Journal of Glaciology, Vol. 13, No. 69, 1974

SUBGLACIAL GEOMORPHOLOGY SURROUNDING THE ICE-FREE VALLEYS OF SOUTHERN VICTORIA LAND, ANTARCTICA

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ABSTRACT. The results of airborne radio-echo (R/E) depth sounding over Wilson Piedmont Glacier, Mackay, Ferrar and Taylor outlet glaciers, and over the ice sheet bordering the mountains, provide ice thicknesses and subglacial topography accurate to 20 m and to 1 km areally. The R/E records show that floors of the Debenham, Wright and Victoria Valleys occur beneath the Wilson Piedmont at elevations of −260 m, and up to 260 and 670 m, respectively. The 670 m "threshold" may have blocked easterly marine and glacial invasions experienced by lower valleys. Profiles along the outlet glaciers display large depressions, some below sea-level. These are associated with erosion by tributaries and with glacial erosion through thick dolerite sills. Elevated ridges thought to be sills submerged beneath the heads of these glaciers also limit nourishment from the adjacent part of the ice sheet. The subglacial west flank of the mountains is formed by a series of high steep-sided plateaux with gentle west-sloping surfaces. Block faulting, west-dipping dolerite and sandstone units, and glacial erosion must explain this topography.

RÉSUMÉ. *Géomorphologie sous-glacaire entourant les vallées déglacées du Sud du Victoria Land dans l'Antarctique.* Les résultats de sondages profonds par radio-écho (R/E) aérien audessus du Wilson Piedmont Glacier, des glaciers effluents Mackay, Ferrar et Taylor et au-dessus des calottes englacées bordant les montagnes, donnent les épaisseurs et la topographie sous-glaciaire à une précision de 20 m et jusqu'à 1 km en surface. Les enregistrements R/E montrent que le lit des vallées de Debenham, Wright et Victoria se trouve sous le Wilson Piedmont, à des altitudes de −260 m et, respectivement, jusqu'à 260 m et 670 m. Le seuil de 670 m peut avoir bloqué de bonne heure les invasions marines et glaciaires subies par les vallées inférieures. Les profils le long des glaciers effluents révèlent de larges dépressions, quelques unes en-dessous du niveau de la mer. Elles sont liées à l'érosion hydrique par la présence d'affluents et à l'érosion glaciaire par d'épais seuils de dolerite. Des reliefs alignés que l'on pense être des seuils ennoyés sous le front de ces glaciers, limitent aussi leur alimentation latérale à partir de la partie adjacente de la calotte. Le versant sous-glaciaire ouest des montagnes est formé d'une série de plateaux à flancs abrupts avec des surfaces en pente douce regardant l'ouest. Les chutes de blocs, entaillant vers l'ouest les formations de dolérite et de grès, ainsi que l'érosion glaciaire, doivent expliquer cette topographie.

ZUSAMMENFASSUNG. *Subglaziale Geomorphologie in der Umgebung der eisfreien Täler des südlichen Victoria Land, Antarktis.* Die Ergebnisse von Radar-Tiefenlotungen aus dem Flugzeug über dem Wilson Piedmont Glacier, den Mackay-, Ferrar- und Taylor-Abflussgletschern und über dem Eisschild, der sich an das Gebirge anschliesst, liefern die Eisdicke und die subglaziale Topographie mit einer Genauigkeit von 20 m der Höhe und 1 km der Lage nach. Die Aufzeichnungen zeigen, dass die Sohlen des Debenham-, Wright- und Victoriatals unterhalb der Wilson-Fussfläche in Höhen von jeweils −260 m, 260 m und 670 m liegen. Die 670-m-Schwelle dürfte das Eindringen des Meeres und der Gletscher von Osten her verhindert haben, wie es bei tieferen Tälern auftrat. Profile entlang der Abflussgletscher zeigen grosse Einsenkungen, einige unter das Meeresniveau. Diese sind mit der Erosion durch Seitengletscher und mit der glazialen Erosion durch dicke Doleritschwellen verknüpft. Rückenartige Erhebungen, von denen anzunehmen ist, dass es sich um Schwellen handelt, die von den Anfängen dieser Gletscher überflossen wurden, beschränken ihrerseits den Zufluss vom angrenzenden Teil des Eisschildes. Die subglaziale Westflanke des Gebirges wird duch eine Reihe von hohen, steil abfallenden Plateaus mit sanften, nach Westen geneigten Oberflächen gebildet. Diese Topographie muss mit Blockfaltung, Kippung von Dolerit- und Sandsteinschichten nach Westen sowie durch glaziale Erosion erklärt werden.

INTRODUCTION

This paper presents the results and geomorphic interpretation of airborne radio-echo depth sounding of the glaciers bordering the ice-free valleys near McMurdo Sound, southern Victoria Land. The glaciers covered include Wilson Piedmont Glacier and its distributaries into the Victoria and Wright Valleys, Mackay, Ferrar and upper Taylor outlet glaciers, and finally that part of the ice sheet of east Antarctica (Victoria Land ice plateau) bordering the valleys on the west (Fig. 1). The radio-echo (R/E) sounding records considered here were collected by personnel of the Scott Polar Research Institute (SPRI)/U.S. National Science Foundation flights of 1967–68 and 1969–70 (Robin and others, 1970[a], [b]). The primary

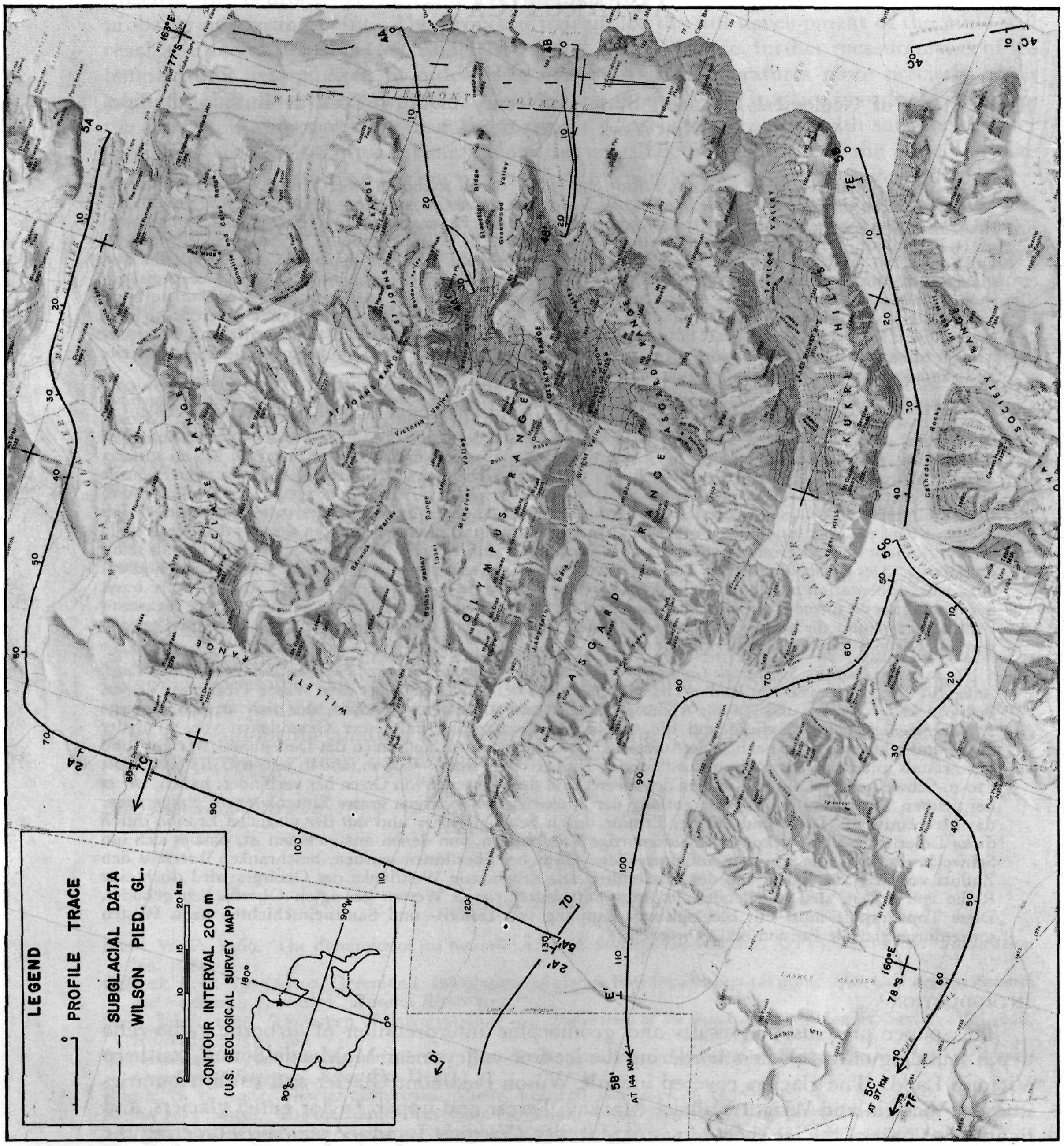

Fig. 1. Topographic map of the ice-free valleys and surrounding glaciers of southern Victoria Land with superimposed profile tracks of Figures 2, 4A–C, 5 and 7C–F, and location of data for Figure 3.

interest of this paper is interpretation of the subglacial information in the context of some glacial geomorphic and structural studies undertaken in the ice-free valley area in recent years. A few preliminary records from this area have been published previously (Calkin, 1971[b], 1973, 1974).

The ice-free valleys, including Taylor and Wright Valleys, and the Victoria Valley system, which themselves have been the subject of many studies (see Bull, 1972), have been cut into the eastern edge of the Transantarctic Mountains to the sea. Together they comprise an area of about 4 000 km² which is the largest ice-free area in Antarctica. Short tongues of the former outlet glaciers, which cut the valleys to their present form, still extend into their western ends from the ice sheet. However, the major valley-cutting glaciers, which were wet-based, retreated from these valleys more than 4 000 000 years ago following emergence of bedrock sills at their western ends (Calkin and others, 1970; Denton and others, 1970, 1971; Calkin, 1971[a], Nichols, 1971; Bull and Calkin, [1972]; Bull and Webb, 1972). Present glaciers are cold-based and probably erode very slowly (Calkin, 1974).

The outer glacier valleys, glacierized or ice-free, are underlain by Precambrian–Ordovican crystalline basement rocks which are overlain unconformably by quartz-sandstones of the late Palaeozoic Beacon Supergroup. Both basement and Beacon Supergroup rocks are intruded by sills of the Jurassic Ferrar dolerite. The sills, each up to 300 m thick, are resistant and form extensive sections of valley walls, peaks, the cap rock of table-land surfaces, and the thresholds and uplands at the western edge of the valleys (Gunn and Warren, 1962; Warren, 1969). The Beacon Supergroup sandstones and Ferrar dolerite sills dip up to 5° inland (westward) in southern Victoria Land.

The surficial characteristics of the glaciers considered here have been described by members of Scott's and Shackleton's Antarctic expeditions of 1901 to 1913 and/or by Gunn and Warren (1962). Bull (1960) computed a bottom profile across Wilson Piedmont Glacier between Wright Valley and Marble Point on 26 gravity stations. Smithson (1972) extended gravity measurements to adjacent ice-free valleys, including scattered stations on Wilson Piedmont and Mackay Glacier. Waite (1962), in the earliest airborne radio-echo sounding tests, made widely spaced sounding from a helicopter over Ferrar Glacier. In addition, three oversnow geophysical/glaciological traverses were undertaken by American parties in this part of the ice sheet during the 1958–59, 1959–60 and 1961–62 seasons. These extended through the mountains and westward across the ice sheet from Skelton Glacier and produced results which were applied to early accounts of the glacial geomorphic and structural history of the study area (Crary, 1963; Robinson, unpublished).

The radio-echo sounding data allow revision of the detailed geophysical conclusions. Although the radio-echo method gives little or no information about the rock composition of sub-ice bottom reflectors, it is equivalent in accuracy and resolution to seismic reflection (Robin and others, 1970[a]). Furthermore, the continuous profiles obtained by the radio-echo profiling methods contain much more geomorphic information and are certainly more reliable in geomorphic interpretation than a linked series of seismic or gravity depths (Evans and Smith, 1969).

Equipment, methods and accuracy

A SPRI Mark II R/E apparatus mounted in a U.S. Navy Super Constellation aircraft was used in 1967 and a SPRI Mark IV in a C13F Hercules aircraft in the 1969–70 season. Both models utilized a pulse-modulated radio system using a carrier frequency of 35 MHz. Ranges are determined from the delay of the returning radio echo from the ice surface and sub-ice bed surface, respectively, and ice thicknesses by subtraction assuming a signal propagation velocity in ice of 169 m μs^{-1}. Evans and Smith (1969) have given a detailed discussion of the equipment and Robin and others (1969) an account of its performance on polar ice sheets.

Records obtained in 1967 over the ice sheet displayed the ice bottom only and ice thicknesses had to be computed indirectly using the aircraft's terrain-clearance radar. Thicknesses derived from these may show unsystematic errors up to 60 m with an average error of 20 m. However, the radar pulse was such that, where both bottom and surface reflections were recorded, ice thicknesses could be measured to an accuracy approaching 10 m. The systematic range error due to uncertainties in knowledge of the velocity of propagation through ice is 10 m or 1.5% of ice thickness whichever is greater. Errors, largely random, within this magnitude may also occur in timing the echo delay. In addition, to ice thicknesses calculated using 169 m μs^{-1} wave velocity in ice, one must add 8–10 m in areas of snow accumulation (Robin and others, 1969).

Gaps which occur in some records, particularly those over the ice sheet, may result from high flight altitudes over unusually deep bottom and consequent lengthening of the echo returns from oblique angles. The occurrence of very steep bottom slopes or changes in the properties of the ice may also be major reasons for the missing record. Echoes are not recorded from a smooth surface with gradients greater than the critical angle in ice, i.e. about 34°.

Film profile records such as that shown in Figure 2, were enlarged and digitized on an *x–y* reader with an accuracy and precision within 5 m. Film traces of echoes were read at 6 s intervals, corresponding to an overland distance of approximately 600–800 m.

Problems of interpretation of the R/E film record similar to those of marine echo sounding or seismic surveys arise due to the broad fan-like beam of the R/E antennae. Resulting errors in interpretation of thickness may be in excess of the systematic range error at points of sudden change in topography (Harrison, 1970). Hyperbolic record traces occur where either the ice surface or the ice–bottom interface is sloping. Furthermore, although the beam is narrowest normal to the flight line, side echoes may occasionally be confused with those from the bottom in runs along narrow valleys or past nunataks. However, the hyperbolic traces were accounted for by eye during digitization from the film record and obvious side echoes were removed from the profiles. Harrison (1970) has developed a computer program to undertake this profile reconstruction more accurately by geometric transformations; however, it was not adapted for this study.

Navigational error for the flights was normally of more importance than the total 10–20 m error considered above. In and adjacent to the mountains, positioning was on U.S. Geological Survey 1 : 125 000 Reconnaissance Series topographic maps (Fig. 1) aided by resection from trimetrogon photographs. Here, navigational inaccuracies were within 1 km. However, on the ice sheet, beyond good map control, positions were determined largely by a flight recorder and the error may be up to about 5 km (see Drewry, 1972[a], [b]).

The radio echo-sounding results and interpretation

Wilson Piedmont Glacier and distributaries

Wilson Piedmont Glacier, stretching for 54 km between New Harbour and Granite Harbour (Fig. 1), is the largest of a series of ice piedmonts which border the west side of the Ross Sea in southern Victoria Land. Data obtained from three north–south R/E flights along this glacier and two flights, each which traverse it along Victoria and Wright Lower distributary glaciers, are summarized in Figures 3 and 4A and B; flight lines are located on Figure 1. Figure 4C and D of Bowers Piedmont Glacier to the south and Oates Piedmont Glacier to the north (see also Figures 1 and 6) are the results of single flights.

Glacier thickness and bed topography. Wilson Piedmont Glacier averages between about 250 and 350 m in thickness, with a maximum of 500 m east of the divide below Victoria Lower Glacier. This is thicker than Bowers Piedmont Glacier but similar to thicknesses suggested by the short, Oates Piedmont Glacier profile.

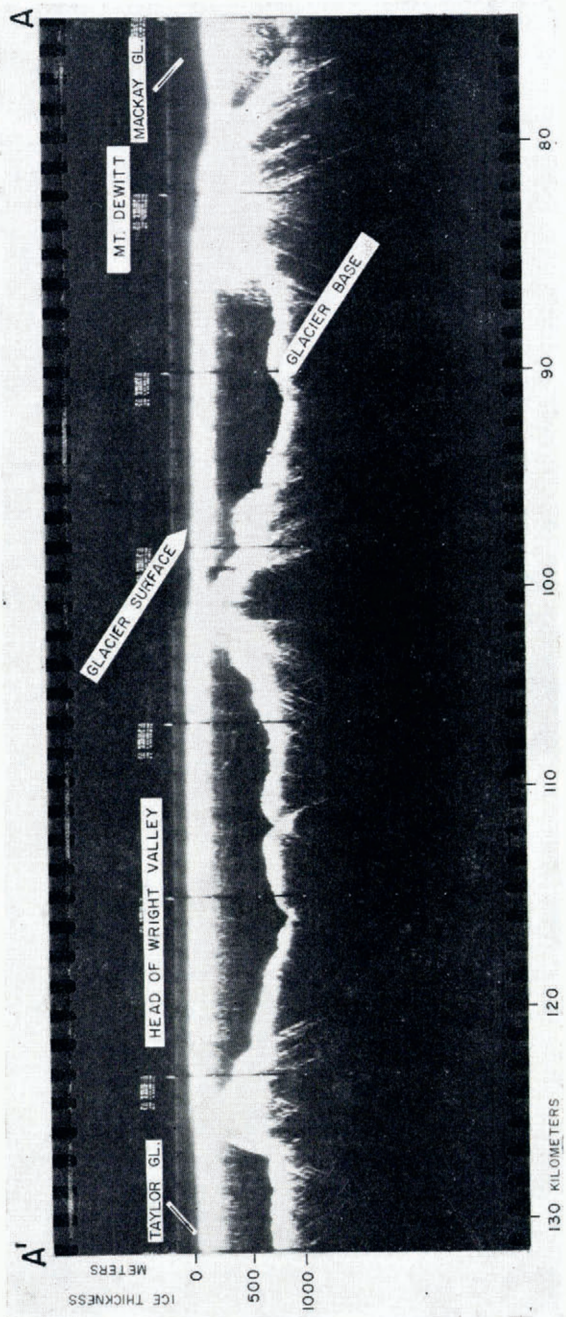

Fig. 2. Photographic enlargement of radio-echo film record from between the heads of Mackay and Taylor Glaciers. The location is shown on Figure 1. Horizontal time-calibration marks (across top) are at 1 min intervals and underlying vertical (depth) calibration marks at 5 μs intervals. Distortion of bottom relief due to differing scales and errors in altimetry are shown corrected as the left part of Figure 5A.

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Subglacial elevation contours (Fig. 3) show that Wilson Piedmont overlies an irregular, seaward-sloping surface between 200 and −200 m elevation which is partitioned by the coastal ends of three east–west glacial troughs (Fig. 1). The trough of Debenham Glacier (once a major distributary of Mackay Glacier) is the most marked and extends more than 260 m below sea-level. In contrast, Wright Valley reaches up from below sea-level on the coast to about 270 m above sea-level (Fig. 4B) under the snout of Wright Lower Glacier. Because the snout rests on an unknown thickness of drift, the bedrock threshold must lie below this point but probably above the 130 m divide at 11 km on Figure 4B. Like Debenham Glacier, the sub-ice profile of the Wright Lower Glacier ice stream is more compatible with repeated outlet glaciation than that of lower Victoria Valley. The latter displays a distinct threshold reaching to about 670 m elevation (Fig. 4A; 34 km) that may be a remnant of marked alpine glacier erosion preceding outlet glaciation (see Taylor, 1922, p. 76).

Sub-piedmont bedrock thresholds and late Cenozoic history. Radio-echo sounding definition of subglacial thresholds of lower Wright and Victoria valleys helps to reconcile some apparent major differences in the tectonic and glacial histories of the ice-free valleys and it complements recent detailed studies in the area as follows. Webb (1972) and McSaveney and McSaveney (1972) provided convincing evidence, in the form of fossiliferous marine deposits,

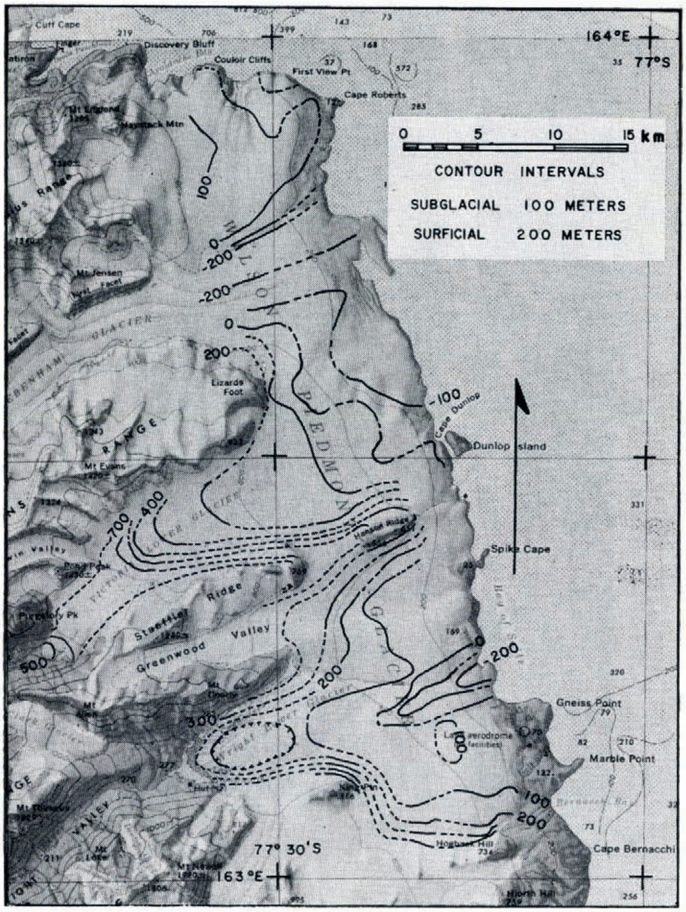

Fig. 3. Subglacial contour map of Wilson Piedmont Glacier superimposed on U.S. Geological Survey topographic map. Location of data used, which includes profiles 4A and B, is shown in Figure 1.

that Wright Valley was a fjord in Tertiary time and hence submerged by an amount at least equal to or greater than its threshold level at the east end. Preliminary work by the McSaveneys and by R. E. Behling and myself suggests that marine deposits may prove to be widespread in the western part of Wright Valley up to 200–300 m elevation with a few possible occurrences to about 600 m (see Bull and Webb, 1972; Calkin, 1974). Thus the bedrock threshold (130–270 m) under Wilson Piedmont Glacier was probably not at a critical elevation for marine invasion as assumed in the past (see Bull, 1972; Bull and Webb, 1972). However, it marked the point of subsequent valley emergence. Furthermore, in the adjacent Victoria Valley system, lack of obvious evidence for marine deposits (Calkin, 1971[a], 1974) may suggest that the marine submergence in this area was limited to below about 670 m, which is the elevation of the eastern threshold shown in Figure 4A. Alternatively, submergence of the Victoria Valley system was blocked by glacial ice (McSaveney and McSaveney, 1972).

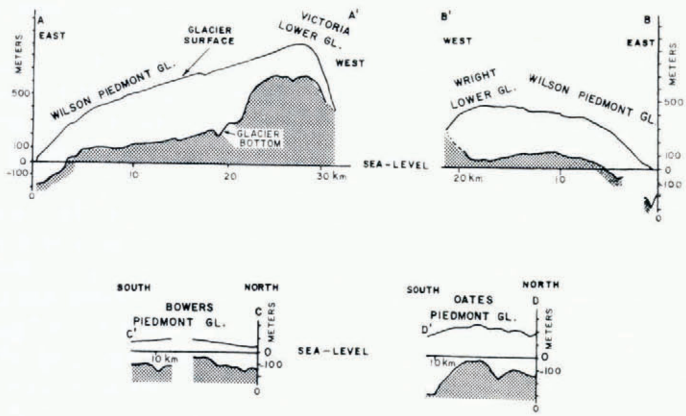

Fig. 4. Radio-echo profiles extending east–west across the Wilson Piedmont (A and B) from the sea to the ice-free valleys and north–south along the Bowers (C) and Oates (D) piedmont glaciers. Profiles A–C are located in Figure 1, D in Figure 6. Dashed lines (B) represent data from adjacent R/E run; dotted lines (A and B) are inferred from topographic map.

Differences in number, magnitude and apparent source of glacial advances which have occurred at the east ends of the ice-free valleys have been a puzzling result of glacial geomorphologic mapping in these valleys. Correlation of these advances is important since they have been ascribed to grounding of the Ross Ice Shelf and in turn provisionally to lowered eustatic sea-levels and Northern Hemisphere glaciations (Calkin, 1971[a]; Denton and others, 1971). Although problems of inter-valley correlation still exist, available information on thresholds now suggests that the following numbers of advances are reasonable. Four westward axial invasions from the Ross Sea occurred into Taylor Valley (Denton and others, 1971), where there is no eastern threshold, while only three advances occurred over the bedrock threshold under Wright Lower Glacier from the east (Behling, 1972; Calkin, 1974). The high threshold at 670 m under Victoria Lower Glacier (Fig. 4A) may have entirely prevented westward flow of ice from the Ross Sea area into this valley system. Although there were west-moving glacial advances at the east end of Victoria Valley that have been provisionally ascribed to the grounded Ross Ice Shelf (Calkin, 1971[a]), no evidence of direct ice movement from the Ross Sea has been found. Therefore, their source may have been local alpine glaciers.

Mackay, Ferrar and Taylor outlet glaciers

Figure 5A, B and C are R/E profiles along Mackay Glacier, lower Ferrar–upper Taylor Glacier system, and upper Ferrar Glacier, respectively (Fig. 1). Substantial surface clutter echoes from intense crevassing have prevented a complete bedrock profile of Mackay Glacier (Fig. 5A) from being obtained. However, runs up Ferrar and upper Taylor Glaciers are much more complete and probably more accurate. Figure 5B and C represent composites of two R/E runs each. General descriptions of these outlet glaciers have been given by Taylor (1922), Wright and Priestley (1922), Hamilton and Hays (1961), and Gunn and Warren (1962).

Glacier thickness and bed topography. The average thicknesses of Taylor, Ferrar and Mackay Glaciers shown by the R/E profiles are similar, varying respectively from about 450 to 550 m through much of their lengths. However, the thickness at the grounding line (sudden glacier thinning and leveling of the surface) of Mackay Glacier (Fig. 5A) is much greater than the other two glaciers, partly reflecting its greater activity. All three glaciers are substantially smaller than when they cut their valleys during an earlier, full-bodied stage of the ice sheet.

The profiles show innumerable bottom irregularities, i.e. steps and adjacent depressions, similar to those throughout the ice-free valleys. Some of the smaller depressions or irregularities may not be either closed and/or so large; they may be due to the varying path of the bottom-echo return relative to the path of the deepest part of the valley cross-section or thalweg. Other explanations are discussed below.

Origin of subglacial steps. Taylor (1922) believed that many of the irregularities on the floors and walls of Taylor Valley and Ferrar Valley represented parts of headwardly eroded cirques or short valleys overwhelmed by the outlet glaciers. However, many exposed steps, thresholds and depressions have been related to outcrops of the dolerite sills by field mapping. The sill-controlled steps most frequently occur in forms referred to as "pseudocirques" by Gunn and Warren (1962). They believed these to have been formed by selective headward retreat of dolerite scarps once breached by concentrated vertical erosion from the outlet glaciers. This origin appears applicable in part to the Mackay, Ferrar and Taylor Valley profiles. Steps and lows are particularly pronounced under Taylor and Ferrar Glaciers just down-valley from where two dolerite sills converge to form very thick units, i.e. below Finger Mountain at 69 km on Figure 5B and also at the west edge of the Kukri Hills at 58 and 43 km. Another pronounced step in the valleys coincides with the eastern edge of the dolerite sill exposure at 29 km.

The steps and lows also correspond in position with incoming tributary glaciers as well as being controlled partly by the dolerite sills as shown by Figures 1 and 5B. The pronounced basin east of the Ferrar–Taylor Glaciers divide (Fig. 5B; 43 km) is analogous in position with the deep basins suggested by the Mackay Glacier records (Fig. 5A) as well as with the basins of Lakes Vida and Vanda immediately down-valley from major tributary junctions in Victoria and Wright Valleys, respectively (Fig. 1).

Mountain crest thresholds and glacier regimen. The basic control on intrusion of the plateau ice to the mountain valleys, including the Victoria, Wright and lower Taylor ice-free areas, must be the delicate balance maintained between the elevation of the adjacent ice sheet and that of the resistant dolerite sills or thresholds which field studies suggest underlie their heads or form the mountain crestal cols (Gunn and Warren, 1962). This balance in turn has been controlled in the past by several factors including climate and by tectonic eustatic and ice-sheet movements as well (Bull and Webb, 1972). Geologic studies have shown the effect of these sills at the heads of the ice-free valleys (Calkin, 1971[a]), where parts of the sills are exposed, and the R/E records confirm the limiting effects on these and on the surrounding glacierized valleys (Figs. 2 and 5A).

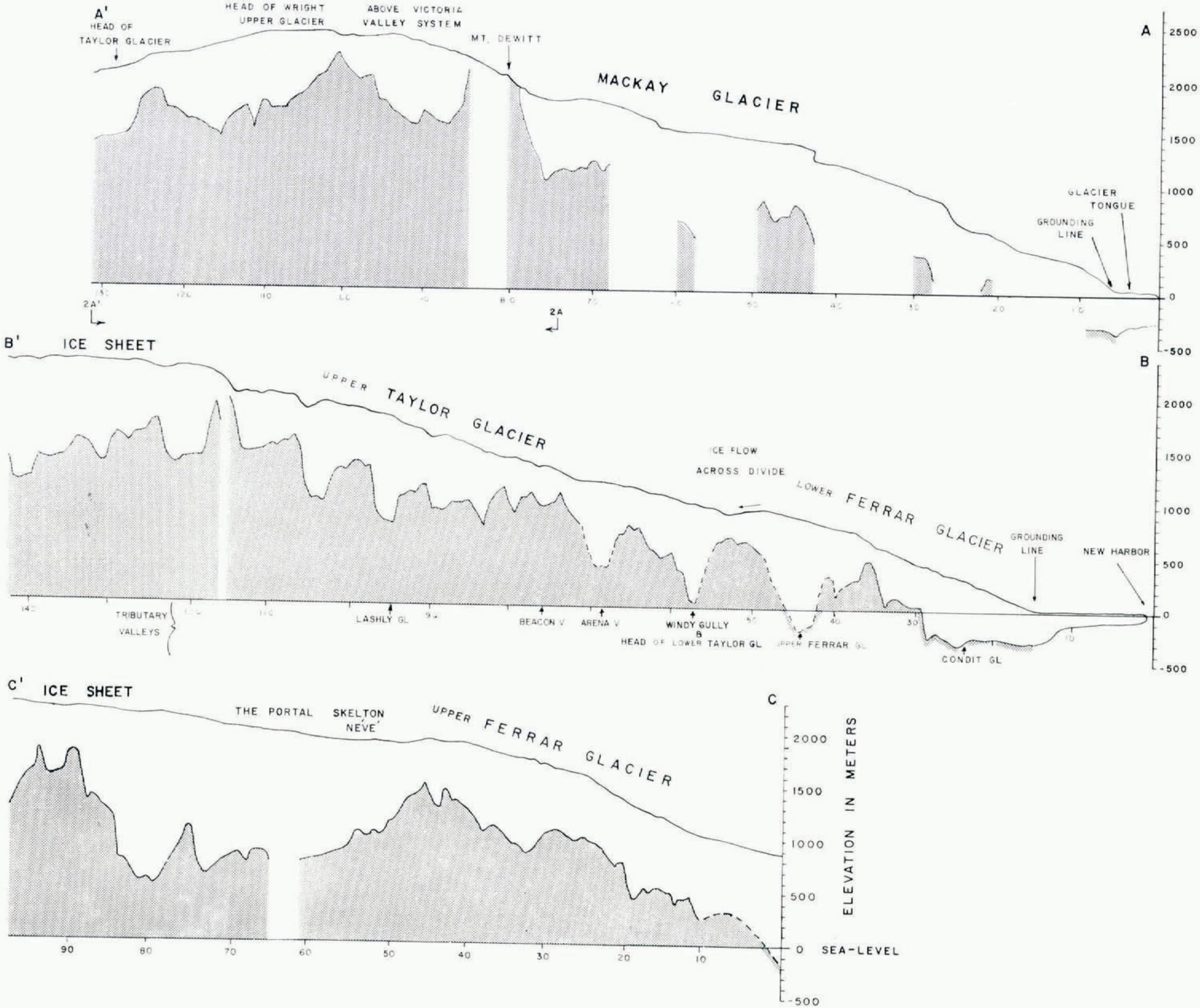

Fig. 5. Radio-echo profiles along: A, Mackay Glacier and ice-sheet margin above the valleys; B, lower Ferrar–upper Taylor Glaciers; and C, upper Ferrar Glacier. Profiles are located in Figure 1. Blanks in base show missing data, dashed lines represent data from adjacent R/E run. Positions of tributary glaciers are indicated on profile B.

At the head of Taylor Glacier (Fig. 5B; 115 km) the subglacial surface rises to about 1 900 m and the glacier thins to 100 m. Ferrar Glacier, on the other hand, heads in the Skelton Névé (Fig. 5C; 45 km) over a col where the ice stream is 400 m thick. Above The Portal (Fig. 1) on the ice sheet near the same latitude as the Taylor Glacier threshold, the sub-surface divide reaches only to 1 800 m (Fig. 5C; 90 km) and ice coming down from the ice sheet is again 400 m thick. Thus Ferrar Glacier reaches the sea, albeit just barely, and Taylor Glacier does not. This is despite the fact that Taylor Glacier is fed by about one-third of the upper Ferrar Glacier discharge via Windy Gully and across the divide of Figure 5B (at 53 km), and also has a wide and direct opening to the ice sheet.

Although the traverse over Mackay Glacier did not extend directly west on to the ice sheet, adjacent data (Fig. 6) indicate that the threshold across the mountain crest is broad and lower than that across Ferrar or Taylor Glaciers, and hence Mackay Glacier is more active. Summer movement of up to 0.85 m d^{-1} has been measured on the Mackay Glacier

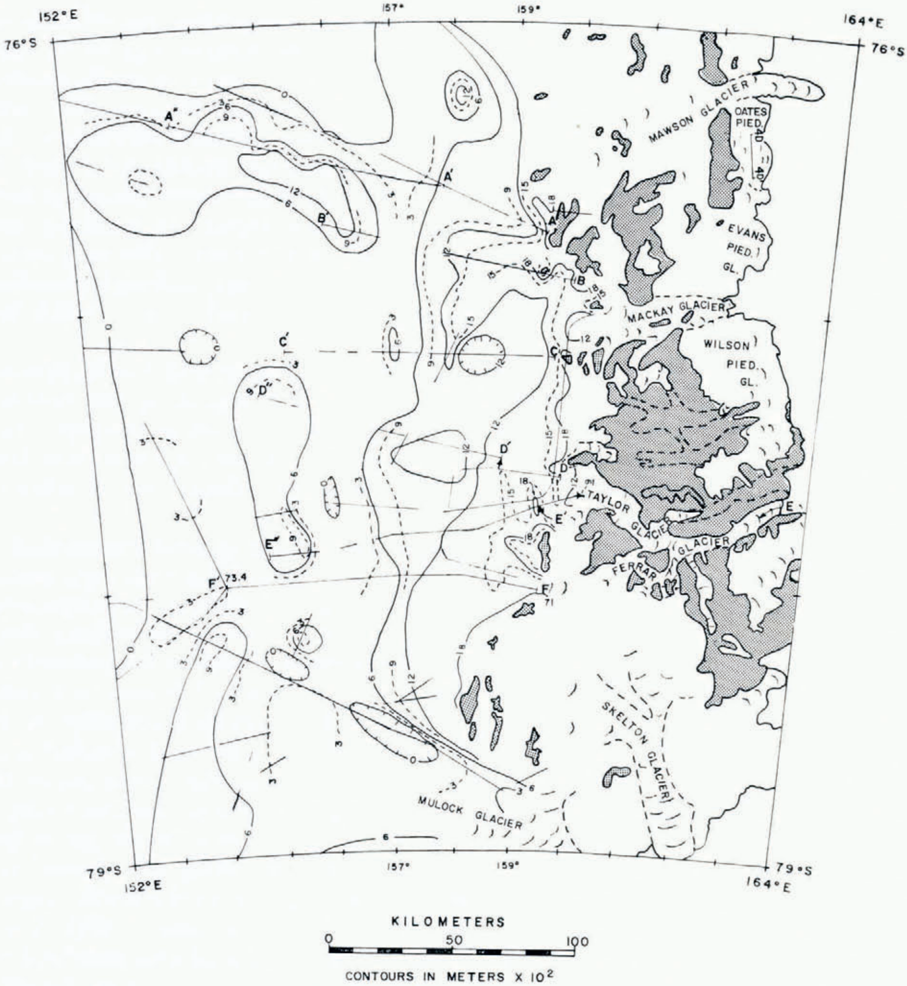

Fig. 6. Subglacial contour map of the west flank of the Transantarctic Mountains between Mawson and Mulock Glaciers. Straight lines show location of data and profiles of Figure 7. Profile line F and two adjoining lines to the west locate oversnow traverse data.

https://doi.org/10.3189/S0022143000023194 Published online by Cambridge University Press

ice tongue (Taylor, 1922) compared to approximate velocities of 0.04 and 0.08 m d^{-1} on steeper sections of lower Ferrar and lower Taylor Glaciers, respectively (Wright and Priestley, 1922; Hamilton and Hays, 1961). Radio-echo sounding records indicate that thresholds are not present at the inland ends of Mulock Glacier (Fig. 6) or Byrd Glacier 100 km farther south; therefore, these glaciers fill their valleys and provide some of the largest discharges to the Ross Ice Shelf that come from the ice sheet of east Antarctica.

The relatively thin connections between ice sheet and valley glaciers in the ice-free valley area, particularly to Taylor Glacier, indicate that local precipitation may play an important part in any current ice-tongue movement. Studies by Wilson and Crary (1961), for example, suggest that Skelton Glacier with a movement of 0.24 m d^{-1} at its snout receives negligible ice supply from the plateau because of high sill thresholds (such as below The Portal; Figs. 1 and 5C) and it is nourished almost entirely by direct precipitation.

Victoria Land ice plateau margin

The records. The subglacial as well as ice-sheet surface topography immediately west of the ice-free valleys on the west flank of the Transantarctic Mountains is given by Figures 6 and 7. These figures were compiled from data of 11 east–west and two north–south radio-echo flights along with information from the oversnow geophysical traverses. The oversnow data are located along profile line 71–73.4 (F–F′) and the two adjoining segments to the west in Figure 6; it was assembled from Weihaupt (1961), Robinson (1962) and Crary (1963). Limitations of detail in the contour map of Figure 6 are due to previously described problems in R/E accuracy, navigation and the broad spacing of the flights. In addition, contours depend on absolute elevation (a.s.l.) determined by aircraft altimetry. This may be in error locally by up to 200 m in areas beyond topographic map coverage, although elevations are tied inland to oversnow-traverse elevations as corrected by Crary (1963) which may be accurate to ± 50 m.

The profiles shown in Figure 7 give more accurate ice thickness than can be interpolated from the contour map in addition to their much greater detail. Data along east–west traverse lines which include profiles B, C and D of Figure 7 may be up to 100 m more in error than those along the other traverses. Malfunction of the flight recorder on these three traverses required that the aircraft terrain clearance used to compute ice thicknesses be taken from a few widely spaced radar height readings.

Subglacial topography. The R/E results show that some of the general assumptions of the cross-sectional shape of the Transantarctic Mountains held since the first geologic exploration (Ferrar, 1907; David and Priestley, 1914) are correct; that is, what may be characterized as the main mountain mass is about 170 km wide and with a decrease in bedrock elevations westward somewhat similar to that on the exposed east flank. Without more closely spaced traverses, it will never be certain just how similar the two sides are; however, it is clear that most of the subglacial western flank recorded by the R/E system does not show in detail a single pronounced west-facing scarp or even a simple descending series of west-facing escarpments as has often been assumed since that time. An exception may occur above Skelton Glacier (Fig. 1), where the oversnow-traverse profile (Fig. 7F; 50 km; see also Crary, 1963, fig. 25a) does indicate a rather simple and steep westward descent of about 700 m between long. 157° and 158° E. Although the changes in topography along this ground traverse correspond very generally with those of the R/E profiles adjacent to the north (Figs. 6 and 7E), effective seismic depth control on the traverse was not obtained (Crary, 1963). Therefore, the true west slope here may be much more irregular than indicated and structural conclusions partly based on this profile such as that by Calkin and Nichols (1972) need re-evaluation.

The R/E profiles are characterized by a series of isolated subglacial mountains, particularly plateau-like or dissected tablelands with gently west-sloping tops, flanked by very

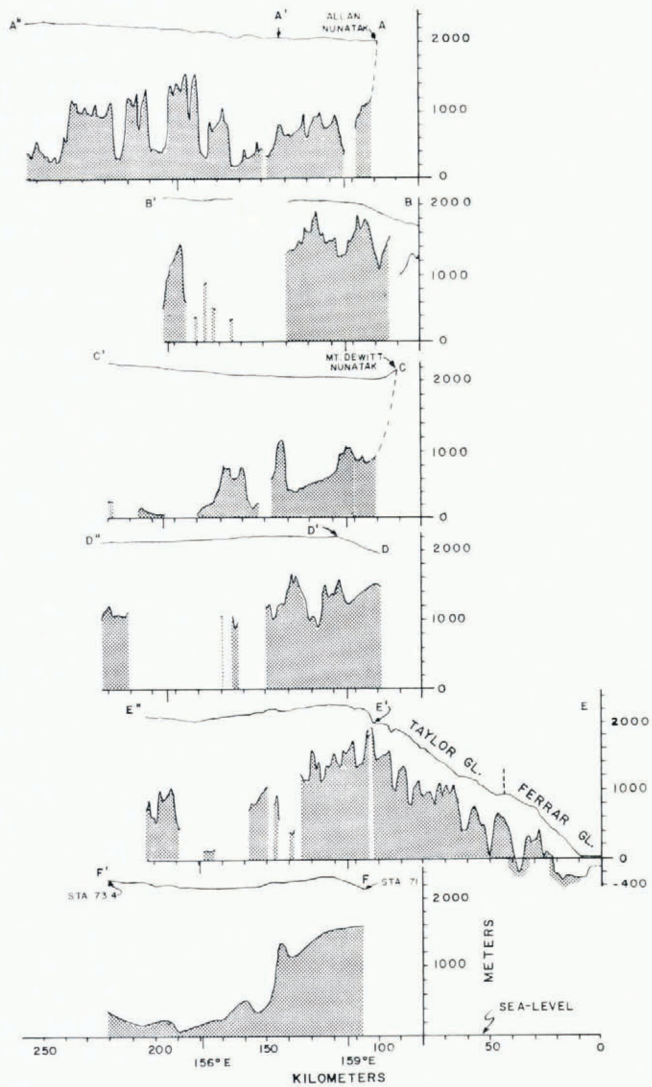

Fig. 7. Radio-echo profiles 7A–E and oversnow traverse profile (F) across the west flank of the Transantarctic Mountains in the ice-free valley area. Profile F was assembled for geophysical data (largely gravity and magnetic) of Weihaupt (1961), Robinson (1962) and Crary (1963). Profiles are aligned in their respective positions along long. 159° E. Profiles A–F are located in Figure 6, and C–F are also located in Figure 1.

steep slopes on at least east and west sides. These show relief of 400–900 m and are separated similarly by gently west-sloping valley or broader lowland surfaces 1–20 km wide. Most profiles are too far apart (about 30 km) to establish whether there are also north- or south-facing escarpments. The average inclination of the major slopes, which appear near vertical on the scale of Figure 7, probably vary between 20° and 55° as interpolated between R/E readings on near-horizontal parts of the top and bottom. The common gentle west-sloping surfaces vary in steepness, but profiles A, C and E show many broad lows with slopes of about 0.01 or somewhat less than 1°.

The crestal upland area is relatively narrow, being about 50 km wide between the 1 000 m contours. This is less than one-quarter of the width of the upland in the Beardmore Glacier area (Drewry, 1972[b]). Bed elevations uniformly reach down westward along the study area to at least 300 m above sea-level 60–70 km west of the crest, giving the mountains a gross symmetrical appearance; however, in detail, they are slightly skewed westward. Several plateaux project more than 900 m above the basins beyond the 300 m elevation contour and the broad open areas below sea-level of the Wilkes Subglacial Basin occur more than 100–150 km from the mountain divide (Fig. 6). The most pronounced and interesting of the subglacial plateau(x) occur in this area (Fig. 7A) strung out east–west between long. 154° and 157° E. Traverses here (Fig. 6) are close enough to show that these are probably terminated by very steep slopes on the north side as well as to the east and west. With this major exception, the grain of the topography is north–south.

Origin of subglacial topography. Structural effect. Faulting and differentially tilted fault blocks, accentuated by sub-horizontal surfaces of the thick Beacon Supergroup sandstones and the Ferrar dolerite sill units, have been the major structural elements defined by field studies throughout the Transantarctic Mountains (Grindley and Laird, 1969; McGregor and Wade, 1969; Warren, 1969; Robinson, unpublished). These faults consist of a complex of longitudinal and transverse faults related to the Victoria orogeny of Tertiary and Quaternary time (Gunn and Warren, 1962; Webb, 1972; Maigkov, 1973). The early concept of David and Priestley (1914) that the mountains are bounded by two major normal fault systems, or that they are part of a "Great Antarctic horst" (Taylor, 1922; Gould, 1935) has not been supported by most previous geophysical work (McGinnis and Montgomery, 1972; Smithson, 1972; Robinson, unpublished) nor by the R/E data given here. Alternative anticlinal or monoclinal tectonic hypotheses for this area (Hamilton, 1960, 1965; King, 1966) have similarly had little definitive support.

Considering the pervasive occurrence of block faulting, it is not surprising that this pattern appears to be repeated on the submerged flank of the mountains. Analysis by Drewry (1972[b]) of R/E data from the Beardmore Glacier area to the south has confirmed a multi-scarped decent of the inland mountain flank there which is similar to that described above for the present study area. Drewry has attributed this topography to westward-tilted fault blocks affected by varying amounts of erosion.

Many of the steep slopes suggested by the subglacial profiles and the contour map, particularly those aligned north–south or near parallel to the coast at about long 157° E. and farther westward, appear to be most easily explained by large faults. In addition, the isolation of the high mountains or plateaux, such as that series revealed by Figure 7A, may be related to transverse or east–west fault lines. For example, a fault delineated along the northern margin of the Mackay Glacier valley (Warren, 1969) and/or the low immediately inland, if real, could be an eastern extension of such a system. However, there is no indication of a single major normal fault system bounding the inland side of the Transantarctic Mountains here. The general westward slope of the summits on the western mountain flank, as well as the gentle, near-uniform, westward-tilted plateau-like surfaces (high or low) seen in the profiles (Fig. 7) can be reproduced crudely by projection of the regional westward-dipping Beacon Supergroup rocks and Ferrar dolerite sills which crop out at the mountain crest. Whether these slopes are directly due to differential tilting during periods of block faulting and isostatic movement or are also due in part to a distinct folding episode(s), as suggested by the tectonic hypotheses of Hamilton (1965) or King (1966), is not known.

Glacial effect. Erosional features, formed by alpine glaciers before build-up of the ice sheet of east Antarctica must also account for some of the texture, particularly the broadened longitudinal or transverse valleys suggested by Figure 6 and the subglacial profiles of Figure 7 (Selby and Wilson, 1971; Calkin, 1974). In addition, inland-flowing valley glaciers nourished from successively diminished local glacier-covered plateaux (Gunn and Warren, 1962) may

https://doi.org/10.3189/S0022143000023194 Published online by Cambridge University Press

have contributed to the build-up of the ice sheet in this area (Bull and others, 1962). Drewry (1972[a], [b]) has shown evidence from the R/E data for inland-trending glacial valleys in the Queen Maud Range.

The broad contour interval of the sub-surface data here precludes definition of most glacial features but some large inland-trending valleys of possible glacial origin may be represented in the areas of profiles A and C of Figures 6 and 7 near the heads of Mackay and Mulock Glaciers. The apparent back-to-back alignment of east- and west-sloping valleys here is similar to that in the Queen Maud Mountains (Drewry, 1972[b]) and may be evidence of the local plateau-glacier initiation of opposing glacial valleys. However, the narrowness of the mountains in this area suggests that most glacial valleys should be very small compared to those described by Drewry for the southern Transantarctic Mountains, and hence the examples cited above may be suspect.

The erosional effect of eastward movement of the ice sheet over the mountains during the outlet-glacier-cutting phase is of uncertain importance. Local accumulations of wet-based till believed to have been deposited during an early, more extensive (full-bodied) stage of the ice sheet have been described from nunataks near the heads of Ferrar and Mackay Glaciers (Mayewski, in press); thus some erosion must have occurred on the west mountain flank.

Summary and conclusions

The following are some major topographic and glacial geomorphic conclusions suggested by the radio-echo sounding data:

1. Three outlet glacier valleys occur beneath Wilson Piedmont Glacier; of these, Debenham Glacier is 260 m below sea-level and Wright and Victoria Valleys are respectively up to about 260 m and to 670 m above sea-level. The 670 m "threshold" may have been high enough to prevent late Cenozoic marine and glacial incursions experienced by Wright and Taylor Valleys.
2. Mackay, Ferrar and Taylor outlet glaciers display several large steps and basins similar to those in the ice-free valleys; some reach below sea-level. The basins coincide with incoming tributaries and with areas of thickened dolerite sills. Sill thresholds at the mountain crest also severely limit flow from the ice sheet, particularly to Taylor Glacier.
3. The Transantarctic Mountains here are narrow-crested compared to the mountain range to the south, with elevations skewed slightly westward in transverse section. The subglacial western flank is characterized by an irregular series of plateau-like mountains, many 900 m in relief with steep sides and very gently west-sloping surfaces and intervening lows. The gentle surfaces are probably related to west-dipping Beacon Supergroup sandstones and tabular dolerite sill bodies. The topography is not consistent with the old "Antarctic horst" structural hypothesis but rather with the more recent hypotheses of complex block faulting.
4. Alpine glaciers and westward movement of the incipient ice sheet of east Antarctica must also have contributed to the overall subglacial topography of the west flank of the mountains here. However, the orientation and wide spacing of flight lines does not allow clear identification of glacial features as such. Nevertheless, inland-sloping valleys heading above Mackay and Mulock outlet glaciers may be at least partly of glacial origin.

Acknowledgements

This work was funded by U.S. National Science Foundation Grant GV-22761 and by a State University of New York Institutional Funds Award. G. de Q. Robin and S. Evans provided facilities and encouragement at the Scott Polar Research Institute, University of

https://doi.org/10.3189/S0022143000023194 Published online by Cambridge University Press

Cambridge, where most of this work was undertaken. I am grateful to D. J. Drewry and C. W. M. Swithinbank for many stimulating discussions on related problems and for review of this manuscript. D. Hodge reviewed an early edition of the paper. Substantial help in data reduction was very kindly given by B. M. E. Smith.

MS. received 15 October 1973 and in revised form 1 March 1974

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