—研究論文—
Scientific Papers

Velocity, Heat Budget and Mass Balance at Anvers Island Ice Cap, Antarctic Peninsula

Gino CASASSA*

南極半島アンバース島氷冠における流動速度，熱収支，質量収支

Gino CASASSA*

要旨：南極半島のアンバース島氷冠において，1982年夏季に雪氷学的調査を行った．氷厚は重力値から計算により，表面質量収支は2地点の積雪層位観測から求められた．平衡線高度は海抜 180 m と推定された．キャンプ 1（標高 474 m）における熱収支の計算結果によると，夏季12日間では，氷河表面の受熱量と放熱量とはほぼ等しく，主な受熱源は短波放射収支，放熱源は蒸発潜熱と長波放射収支であった．この観測期間中，少量の融解が起こっていた．深さ 11 m までの掘削コアには，数枚のアイスレンズが観察され，融解水の再凍結が生じていたことを示す．表面質量収支分布をもとに，46.5 km^2 の地域における質量収支を算出した．その結果，6×10^{13} g·a^{-1} の正の質量収支が得られた．この結果は，1965–1967 年の同観測域を含むアンバース島氷冠における見積もり結果とほぼ一致した．

Abstract: General glaciological characteristics of a part of Anvers Island, Antarctic Peninsula, were studied during the summer of 1982. Ice thickness was calculated from gravity data. Net balance was determined by snow stratigraphy at two campsites and the equilibrium line altitude was estimated to be at 180 m above sea level. A heat balance was assessed for Camp 1 (474 m a.s.l.) in two ways by using different equations to estimate long wave and turbulent fluxes. Both calculations show that heat sources are approximately equal to heat sinks for the 12-day summer period. The main heat source is short wave radiation while the main sinks are evaporation and long wave radiation. A small amount of melting occurred during the period. Hand drilling down to 11 m was carried out and several ice lenses were found, an evidence that meltwater refreezes as it percolates down due to sub-zero englacial temperatures. A mass balance was calculated for an area of 46.5 km^2 by iso-balance curves. A positive net balance of 6×10^{13} g·a^{-1} was obtained. This result was comparable to that of 1965–1967 for a part of the ice cap of Anvers Island which included our study area.

1. Introduction

Anvers Island is the biggest and southernmost island of the Palmer Archipelago, west coast of the Antarctic Peninsula (Fig. 1). It has an approximate area of 2700 km^2 and its mean latitude is 64°30′S and longitude 63°30′W. Located within the subpolar belt of low pressures, cyclonic conditions are predominant throughout the year, with

* 北海道大学低温科学研究所. Institute of Low Temperature Science, Hokkaido University, Kita-19, Nishi-8, Kita-ku, Sapporo 060.
Present address: Byrd Polar Research Center, Ohio State University, 125 South Oval Mall, Columbus, OH 43210–1308, U.S.A.

high annual precipitation.

Except for a few rock outcrops along the coast, Anvers Island is entirely covered by ice which calves into the sea. The eastern half of the island has a very rough topography and consists of glacier-covered mountains, the highest being Mount Français (2822 m). The western half consists of an unbroken ice cap which flows from the base of the mountains at 850 m to the sea. This ice cap is thus a piedmont which is known as Marr ice piedmont. Glaciers have subpolar characteristics, being in a transition zone between the temperate glaciers of Patagonia and the cold polar West Antarctic ice sheet.

RUNDLE (1970, 1973) carried out the first systematic glaciological study on Anvers Island, covering a broad area of the island from Palmer Station, located on the southwest coast. He made extensive measurements over a 3-year period including ice

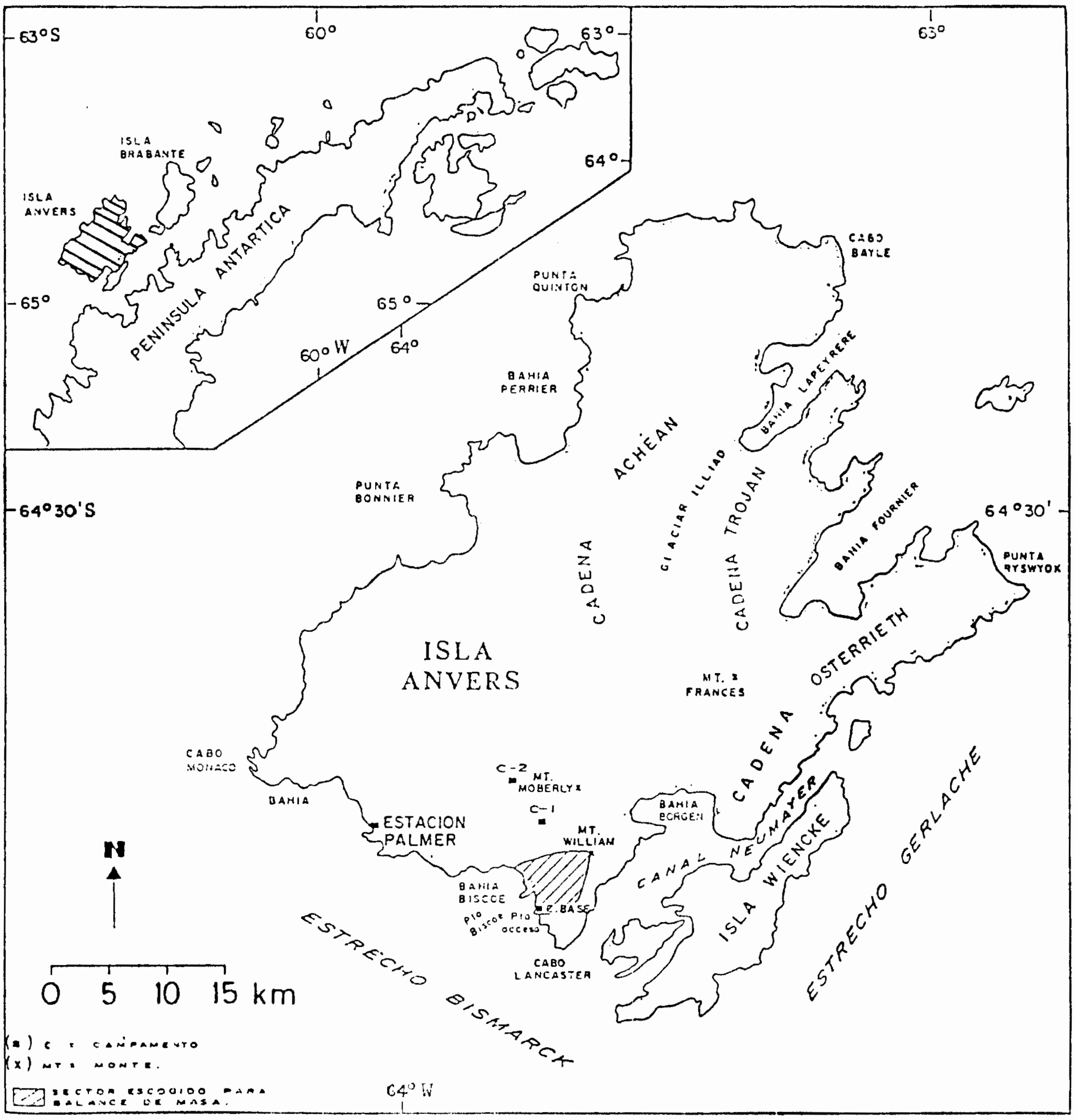

Fig. 1. Location of Anvers Island.

velocity, ablation/accumulation by stake method, stratigraphy of the upper layers of the ice cap and meteorological observations. As part of the same research project, DEWART (1971) made a gravity survey over Anvers Island and determined ice thicknesses. RUNDLE's conclusion was that the mass balance of a representative part of the ice cap is approximately in equilibrium.

The object of the present work was to study the general glaciological characteristics of an area of Anvers Island. The field work was carried out in January–February 1982 as part of the first expedition of a 3-year long joint project of the Chilean Antarctic Institute (INACH) and Hannover University on the determination of the ice cap flow by TRANSIT satellite system.

This work was presented as a thesis to Universidad de Chile (CASASSA, 1984) and in this paper main results are summarized and discussed, including a recalculation of the heat balance.

2. Glacier Survey

Three sets of Marconi TRANSIT equipments for satellite positioning were used by the German investigators on each of the three camps (SCHMIDT, 1983; HINZE, 1983). Base Camp was set up on bare rock on the coast and two camps were set up on the glacier: Camp 1 at 474 m a.s.l. and Camp 2 at 687 m a.s.l. Analyzing a 15-day common record by translocation method with base camp, the coordinates and summer ice velocities at Camp 1 and Camp 2 were determined. Markers were left *in situ* and the station reoccupied in 1983, thus determining the annual velocity. Results are shown in Table 1.

Table 1. Ice velocities deduced from satellite positioning measurements (TRANSIT).

Station	Summer period				Annual period	
	1982		1983		1982–1983	
	Velocity m/a	Azimuth °	Velocity m/a	Azimuth °	Velocity m/a	Azimuth °
Camp 1	102.2	222	84.0	197	96.7	201
Camp 2	62.1*	231*	47.5	158	42.1	215

* These values are not accurate due to problems of the oscillator of the receiver.

Chilean investigators carried out a standard survey between camps and determined the coordinates of the intermediate stations E1 to E12 (Fig. 2).

3. Ice Thickness

Precise gravity measurements were made during several traverses on the ice cap by H. LANGE with a LaCoste & Romberg gravimeter. Accounting for instrument drift and tides, the mean square error of gravity data was less than 0.07 mgal (LANGE, personal communication).

Standard corrections were applied by the author to these data. Units are mgal per meter of elevation (h) above sea level of the station.

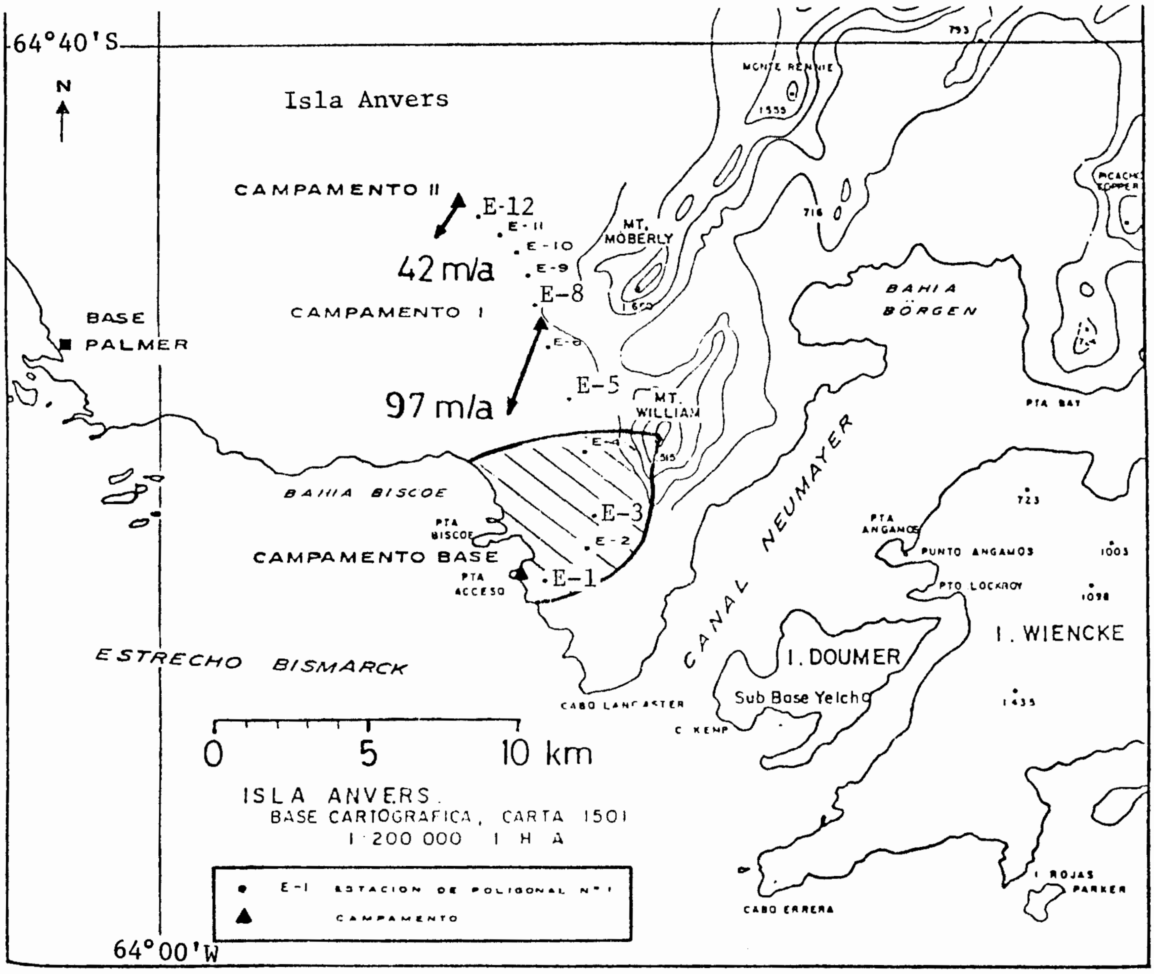

Fig. 2. Location of camps and stations at Anvers Island. Mass balance was computed for the hatched area.

—Free air corrections (FC): $FC = +0.30886$ mgal/m
—Bouguer corrections (BC): $BC = -0.11320$ mgal/m
—Latitude corrections (LC): $LC = +0.6253$ mgal/km
(the distance (s) measured in km northward from Base Camp)

—Terrain corrections (TC): computed from map of the Chilean Navy Hydrographic Institute (IHA) at 1 : 200000 scale where many contour lines had to be estimated; the major contribution is due to Mt. Moberly and Mt. William (see Fig. 2). TC varied between 1.3 and 4.5 mgal.

Bouguer anomalies (BA) were calculated by applying the above corrections to LANGE's data (G):

$$BA = G + (FC + BC)h + LCs + TC.$$

Regional anomalies (RA) were estimated from the gravity difference of 1.6 mgal measured between Base Camp and Palmer Station. Projecting this value to an axis perpendicular to the main range in the Antarctic Peninsula a gradient of -0.1 mgal/km away from the mountains is obtained. This is probably due to local effects which modify the positive trend of the gradient as measured by DEWART (1971). Thus the gravity anomaly due only to density contrast between snow and ice was calculated as:

$$GA = BA + RA.$$

For the interpretation of the results the ice cap was considered to be an infinite slab of ice. Average ice density was estimated as 870 kg/m^3, while a rock density of 2700 kg/m^3 was used, which corresponds to granodiorites with partial hydrothermal alteration, as found in base camp. This way, the thickness of ice per mgal of anomaly is 13.03 m/mgal and ice thicknesses were computed for each station. Bedrock elevations were calculated from the surface elevations determined in the survey.

In Fig. 3 the results are shown. Ice thickness errors are estimated to be less than 17%. A maximum thickness of 572 m was obtained for Camp 2. DEWART (1971) calculated ice thicknesses of 500–550 m for the same area by gravity survey.

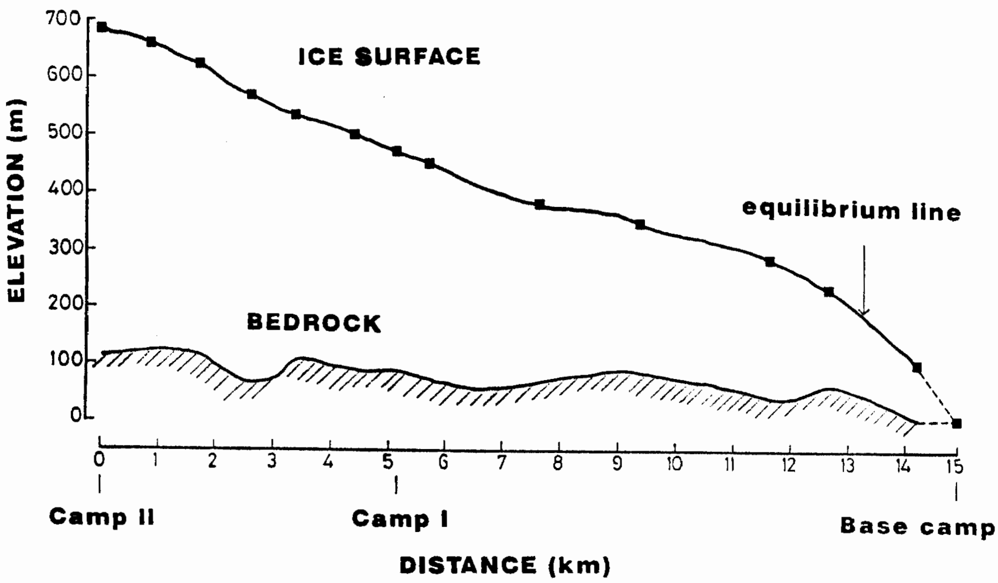

Fig. 3. Longitudinal profile of the ice cap along the line joining the three camps and intermediate stations (see Fig. 2). Ice thickness was obtained by the gravity method.

4. Heat Balance

In order to assess the heat sources and sinks a heat balance was calculated for a 12-day period at Camp 1, every 12 hours. Cyclonic conditions prevailed over most of the period, with a net accumulation depth of 0.22 m of new snow, average of 9 stakes. Only solid precipitation occurred and the heat flux due to snow fall was neglected.

The heat balance equation at the snow surface can be expressed as follows:

$$Q_{SW} + Q_{LW} + Q_S + Q_L + Q_M = X,$$

where all terms are expressed in MJm^{-2}d^{-1}:

Q_{SW}: net short wave radiation,
Q_{LW}: net long wave radiation,
Q_S: sensible heat flux,
Q_L: latent heat flux (evaporation (−) or condensation (+)),
Q_M: heat used for melting snow,
X: remainder term.

The remainder term X includes the heat capacity change within the layer, refreezing of meltwater in the layer at 0°C and the conductive heat flux to deep layers.

Incoming short wave radiation was measured with an actinograph. Rough measurements of albedo with the actinograph gave values of about 0.8. This constant value was used for the calculation, which is reasonable if we consider that the surface consisted mostly of new snow.

Incoming long wave radiation for clear skies (L_0) was estimated from a Brunt-type empirical equation derived by YAMANOUCHI and KAWAGUCHI (1984), for Syowa Station, East Antarctica, where there is a weak inversion layer:

$$L_0 = L_u(0.563 + 0.093 e_1^{0.5}),$$

L_u is the upward longwave flux given by $L_u = \sigma T_s^4$, with σ being the Stefan-Boltzmann constant (5.67×10^{-8} Wm^{-2}K^{-4}) and T_s the surface snow temperature. Absolute vapor pressure e_1 (mb) at screen level (1 m) was calculated from the screen level temperature (T_1) and relative humidity measurements from a hygrothermograph. At Camp 1 snow temperature was measured at five levels and no clear stratification or inversion layer was observed. No direct measurements exist for T_s and it was estimated from the screen-level temperature T_1 in the following way: $T_s = T_1$ for cloudy days and $T_s = T_1 - 3$ (°C) for clear days.

The incoming longwave flux for cloudy skies (L_0^*) was estimated from the following empirical formula (OKE, 1978):

$$L_0^* = L_0(1 + a n^2),$$

where n is the cloud amount in tenths and a depends on the cloud type. Mean cloud amount for the period at Camp 1 was 8/10 and predominant clouds were stratus, so that $a = 0.24$. Finally, the net long wave radiation is the difference between downward and upward long wave fluxes:

$$Q_{LW} = L_0^* - L_u.$$

Because of lack of field measurements of turbulent heat fluxes bulk equations were used to make an estimation. As no calibration for Antarctica was found, values of parameters derived for northern Hokkaido (MOTOYAMA, 1986) were used:

$$Q_S = 0.22 u_1 (T_1 - T_s) \quad \text{Ly/h},$$
$$Q_L = 0.35 u_1 (e_1 - e_s) \quad \text{Ly/h},$$

where u_1 is the wind speed in m/s at screen level and e_1 and e_s are the vapor pressure in mb at screen level and the snow surface, respectively.

On the stakes it was observed that the snow surface level decreased a few times

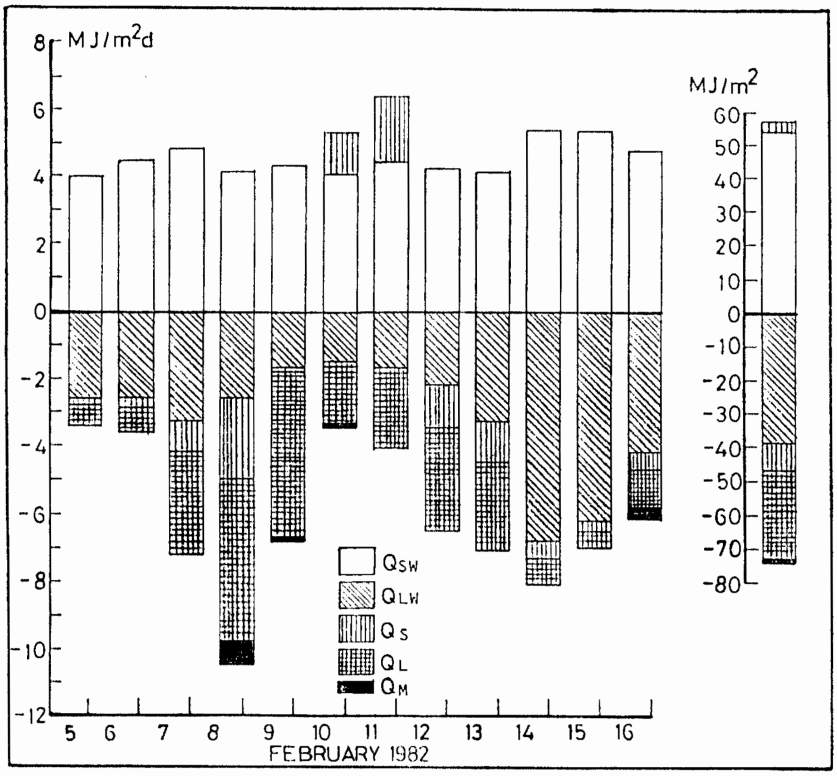

Fig. 4. Heat balance at Camp 2, daily values. The right-hand column shows total values for the period. Q_{SW}: net short wave radiation; Q_{LW}: net long wave radiation; Q_S: sensible heat flux; Q_L: latent heat flux; Q_M: heat used for melting snow.

over the period. This lowering (h) amounts only to 10% of the total accumulation and is both due to melting, compaction and evaporation. Neglecting the latter two causes, melting heat was calculated considering a mean surface density of 200 kg/m³.

Due to lack of precise field data only a rough estimation of heat balance can be done, but the relative contribution of each component should be a good approximation. Results are shown in Fig. 4. The main source is short wave radiation (94%), the rest is due to sensible heat flux during a relatively warm period between February 10 and 11. Heat sinks consist of long wave flux (52%), which is specially high during clear skies observed at the end of the period; evaporation accounts for 35% and is maximum during a strong wind period of two days (February 8 and 9); sensible heat accounts for 11% and melting only 2%. The difference between the total sources (57.2 MJ/m²) and the total sinks for the period (73.9 MJ/m²) should correspond to the remainder term X. Actually this difference is not so large (16.7 MJ/m²) and considering the lack of field data it is not possible to say whether it corresponds to the calculation error or in fact to the remainder term (*i.e.* heat capacity change, refreezing of meltwater or heat conduction to deep layers).

5. Shallow Drilling and Accumulation

At Camp 1 a snow pit of 1 m² was dug beside a crevasse down to 5.8 m and boring was continued down to 11.55 m with a hand drill. Density and grain size were measured, the results being shown in Fig. 5. The high density of 500 kg/m³ measured at the surface is probably due to wind packing. At a depth of 11.5 m the density already becomes 750 kg/m³, an evidence of the rapid metamorphism. Snow density averaged 550 kg/m³ over the whole depth (11.55 m). Several ice lenses were found, which shows that the melt and refreeze process of meltwater is important. RUNDLE (1973) located the wet-snow line at about 500–600 m a.s.l., so that Camp 1 (474 m) would be within the wet-snow zone.

During summer months rapid grain growth is expected due to high temperatures. In fact, grain size maximums occurred at 3.5, 7.4 and 10.7 m. These layers were estimated to correspond to the summer layers and thus the net balance was calculated. According to measurements on a pole left *in situ* at Camp 1, SCHMIDT (1983) reported a net balance of 2.8 m of snow for the period February 1982–January 1983, which agrees roughly with our measurements, but indicates that annual variations can be important, a fact already reported by RUNDLE (1973).

At Camp 2, hand-drilling was carried out down to 6 m. Big grain size was observed at 5.4 m and this was taken as the annual layer. As no density measurements exist for this layer, the average value of 550 kg/m³ of Camp 1 was used to calculate the water equivalent.

From field observation at the end of February, the equilibrium line altitude was estimated to be at 180 m, which is higher than the elevation of 60 to 120 m given by

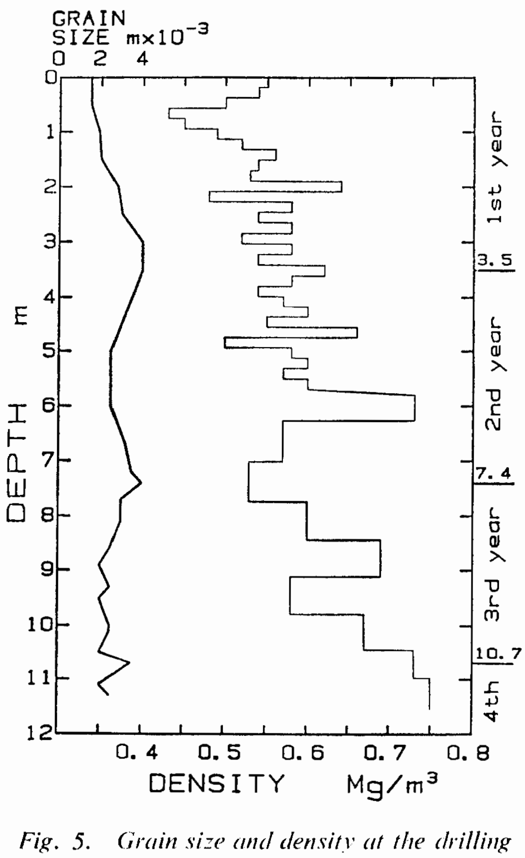

Fig. 5. Grain size and density at the drilling site of Camp 1.

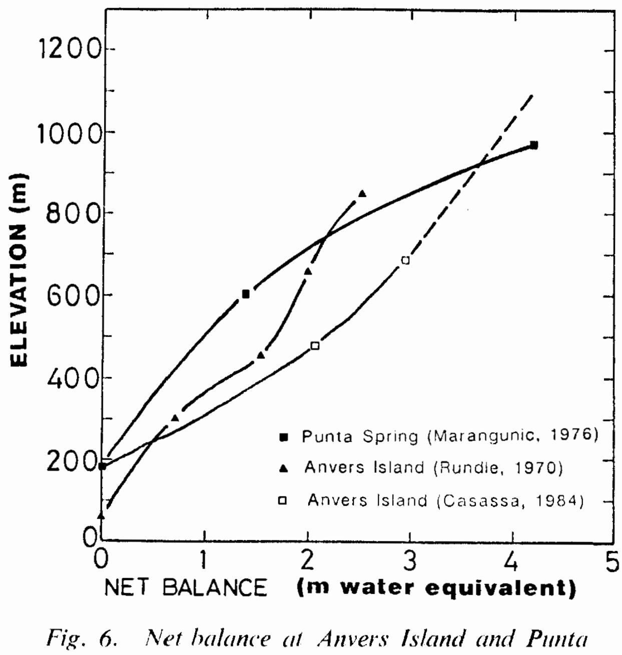

Fig. 6. Net balance at Anvers Island and Punta Spring, western coast of the Antarctic Peninsula.

RUNDLE (1973). In Fig. 6 net balance curves are shown for Anvers Island and Spring Point, located at 64°25′S on the mainland of the west coast of the Antarctic Peninsula. All data show the high accumulation which characterizes the western part of the Antarctic Peninsula.

6. Discussion

In the original calculation of heat balance done by the author in the thesis at Universidad de Chile (CASASSA, 1984), different expressions for heat fluxes were used. The new calculation presented in this paper (see Section 4. Heat balance) was carried out in order to introduce more updated expressions for the heat fluxes. For the original calculation the profile method was used to determine the turbulent fluxes and a different empirical equation was used for the long wave flux (AMBACH, 1960). Those results for the period were of 66.3 MJ/m² for the sources and 54.9 MJ/m² for the sinks. The sinks were composed of evaporation (44%), sensible heat (31%), long wave (21%) and melting (5%), while the sources were short wave (71%), long wave (25%) and sensible heat (4%).

Sinks were smaller in the original calculation (54.9 MJ/m²) than in the new calculation (73.9 MJ/m²). The main cause is that the profile method gave lower values of turbulent heat fluxes compared to the bulk equations, and that longwave flux by AMBACH's equation is positive for most of the period, whereas the Brunt-type equation used in the new calculation gives negative (upward) longwave fluxes for the whole period.

With the new field data available to estimate turbulent fluxes and longwave flux, it is not possible to ascertain which method is more reliable. However, the general conclusion from both computations is that the sinks are approximately equal to the sources. In fact, melting was quite small for the period, which supports the above statement. The main source in both computations was short wave radiation and the main sinks were evaporation and long wave radiation for the original computation and the new computation, respectively.

As respects the thermal regime of the glacier at Camp 1, snow temperatures at the drilling site were measured to be 0°C from surface to bottom (11.55 m). These temperatures cannot be regarded as representative, specially at deep layers, because they are affected by air temperature due to the location beside an open crevasse. In fact, RUNDLE (1973) measured englacial ten-meter temperatures of −0.8°C near the coast to −4.9°C inland at about 800 m a.s.l. Considering that RUNDLE gives a mean annual temperature of −3.3°C for Palmer Station (located on the coast), this ten-meter englacial temperature does not reflect the mean air temperature.

The englacial temperatures were estimated for Camp 1 by using a finite differential method to solve the energy equation:

$$\frac{\partial T}{\partial t} = b^2 \frac{\partial^2 T}{\partial z^2} ,$$

T is the englacial temperature, which varies with time t and depth z, b^2 is the thermal diffusivity of ice which depends on snow density ρ, heat conductivity k and specific heat of snow C_p:

$$b^2 = \frac{k}{C_p \rho}.$$

By inputting the density profile measured at Camp I (Fig. 5), b^2 could be calculated as a function of depth. The energy equation was solved estimating mean monthly temperatures at Camp I from data at Faraday Station, Lat. 65°15′S, Long. 64°15′W (SCHWERDTFEGER, 1970), and a lapse rate of 6.5°C/km (RUNDLE, 1973). Results of the computation are shown in Table 2. Englacial temperature at ten-meter is nearly constant at −8.2°C throughout the year, which is the estimated mean annual temperature at Camp I.

In the above calculation, water content was supposed to be zero and therefore the heat released by refreezing of meltwater was not taken into account. In fact, some melting occurs at Camp I so that refreezing in deep layers should play an important role to heat up the snow and therefore much higher englacial temperatures occur, as measured by RUNDLE. It is interesting to note that attempts by C. SWITHINBANK in 1967 to measure ice thickness using an ice radar desinged for cold ice were "largely unproductive" (RUNDLE, 1973). This was attributed to the relatively high englacial temperature.

A rough estimate of mass balance was made for an area of 46.5 km² as shown in Fig. 2. For this purpose a 1 : 25000 map of the Directorate of Overseas Surveys was used (Fig. 7). Ice divides were estimated from the map, passing through the summits of Mt. William and Mt. Hindson. The area was divided into nine zones according to their elevation, and from the net balance curve determined at Anvers Island (CASASSA, 1984) the annual mass balance was determined. As no net balance data are available in the ablation zone, it was estimated to be −0.5 m of ice per year on the ice front at the coast (50 m a.s.l.) and increases linearly to zero at the equilibrium line (180 m). The contribution of superimposed ice was neglected, even though RUNDLE (1973) observed a superimposed ice layer of a thickness of about 4 m at 90 m a.s.l., gradually thinning down to disappear at about 70 m a.s.l.

Table 2. Englacial temperatures calculated by solving the energy equation.

Month	Air temperature °C	5 m temperature °C	10 m temperature °C
January	− 2.8	−8.4	−8.3
February	− 3.2	−8.1	−8.4
March	− 4.3	−7.7	−8.4
April	− 7.7	−7.5	−8.3
May	− 9.6	−7.4	−8.2
June	−12.6	−7.6	−8.1
July	−14.2	−7.9	−8.0
August	−14.9	−8.3	−7.9
September	−12.0	−8.7	−7.9
October	− 8.2	−8.9	−8.0
November	− 5.7	−9.0	−8.1
December	− 3.8	−8.8	−8.3
Annual	− 8.2	−8.2	−8.2

* Surface temperature on the glacier is equal to air temperature.

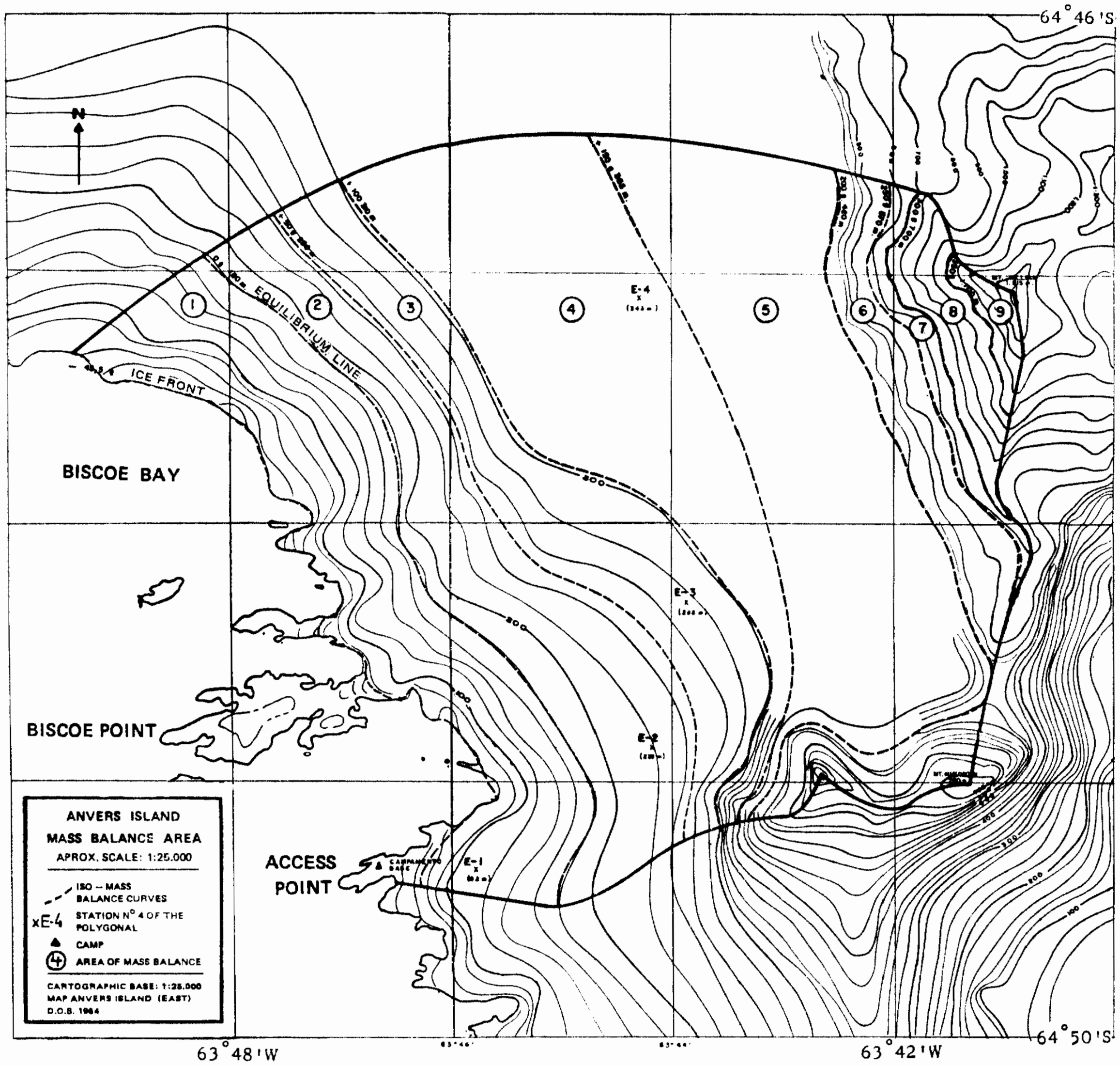

Fig. 7. Mass balance area. Dashed lines are iso-mass balance curves with annual net balance indicated in g and elevation in m.

The calculations show that the total positive balance was 6.15×10^{13} g·a^{-1} and the negative balance 1.5×10^{12} g·a^{-1}, resulting in a net positive balance of 6×10^{13} g·a^{-1}. RUNDLE (1970) carried out a mass balance study for an area of 300 km^2 which included our area and was limited by the ice divides joining Monaco Cape, Mt. Moberly and Lancaster Cape. RUNDLE concluded that the positive net balance was 3.8×10^{14} g·a^{-1}, of which 3.6×10^{14} g·a^{-1} calved at the front. The difference (0.2×10^{14} g·a^{-1}) was comparable to the possible errors and he concluded that the area was approximately in equilibrium. Considering that RUNDLE's area is 6.5 times greater than our mass balance area, RUNDLE's value for the net positive balance agrees quite well with our value, since it is 6.3 times greater. There was no field evidence of either glacier growth or shrinkage, so it can be assumed that the mass balance for the area of this study is roughly in equilibrium. In order to verify this both velocity and thickness of the ice at the front should be measured to calculate calving rate.

Acknowledgments

The Chilean Antarctic Institute (INACH) provided all the field logistics for this study; at The Institute of Low Temperature Science Dr. N. ISHIKAWA was very kind to guide the author in the calculation of heat balance and Mr. H. OZAWA provided thermal regime calculations. Finally, the author expresses his sincere thanks to the Professors of the original thesis presented at Universidad de Chile, Dr. Č. MARANGUNIĆ and I. H. PEÑA.

References

AMBACH, W. (1960): Investigations of the heat balance in the area of ablation on the Greenland Ice Cap. Arch. Meteorol. Geophys. Bioklimatol., Ser. B, **10**, 279–288.

CASASSA, G. (1984). Estudio de un sector del casquete de hielo de Isla Anvers, Península Antártica. Civil Engineer Thesis, Department of Civil Engineering, Universidad de Chile, Santiago (unpublished).

DEWART, G. (1971): Gravimeter observations on Anvers Island and vicinity. Antarctic Snow and Ice Studies II, ed. by A.P. CRARY. Washington, Am. Geophys. Union, 179–190 (Antarct. Res. Ser., Vol. 16).

HINZE, H. (1983): Zwischenbericht 1983 zu SE 313/4–3. Eis Bewegungsbestimmung mit Doppler-Satellitenmessungen im DFG-Schwerpunkt-programm Antarktisforschung. Institut für Erdmessung, Universität Hannover.

MARANGUNIĆ, Č. (1976): Contribution by Chile to the GAP Project. SCAR Working Group of Glaciology Meeting. Mendoza, Argentina.

MOTOYAMA, H. (1986): Studies of basin heat balance and snowmelt runoff models. Contrib. Inst. Low Temp. Sci., Hokkaido Univ., Ser. A(Phys. Sci.), **35**, 1–53.

OKE, T.R. (1978): Boundary Layer Climates. London, Methuen, 372 p.

RUNDLE, A.S. (1970): Snow accumulation and ice movement on the Anvers Island ice cap, Antarctica; A study of mass balance. International Symposium on Antarctic Glaciological Exploration (ISAGE), Hanover, New Hampshire, 3–7 September 1968, ed. by A.J. Gow *et al.* Cambridge, Heffer, 377–390 (IASH Publ. No. 86).

RUNDLE, A.S. (1973): Glaciology of the Marr Ice Piedmont, Anvers Island, Antarctica. Inst. Polar Stud., Rep., **47**, 237 p.

SCHMIDT, K.H. (1983): Arbeitsbericht zur Messkampagne 1983 DFG Se 313/4–2. Eis Bewegungsbestimmung mit Doppler-Satellitenmessungen. Institut für Erdmessung, Universität Hannover.

SCHWERDTFEGER, W. (1970): The climate of the Antarctic. Climates of the Polar Regions, ed. by S. ORVIG. Amsterdam, Elsevier, 253–355 (World Survey of Climatology, Vol. 14).

YAMANOUCHI, T. and KAWAGUCHI, S. (1984): Longwave radiation balance under a strong surface inversion in the katabatic wind zone, Antarctica. J. Geophys. Res., **89**, 11771–11778.

(Received November 1, 1988; Revised manuscript received September 8, 1989)