THE SNOW ACCUMULATION BUDGET AT HALLEY BAY IN 1959, AND ASSOCIATED METEOROLOGICAL FACTORS

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ABSTRACT. Frequent measurements at three different stake patterns covering disproportionate areas of the Brunt Ice Shelf near Halley Bay were used to study daily, monthly and annual accumulation, and local variation in accumulation. A study of pit stratigraphy, daily accumulation measurements and changes in snow density gave an estimate of settling, which was half the apparent ablation loss. The effects of evaporation at the surface, evaporation of drift snow in flight and drift snow losses by wastage out to sea are discussed in relation to the true ablation loss. True ablation losses amounted to about one-quarter of the measured gross accumulation, which is about one-third of the net annual accumulation.

DURING the period 1956–58 routine accumulation and temperature-depth measurements were made, and a 10 m. accumulation pit was dug as part of the meteorological programme (MacDowall, 1960, p. 153–61). Throughout 1957 and 1958 various short manhaul journeys were made to explore the extent of the ice shelf (Burton, 1960, p. 197–200).

Halley Bay is located in lat. 75°31′S. on the ice shelf abutting the Caird Coast of the Weddell Sea, a part of Coats Land (Fig. 1). The Filchner Ice Shelf lies about 200 miles (320 km.) to the south-west of Halley Bay, and west Dronning Maud Land is in lat. 71°S. to the north-east. Halley Bay is a shallow bight in the ice front, with a drift slope that has formed

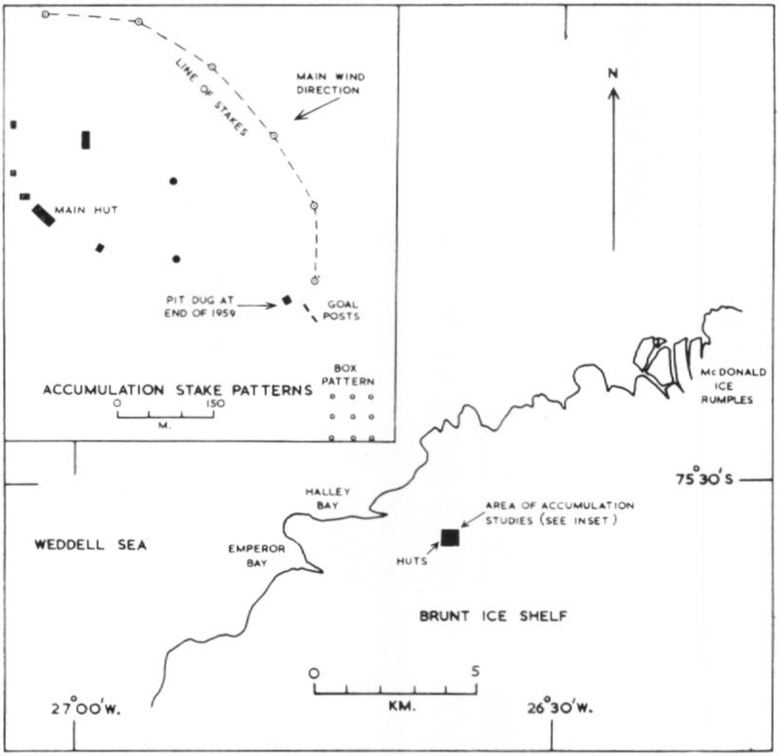

Fig. 1. Sketch map of the Brunt Ice Shelf near Halley Bay, showing the location of accumulation stakes in 1959. The coastline is based on a map prepared by L. W. Barclay in December 1959. The inset shows the respective positions of accumulation stake patterns.

in one of the larger clefts in the ice shelf, thus giving access on to the ice shelf. The base site is about 1·25 miles (2 km.) from the ice front, which is broken by a number of fissures along its west and north-western edges. The largest of the fissures is that forming Emperor Bay 2·5 miles (4 km.) west-south-west of the base. The ice shelf extends for about 30 miles (48 km.) westward from the main edge of the continent between long. 24°W. and long. 27°W., and its total area is about 1,500 sq. miles (3,900 km.^{2}). At its western edge the ice cliff heights vary between 6 ft. (18 m.) and 90 ft. (27 m.) above sea-level, but to the south this decreases to between 30 ft. (9 m.) and 50 ft. (15 m.). The elevation of the base area, within the error of the survey, remained constant at 94 ft. (28·6 m.) a.s.l. for the period 1956–59. During 1960 there were indications that sinking of the ice shelf might be taking place.

The Brunt Ice Shelf is maintained by outflow of the inland ice and the net annual snow accumulation on the ice shelf, the latter equivalent to about 37 g. water/yr. The true continental edge, i.e. the junction between the ice shelf and the inland ice, is orientated east-north-east to west-south-west. The ice of the high inland ridge (1,150 m. a.s.l.) spills over as a broad ice stream south of the Brunt Ice Shelf to form the Dawson-Lambton Glacier. Smaller, though active distributary streams, are the main source of supply to the ice shelf.

About 6 miles (9·6 km.) north-east of the station there is a prominent ice ridge and heavily disturbed area named the MacDonald Ice Rumples. This area is certainly grounded and it is thought to be the feature described and named by Shackleton (1919, end map) as the "Allan McDonald Glacier".

The Stake Patterns

During 1959 three stake patterns were used simultaneously:

i. A box pattern of nine stakes at 30 m. intervals.
ii. A line of six stakes in a semi-circle 0·75 miles (1·2 km.) long.
iii. Two horizontal bars at known levels above the snow surface in the form of "goal posts", by which name they will be referred.

The relative positions of these stake patterns are shown in Fig. 1. The first two were in use throughout 1958, but the "goal posts" were not introduced until 1959. Each "goal post" was set initially with the 3·5 m. long cross-bar 1·5 m. above the snow surface, and the snow level was measured from 13 points, each 25 cm. apart, along the length. Two such "goal posts" were erected; one was at the crest and the other in the trough of the main sastrugi pattern across the direction of the sastrugi, that is, at right angles to the prevailing wind (060°–080° true). The sastrugi wave-lengths were varied but they appeared to form harmonics of the main wave-length which was about 44 m. Accordingly, the "goal posts" were set with centres 22 m. apart.

The argument for such an arrangement was that the main sastrugi would maintain the same relative separation, even when sastrugi migration occurred, and thus the two "goal posts" would be measuring complementary features on the surface. The comparability between this type of measurement with that given by a more conventional stake pattern was borne out by experience. For example, the standard deviation of accumulation recorded at the box pattern was 4·9 cm. for the period March–December 1959. Over the same period and for the same dates the standard deviation of the combined "goal posts" was 4·7 cm. Both these values (σ_s) are the measure of the surface inequalities, as opposed to errors in measurement (σ_m), and show that the "goal posts" are no better or worse than a set of nine stakes over a larger area, provided that the period is of reasonable length. For shorter periods, σ_s is greater for the "goal posts". Thus, for the month of October σ_s ("goal posts") was 8·6 cm. compared with 7·5 cm. for σ_s (box pattern).

The variation of accumulation at "goal posts" A and B (Fig. 2) illustrates the variability of surface levels only 20 m. apart and shows the inaccuracy of single "goal post" observations (A or B) as a basis for daily observations. This is even more true of a single stake. The mean accumulation of the two "goal posts" compared reasonably well with the other patterns (Fig. 3), and after a trial period of four months regular readings of the single line of stakes were discontinued.

Readings were made daily when conditions were changing rapidly but less frequently in calm weather.

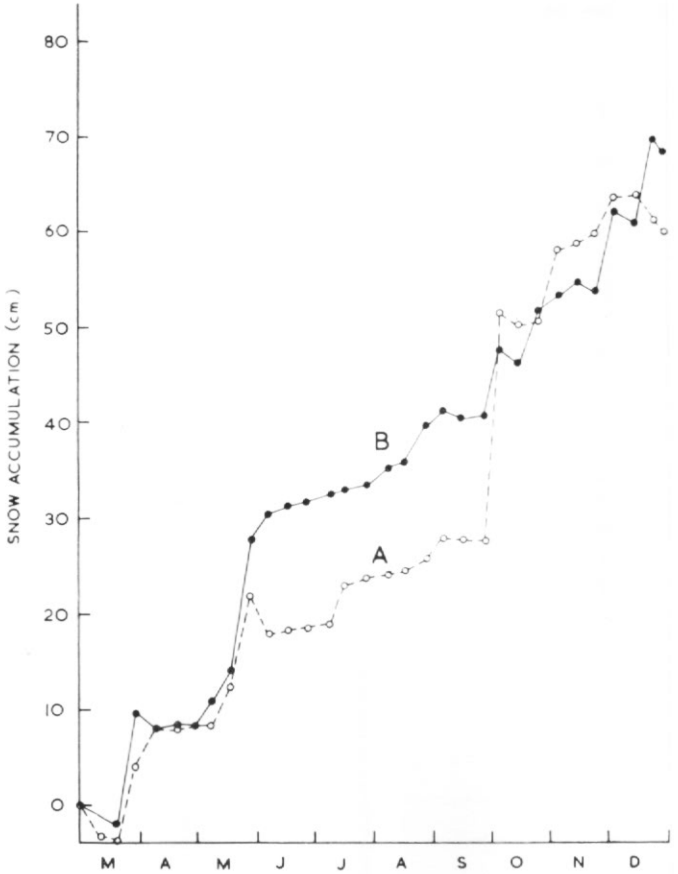

Fig. 2. Comparison of net snow accumulation at "goal posts" A and B.

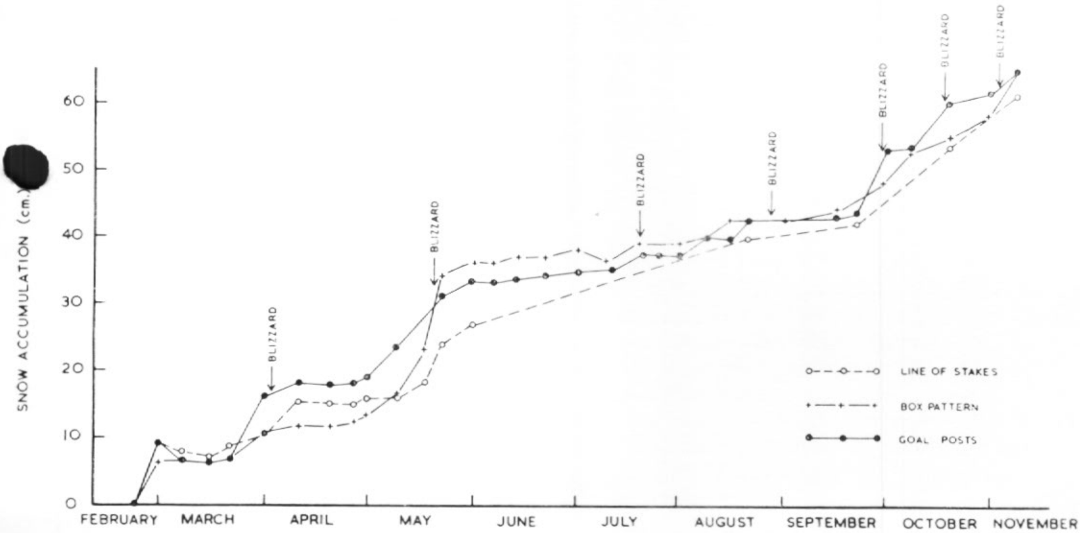

Fig. 3. The measurement of snow accumulation by means of three different stake patterns.

Accuracy of Stake Measurements

The accuracy of reading stakes depends on two factors: the wind strength at the time of reading, and the surface roughness. Several cases of 24 hr. accumulation were examined and it was found that the standard deviation of measurement (σ_m) at each "goal post" varied between 2 and 4 cm. when there was drift snow in association with a rapidly changing surface. This value of σ_m was independent of the sign of the net accumulation over the 24 hr. period. When the wind velocity reached 20 m./sec. (39 kt.) during actual periods of accumulation the snow surface was so indistinct that the readings were not considered valid, although even at these velocities the standard deviation was within the above limits.

When the surface was fairly smooth, such as that found after light snowfall in calm weather, the standard deviation (σ_m) fell to 0·1 to 0·2 cm. With a very uneven post-blizzard surface σ_m for the 13 points of each "goal post" was still as low as 0·2 to 0·3 cm.

It follows that the probable error in the measurement of accumulation was between 0·07 and 0·20 cm. for calm conditions (probable error = 0·67 standard deviation), and 1·3 to 2·7 cm. in rapidly changing surface conditions. Since accumulation values are dependent on two successive readings at the stakes, the probable error in accumulation is given by $2^{\frac{1}{2}}$ multiplied by the uncertainty of a single reading. Thus, the uncertainty of a single reading at a stake is given by the probable error of accumulation divided by $2^{\frac{1}{2}}$, which is

$\pm 0{\cdot}05$ to $\pm 0{\cdot}14$ cm. in calm conditions (approx. 0·1 cm.)

and $\pm 1{\cdot}4$ to $\pm 2{\cdot}8$ cm. in stormy conditions (approx. 2·5 cm.).

Therefore, the accuracy that can be claimed for individual stake readings changes by a factor of 25 from calm to stormy conditions.

The problem of the accuracy with which a stake can be measured leads naturally to the amount of surface level change that can be accepted as significant. By far the simplest method is to assume that the probable error in accumulation is the level of significance. In calm conditions 0·1 cm. is significant but in blizzard conditions the change should exceed 2·7 cm. for each "goal post". If one measurement is made in stormy conditions and the next in calm weather the probable error will be $(0{\cdot}1+2{\cdot}5)^{\frac{1}{2}}$ cm. = $\pm 1{\cdot}6$ cm., which will be the level of significance in this case.

The levels of significance depend not only on the wind strength and general storminess but also on the hardness of the surface. With a well-compacted surface a significant level as low as 0·1 cm. in stormy conditions can be accepted, provided that general deflation is in progress with no fresh accumulation. If varying conditions are allowed for, the day to day measurements at the "goal posts" were valid at all times. These daily values and those of the original line of stakes have been used to determine the amount of ablation and settling that occurred throughout one year.

Monthly Net Accumulation

The net monthly accumulation given by the box pattern of stakes and the values found from daily measurements (line and "goal posts") are compared in Table I. It will be seen that the mean wind speed has little effect on the standard deviation (σ_s) of monthly accumulation. The sole factors operating are the weather and surface conditions on the actual days of measurement. Clearly, there is a difference between the accumulation at the two sites, which by the end of the year almost vanishes. This may be due to the migration of long waves across the ice shelf, or the settling of one set of stakes relative to the other. The effect of the latter may be deduced. From March onwards the "goal posts" were in operation and, if there was relative settling, accumulation at the "goal posts" (*B* in Table I) would be expected to be in excess of the box pattern values (*A* in Table I) for a few months. Similarly, from October onwards, with a new box pattern *A* would be expected to be greater than *B*. Referring to the values of $A-B$ in Table I, it is found that although from October onwards *A* is greater than *B*, as expected, there are inconsistencies in the size of *B* relative to *A* in the winter months. Even allowing for relative settling of approximately 1 cm./month ($A-B$) varies from negative to positive in a periodic fashion giving support to the hypothesis of long migratory waves in the surface of the ice shelf.

The monthly values for accumulation give only a crude representation of the actual snow-

TABLE I. MONTHLY ACCUMULATION AT HALLEY BAY, 1959

Month	*Monthly Net Accumulation*				*Mean Surface Density*	*Mean Wind Speed*
1959	*A* (cm.)	σ_s (cm.)	*B* (cm.)	*A–B* (cm.)	(g./cm.^{3})	(m./sec.)
January	13·2	8·4	15·7(*L*)	−2·5	0·36	7·15
February	7·6	5·5	9·4(*L*)	−1·8	0·30	4·28
March	4·2	5·2	6·4	−2·2	0·31	4·59
April	2·1	4·8	1·8	+0·3	0·33	6·85
May	23·0	5·4	15·6	+7·4	0·36	10·00
June	2·1	0·5	1·4	+0·7	0·10 (hoar frost)	3·30
July	1·0	3·4	2·4	−1·4	0·37	7·94
August	5·5	4·9	5·6	−0·1	0·30	5·87
September	5·6	5·9	10·6	−5·0	0·34	7·00
October	11·4	7·5	8·3	+2·9	0·37	7·67
November	17·3	5·5	14·4	+2·9	0·35	7·42
December	−0·7	5·8	−2·2	+1·5	0·35	8·09
Year	92·3	5·6	89·4	+2·9	0·34	6·70

A. Measurements made at the "box pattern".
B. Measurements made at the "goal posts".
L. Measurements made at the line of stakes.

fall. Not only is it necessary to account for ablation by evaporation and deflation, but it is also important to gauge the *apparent ablation* losses caused by compaction of the snow. In addition, a true water equivalent cannot be derived by multiplying net accumulation by the mean surface density for the month, because the snow surface density bears no relation to the final density throughout the compacting surface layers and the resultant net accumulation. The net annual accumulation was taken as the mean of *A* and *B*, which was 91 cm. of snow.

SETTLING (OR COMPACTION)

LaChapelle (1959, p. 459–62) has described how errors in ablation measurements may be overcome by using stakes set at different depths. If S_1 and S_2 are the settling rates of compacting snow layers at two stakes set with their bases at different levels, then the settlement between the layers in which these stakes are embedded is given by

$$S_1 - S_2 = dh_1 - dh_2, \qquad (1)$$

if dh_1 and dh_2 are simultaneously recorded changes in snow level.

The introduction of a new box pattern of stakes interspersing the old box pattern led to a comparison of this nature. The two sets of stakes were embedded at depths of approximately 1 m. and 2·5 m., or about two annual layers apart, and they gave a settling rate of 0·04 cm./day or 14·6 cm./yr. The surface layer of 1 m. depth had a considerably greater settling rate as will be shown later.

Further estimates of settling were made by close scrutiny of the daily observations in conjunction with meteorological records. All negative values of accumulation (i.e. ablation) that occurred on days of drift snow or on warm sunny days were eliminated. A few days of apparent ablation remained, even though there had been slight snowfall during the preceding

24 hr. These changes were assumed to have been caused solely by settling and mean values are shown in Fig. 4. It should be noted that this is combined bulk and surface settling. Where no such cases were evident relative rates between stakes were used, if available. These latter values are an underestimate. It is clear from Fig. 4 that the settling rate in summer is much

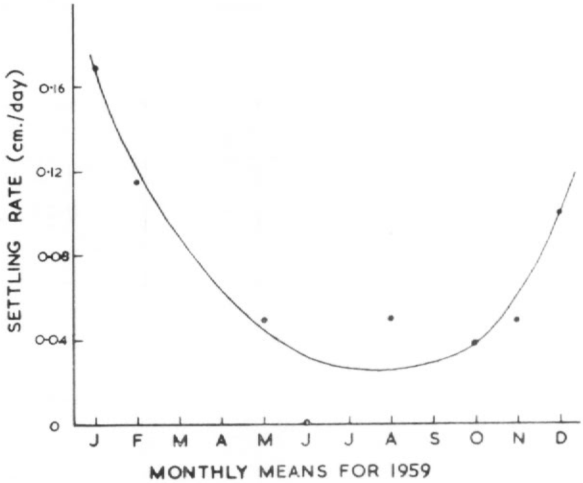

Fig. 4. Estimation of settling from daily stake measurements in 1959.

greater than that in winter, which was approximately 0·04 cm./day. Monthly settling rates derived from Fig. 4 are tabulated in Table II and give the total settling in the first metre for the year as being not less than 24 cm./yr.

TABLE II. APPROXIMATE SETTLING RATES DERIVED FROM A SCRUTINY OF DAILY STAKE MEASUREMENTS AND METEOROLOGICAL RECORDS, AND ALSO FROM THE COMPARISON BETWEEN STAKES BURIED TO DIFFERENT DEPTHS

Month	*Settling Rate*	
1959	(cm./day)	(cm./month)
January	0·16	5·0
February	0·12	3·4
March	0·09	2·8
April	0·06	1·8
May	0·04	1·2
June	0·03	0·9
July	0·03	0·9
August	0·03	0·9
September	0·03	0·9
October	0·04	1·2
November	0·06	1·8
December	0·10	3·1
Year	24 cm./yr.	

Having established the seasonal variability of settling, more refined methods may be used for measuring the settling, first as surface settling and then as bulk settling.

For the measurement of mass loss from the surface snow layer, changes of density and volume must be taken into account (LaChapelle, 1959):

$$\text{Mass loss } (A) = \int_0^{h_0} \rho(ht_0)dh - \int_0^{h_0} \rho(ht_1)dh. \tag{2}$$

The integral is over the layer from the surface (0) to a depth h_0, where $\rho(ht)$ has not changed within the time $(t_1 - t_0)$. dh is the apparent ablation at the surface, i.e. settling plus ablation. The amount of settling may be determined from the surface density samples. Thus, after a period of snowfall the surface layer compacts from an initial density ρ_0 to a new density ρ_1. Assuming the period between snowfalls as $(t_1 - t_0)$, the mass loss is given by

$$A \simeq \rho_0(ht_0)(h - h_0) - \rho_1(ht_1)(h_1 - h_0),$$

where h_1 are levels of the surface measured from a stake. h_0 is the level of the bottom of a new snow layer.

Therefore,

$$A \simeq \rho_0 H - \rho_1 H_1 = \rho_0 H - \rho_1(H - \Delta H),$$

where $H = h - h_0$ and $H_1 = h_1 - h_0$.

If there is no ablation, then

$$\Delta H = H\left(1 - \frac{\rho_0}{\rho_1}\right). \tag{3}$$

Replacing H by a (the amount of new snowfall), and denoting ΔH by S_1, then the surface settling is given by

$$S_1 = a\left(1 - \frac{\rho_0}{\rho_1}\right), \tag{4}$$

where ρ_0/ρ_1 is the compaction, i.e. the change in volume.

The most important factors affecting compaction are wind strength, temperature and sun. In Fig. 5 the pecked lines represent the change in density of the surface layer between snowfalls, or throughout settling periods. The full lines are running five-day means of wind speed. Sunshine is shown as a histogram. In all cases, the lower densities are those of fresh snow, and the surface layer in periods between any two points on the density curve have been treated using equation (4), as detailed in Table III. The effect of wind force as a compacting agent is especially clear for August, September and November. The total surface settling for 1959 amounted to 17 cm.

To obtain bulk (or secondary settling, S_2) it is necessary to compare the densities found by surface sampling with those found in the final annual layer by sampling from a pit dug at the end of the year.

At the end of one year the net accumulation recorded as centimetres of snow at a set of stakes should agree closely with a pit dug for comparison, because each has undergone the same surface and bulk settling and ablation. But, layer for layer, the densities recorded at the surface (ρ_1) will have increased to some new value (ρ_2). A plot of density curves for different snow depths for ρ_1 and ρ_2 (Fig. 6) shows that the general correspondence of density/depth curves becomes distorted at the lower end. This is because the surface densities are recorded with a time lapse, whilst those in the pit are at one fixed time. It was possible to date some of the bands in the pit from a knowledge of the daily accumulation, and this gave a measure of the time lapse distortion. The significant feature of the accumulation pit density/depth profile is that the density reaches the mean annual values only 12 cm. below the surface, showing that the majority of the bulk settling takes place directly a surface layer has been covered by a new layer.

The depth of the annual layer given by the pit was $92 \cdot 7 \pm 0 \cdot 7$ cm., at which level there was a well-defined ice band dating from 20 January. This date was taken as a datum, because up to that date ablation had been continuous. For the same period the annual accumulation, measured daily, was $92 \cdot 8$ cm. with a mean surface density (ρ_1) of $0 \cdot 35$ g./cm.^{3} (Table III).

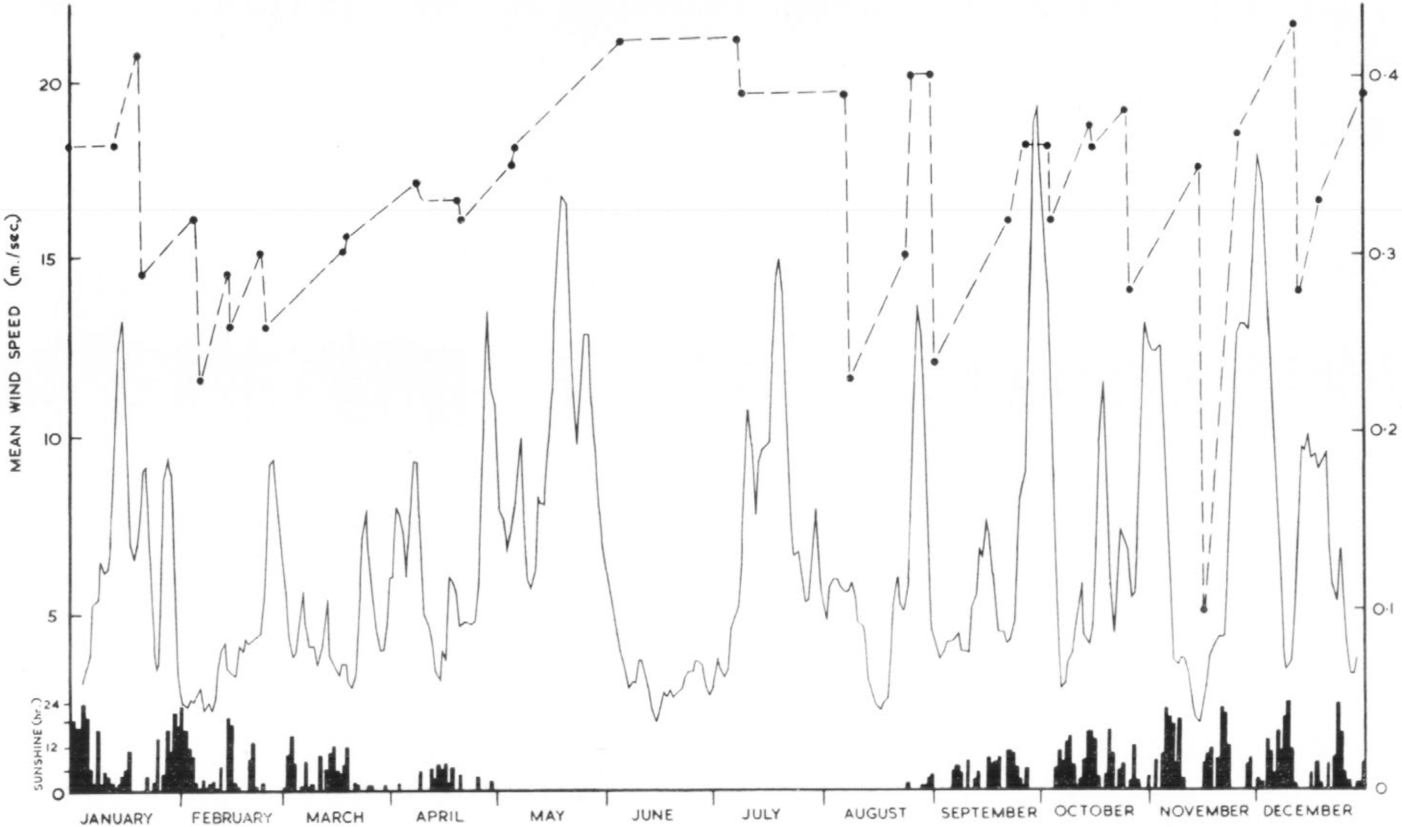

Fig. 5. Comparison of snow surface density (pecked line) with mean wind speed (solid line). The frequency of sunshine is also shown.

Table III. Accumulation, Ablation and Settling for Irregular Periods throughout 1959

Date 1959	*Gross Accumulation* (G) (cm.)	*Apparent Ablation* (B) (cm.)	*Net Accumulation* (N) (cm.)	*Snow Surface Density* *Initially* (ρ_1) (g./cm.^{3})	*Finally* (ρ_2) (g./cm.^{3})	*Compaction* (ρ_1/ρ_2)	*Surface Settling* (S_1) (cm.)	*Ablation plus Bulk Settling* (S_2) ($B-S_1$) (cm.)	*Gross Water Equivalent* ($G\rho_1$) = ($N-B+S_1)\rho_2$ (g.)
1–12 January	+0·2	−2·6	−2·4	0·33*	0·37	0·88*	0·02*	2·6	0·07
13–20 January	+0·7	−1·7	−1·0	0·36	0·41	0·88	0·1	1·6	0·25
21 January–5 February	+20·9	−2·4	+18·5	0·29	0·32	0·90	2·1	0·3	6·06
6–14 February	+2·3	−1·5	+0·8	0·23	0·29	0·79	0·5	1·0	0·53
15–24 February	+0·9	−1·4	−0·5	0·26	0·30	0·87	0·1	1·3	0·23
25 February–18 March	+12·4	−6·0	+6·4	0·26	0·30	0·87	1·6	4·4	3·22
19 March–5 April	+14·2	−1·8	+12·4	0·31	0·34	0·91	1·3	0·5	4·40
6–9 April	Nil	−1·5	−1·5	0·34	0·34	1·00	Nil	1·5	Nil
10–20 April	+1·5	−1·0	+0·5	0·33	0·33	1·00	Nil	1·0	0·50
21 April–5 May	+0·8	−0·8	0·0	0·32	0·35	0·91	0·1	0·7	0·26
6 May–5 June	+19·8	−4·4	+15·4	0·36	0·40	0·90	2·0	2·4	7·13
6 June–8 July	+1·7	−1·4	+0·3	0·40	0·40	1·00	Nil	1·4	0·68
9–28 July	+5·2	−0·8	+4·4	0·37	0·37	1·00	Nil	0·8	1·92
29 July–7 August	+0·8	−0·6	+0·2	0·37	0·37	1·00	Nil	0·6	0·29
8–24 August	+6·4	−2·4	+4·0	0·23	0·30	0·77	0·9	1·5	1·47
25–31 August	+0·6	−0·1	+0·5	0·38	0·38	1·00	Nil	0·1	0·23
1–27 September	+5·1	−4·2	+0·9	0·24	0·32	0·75	1·3	2·9	1·22
28 September–3 October	+18·9	−10·4	+8·5	0·36	0·36	1·00	Nil	10·4	6·80
4–15 October	+6·8	−1·7	+5·1	0·32	0·37	0·87	0·9	0·8	2·18
16–25 October	+3·2	−0·4	+2·8	0·36	0·38	0·95	0·2	0·2	1·15
26 October–15 November	+8·2	−3·2	+5·0	0·28	0·35	0·80	1·6	1·6	2·30
16–20 November	+1·4	−1·0	+0·4	0·10	0·32	0·31	1·0	0·0	0·14
21 November–12 December	+15·6	−9·7	+5·9	0·37	0·43	0·86	2·2	7·5	5·77
13–19 December	+0·7	−1·1	−0·4	0·28	0·33	0·85	0·1	1·0	0·20
20 December–1 January	+4·7	−1·5	+3·2	0·33	0·38	0·87	0·6	0·9	1·55
Year	+153·0	−63·6	+89·4	0·31	0·35	0·89	16·6	47·0	48·55

* Estimated value. $S_1 = G(1-\rho_1/\rho_2)$.

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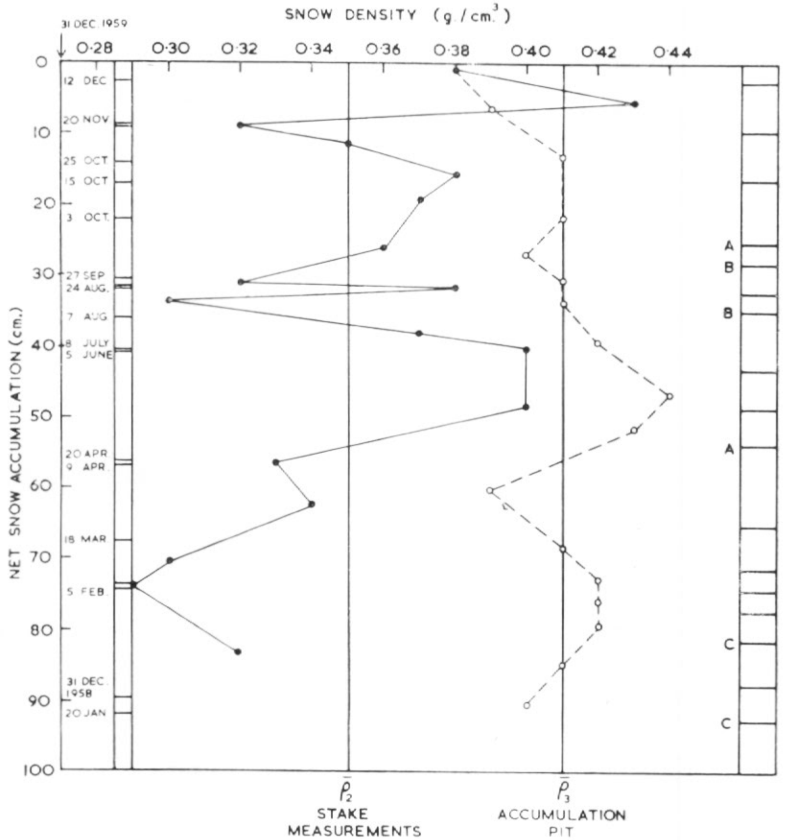

Fig. 6. Density/depth profile of accumulation pit compared with surface density/net accumulation profile given by stake measurements.
$\bar{\rho}_2$ Mean surface density prior to a new snowfall.
$\bar{\rho}_3$ Mean density of layers in the accumulation pit.
A. Change in firn crystals.
B. Wind crust.
C. Ice band.

If N is the net accumulation in centimetres, then the bulk settling is derived by the equivalence of grams water accumulation

$$(S_2+N)\rho_1 = N\rho_2 \text{ g.}$$

or

$$S_2 = N(\rho_2-\rho_1)/\rho_1. \tag{5}$$

Thus, for the period 20 January to 31 December

$$S_2 = 92\cdot 8\times\frac{0\cdot 06}{0\cdot 35} = 15\cdot 9 \text{ cm.}$$

or 17·1 cm./m./yr.

For the period 1–12 January the bulk settling appears to have been about 0·2 cm./day (Fig. 3), i.e. an extra 2·4 cm., and from 13–20 January the rate was about 0·1 cm./day, an additional 0·8 cm. Therefore, the total bulk settling amounted to 19·1 cm./yr. at a mean rate of 0·052 cm./day for the surface layer that contracted to a net accumulation of 91 cm.

Bulk settling may also be given by

$$S_2 = G_1\left(1-\frac{\rho_1}{\rho_2}\right), \tag{6}$$

where G_1 is gross accumulation at a density of ρ_1. Gross accumulation is defined here as the sum of the daily positive accumulation. The total of surface plus bulk settling amounted to 36 cm. compared to a mean net accumulation of 91 cm. for the year.

At deeper levels the settling rate is slower. MacDowall and Blackie (1960, p. 64) state that over a period of 21 months the geomagnetic hut at Halley Bay subsided by 25·4 cm. relative to piles driven to a depth of 20 ft. (6·1 m.). The settling for the last 13 months was 11·2 cm., or 1·72 cm./m./yr. for a layer at a depth of 1–7 m. From a comparison of the two box patterns of stakes it has been shown that the settling between a depth of 1 and 3 m. was approximately 14·6 cm./yr., or 7·3 cm./m./yr. For the first metre there is a bulk settling of 17·1 cm./m./yr. The appropriate compaction ratios, ρ_1/ρ_2, derived from $S = 100\left(1-\frac{\rho_1}{\rho_2}\right)$ are 0·983, 0·927 and 0·827. In a similar approach, Schytt (1958, p. 55) gave a settling rate of 1·57–1·64 cm./m./yr. for the 2–11 m. layer at Maudheim, which is consistent with the Halley Bay value for the 1–7 m. layer.

Evaporation Losses

The loss by direct evaporation from the snow surface presents a particularly difficult problem. With the availability of humidity data and net flux radiation records it is possible to arrive at an estimate of the monthly evaporation rates, though only by making some broad assumptions.

First, consider the humidity lapse from the surface to the level of the meteorological instruments (approximately 1·25 m.). Rider (1954, p. 499) has shown the equivalence of the eddy diffusivities of water vapour (K_v) and momentum (K_m) throughout large ranges of stability. Pasquill (1949) assumed the equivalence and showed that observed values of K_v in near-neutral and unstable conditions agreed closely with values of K_m calculated on the basis of the logarithmic wind-profile law,

$$u_2 = u_{*}k \log_e(z/z_0). \tag{7}$$

Stable conditions did not produce a correspondence, because the logarithmic law is no longer valid (Rider, 1954, p. 500).

For an ice shelf station such as Maudheim inversion conditions prevail for nine months of the year within the 2·5 m. layer next to the surface (Liljequist, 1957*b*, p. 277, table 2). Isothermal conditions were prevalent at Maudheim in November, and in January and December there were lapse rates of 0·2°/2·5 m. and 0·1°/2·5 m., respectively. Halley Bay is similar to Maudheim in many respects and it is reasonable to accept that average Maudheim data are applicable to Halley Bay, in respect of the stability of the surface air layer.

Applying Pasquill's reasoning to the problem of the vapour pressure gradient at Halley Bay, it will be assumed that because the summer temperature gradients are near neutral or are unstable then the logarithmic wind law holds and surface evaporation is given by

$$E = \frac{A}{T}\,u_1(e_1-e_2)\times 10^3 \text{ g./cm.}^2\text{/sec.}, \tag{8}$$

where

$$A = \frac{k^2M(1-u_1/u_2)}{R(\log_e z_2/z_1)^2}, \tag{8a}$$

and e = Vapour pressure in mb.
u = Wind velocity in cm./sec.
z = Height in cm.
k = Von Karman's constant, 0·41.
M = Gram molecular weight of water = 18.
R = Universal gas constant = $83\cdot15\times10^6$ erg/degree.
T = Air temperature in ° K.
Suffixes 1 and 2 refer to the levels of z, level 2 being the higher.

u_1 may be replaced by the friction velocity u_*, if z_1 is replaced by z_0, the roughness length as in equation (7). A comprehensive study by Liljequist (1957*a*, p. 200) provides both z_0 and u_*. z_2 will be taken as the level of the humidity measurements which was approximately 125 cm., and the wind at this level deduced from meteorological data using the logarithmic law.

To obtain the vapour pressure gradient it will be reasonably assumed that the surface is at saturation with respect to ice at the ambient temperature. The mean vapour pressure for each hour over one month at 1·25 m. is compared with the saturation vapour pressure for the mean hourly air temperature for the month. The error in assuming isothermal conditions at the surface is less than 2 per cent of the saturation vapour pressure.

Applying the various contents and the values for u_* and z_0, the evaporation over one hour is given by

$$E_h = 0{\cdot}231 \frac{u_2(e_1 - e_2)}{T} \times 10^{-3} \text{ g./cm.}^2\text{/hr.,} \tag{9}$$

or for one month

$$E_m = N \sum_{h=0}^{h=24} E_h \simeq \frac{0{\cdot}231 N u_2}{T} \left[\sum_0^h (e_1 - e_2) \times 10^{-3} \right] \text{ g./cm.}^2\text{/month} \tag{10}$$

where N = the number of days in the month.

The variation of air temperature at any fixed hour throughout any month is less than $\pm 10^\circ$ C, so that the probable error in using the mean monthly air temperature T is 2·5 per cent. The error in using the mean monthly wind speed u_2 is ± 10 per cent. The overall error using such data is thus of the order 10–15 per cent. There is, however, an additional error which will consistently give evaporation rates too low. The mean vapour pressure is not the same as the vapour pressure at the mean temperature, and by assuming the equivalence of $\frac{\sum^n e}{n}$ to e_T, the value of e_1 is made smaller by 10–20 per cent, so that $(e_1 - e_2)$ is reduced hence giving an underestimate of evaporation $(e_1 > e_2)$, and conversely an overestimate of condensation $(e_2 > e_1)$.

January and December were the only months when $(e_1 - e_2)$ was positive for all hours. The other summer months were positive in daylight hours and negative at night. Negative values of $(e_1 - e_2)$ are manifest as condensation in the form of hoarfrost which is swiftly removed by any wind in excess of 7 m./sec. If it is assumed that this happens, the negative values can be ignored and the positive values of $(e_1 - e_2)$ are summed, giving the total surface evaporation (E_{m_1} in Table IV).

TABLE IV. MONTHLY EVAPORATION AT HALLEY BAY, 1959

Month	*Air Temperature*			*Evaporation*	
1959	(°K)	$\Sigma(e_1 - e_2)$ (mb.)	$\bar{u}_2$ (m./sec.)	Em_1 (g.)	Em_2 (g.)
January	269	8·32	5·8	1·28	1·28
February	264	1·57	3·5	0·14	0·05
March	254	0·05	5·6	negligible	nil
November	261	0·54	6·0	0·08	nil
December	268	9·31	6·5	1·65	1·65
TOTAL				3·15	2·98

e_1 Mean hourly value of saturation vapour pressure at the surface at indicated temperature.
e_2 Mean hourly value of vapour pressure at 1·25 m.
$\bar{u}_2$ Mean monthly wind speed at 1·25 m.
$\Sigma(e_1 - e_2)$ Sum of the hourly differences of vapour pressure, i.e. daily total.
Em_1 Evaporation using only positive values $(e_1 - e_2)$.
Em_2 Evaporation using the 24 hr. values $(e_1 - e_2)$.

Quantitative measurement of hoarfrost is at the best rather subjective, and observation showed that overnight condensation soon evaporated during the summer days when the wind was not strong enough to remove the frost. It is, therefore, necessary to consider the negative values of (e_1-e_2). November is eliminated as a month with any net significant surface evaporation, and the February contribution is reduced to one-third of the gross surface evaporation (E_{m_1}). January and December remain unchanged (E_{m_2} in Table IV).

On this reckoning, the total summer evaporation from the snow surface amounts to 3·0 g. of water. Allowing for the low values of e_1, this may be increased by 20 per cent to 3·6 g. with an overall error of approximately ± 20 per cent.

The second method of calculating evaporation depends on the net flux of radiation and the sub-surface temperature profiles.

The flux of energy at the snow surface is defined by

$$E_{(0)}+Q_{(z)}+B_{(0)}+W_{(0)} = 0, \tag{11}$$

where $E_{(0)}$ is the net flux of radiant energy, $Q_{(z)}$ is the vertical transfer of heat in the air by eddy diffusion at height z, $B_{(0)}$ is the flux of heat in the snow reaching the surface, and $W_{(0)}$ are heat gains or losses due to evaporation or condensation.

At any particular moment the radiation flux at the surface, $E_{(0)}$, is not the same as that given by the net flux radiometer at some level above the surface. But over a long period Δt, the integrated values at both levels, will be substantially the same. Thus, the monthly net flux radiation totals may be applied to a balance equation at the surface.

According to Johnson (1954, p. 157) $B_{(0)}$ may be represented as

$$B_{(0)} = -\rho_s C_s(\Delta T \Delta Z/\Delta t), \tag{12}$$

where $\Delta T \Delta Z/\Delta t$ represents the area enclosed by two temperature/depth profiles of time, Δt, separation. ρ_s is the density of the snow throughout the depth, ΔZ, and C_s is the specific heat of ice.

If L_s is the latent heat of sublimation and $\Delta m/\Delta t$ is the mass of ice evaporation in time, Δt then

$$W_{(0)} = -L_s \Delta m/\Delta t. \tag{13}$$

Any melting that takes place is either refrozen at a different level in the snow with no resultant loss of heat, or it is vaporized from the liquid state. The process of melting followed by vaporization is equivalent to direct evaporation from the solid state as far as the amount of heat exchange is concerned, and is therefore included in the term, $-L_s\Delta m/\Delta t$.

To ascribe values to $Q_{(z)}$ is difficult, depending as it does on the stability of the atmosphere near the ground. The turbulent flux of heat is defined by

$$Q_{(z)} = -K_H \rho C_p\left(\frac{\partial T}{\partial z}+\Gamma\right). \tag{14}$$

K_H is the coefficient of eddy conductivity of heat at height z, and is only constant for a particular set of meteorological conditions; it varies by several orders of magnitude depending on the stability of the air layer. The measure of stability is given by $\left(\frac{\partial T}{\partial z}+\Gamma\right)$, where $\frac{\partial T}{\partial z}$ is the temperature gradient (negative if there is an inversion) and Γ is the adiabatic lapse rate (very nearly 0·01° C/m.). ρ is the air density and C_p the specific heat of the air at constant pressure. When $\left(\frac{\partial T}{\partial z}+\Gamma\right)$ is zero, i.e. in neutral condition, $Q_{(z)}$ is zero. When $\left(\frac{\partial T}{\partial z}+\Gamma\right)$ is negative the flux of heat is positive towards the surface. Conversely, there is a turbulent loss of heat when $\left(\frac{\partial T}{\partial z}+\Gamma\right)$ is positive. There is an additional control provided by K_H, which for positive flux is much smaller than for negative flux, so that for similar values of $\left(\frac{\partial T}{\partial z}+\Gamma\right)$ but of opposite sign there will be disproportionate amounts of heat involved.

An analysis of the temperature gradients at Halley Bay for 01·00 and 13·00 hr. local time gives a rough idea of the frequency of each type of flux (Table V). There were twice as many cases of incoming flux as outgoing flux. But, since K_H is greater for negative flux than for positive flux, there is a case for assuming that $Q_{(z)}$ was near zero for the summer

TABLE V. ESTIMATED NUMBER OF CASES OF POSITIVE AND NEGATIVE FLUX OF TURBULENT HEAT BASED ON TEMPERATURE GRADIENTS AT 01.00 AND 13.00 hr. LOCAL TIME

Month 1959	*Positive Flux Incoming* $\left(\frac{\partial T}{\partial Z}+\Gamma\right)<0$	*Negative Flux Outgoing* $\left(\frac{\partial T}{\partial Z}+\Gamma\right)>0$	*No Flux* $\left(\frac{\partial T}{\partial Z}+\Gamma\right)=0$
January	30	13	19
February	30	16	10
November	25	18	17
December	22	11	29
TOTAL	107	58	75

period. This would be in agreement with the assumption regarding temperature gradients used in calculating evaporation from the vapour pressure gradient.

Neglecting $Q_{(z)}$, the mass loss by evaporation is

$$\frac{\Delta m}{\Delta t}=\frac{E_{(0)}-\rho_s C_s}{L_s}(\Delta T\Delta z/\Delta t). \tag{15}$$

The results of computing equation (15) are given in Table VI. Considering the approximations and assumptions made, there is a fair agreement in the magnitude of the evaporation calculated by the two methods. On the other hand, neglecting $Q_{(z)}$ must introduce some error into the evaporation determined by the energy balance. If allowance is made for a positive turbulent flux of heat of the order of 400 cal./cm.²/month, based on estimates from Maudheim (Liljequist, 1957, p. 292), the monthly evaporation rates are increased by 0·59 g./cm.²/month, so that the overall evaporation increases from 5·0 to 7·4 g. The excess evaporation given by the larger amount may be in the form of evaporation of drift particles in flight, and this possibility will be discussed at some length in the next section.

The main inconsistency in the vapour pressure gradient method is the assumption that the average temperature gradients are applicable for all hours. That they are not is highlighted

TABLE VI. EVAPORATION FROM NET FLUX RADIATION AND TEMPERATURE PROFILES

Month	*Net Radiation Flux*, $E_{(0)}$	$\rho_s C_s \frac{\Delta T}{\Delta t}\Delta Z = B_{(0)}$	*Evaporation Rate*, $\Delta m/\Delta t$ $Q_{(z)}=0$	$Q_{(z)}=400$ cal./cm.²/month
1959	(cal./cm.²/month)	(cal./cm.²/month)	(g./cm.²/month)	(g./cm.²/month)
January	2,001	565	2·12	2·71
February	344	51	0·43	1·02
March	−661	−296	−0·54 (condensation)	0·05
November	1,048	524	0·77	1·36
December	1,712	66	1·64	2·23

$\rho_s \simeq 0\cdot5$ g./cm.² density 0 to 12 m. depth.
$C_s = 1\cdot0$ cal./g.
Negative flux is outwards from the surface.

TABLE VII. METEOROLOGICAL FACTORS AFFECTING VISIBILITY IN DRIFT SNOW CONDITIONS

Temperature Range (° C)	*Visibility Range*											
	30–150 yd. (27–137 m.)			180–600 yd. (165–549 m.)			700–1,500 yd. (640–1,372 m.)			4,400–22,000 yd. (4,023–20,117 m.)		
	*Relative Humidity** (per cent)	*Wind Speed*† (m./sec.)	*Number of Observations*	*Relative Humidity** (per cent)	*Wind Speed*† (m./sec.)	*Number of Observations*	*Relative Humidity** (per cent)	*Wind Speed*† (m./sec.)	*Number of Observations*	*Relative Humidity** (per cent)	*Wind Speed*† (m./sec.)	*Number of Observations*
0·0 to −10·0	100·0	18·4	17	98·1	13·8	42	96·5	12·3	17	95·3	19·5	39
−10·1 to −20·0	99·8	16·5	29	100·2	12·9	46	98·3	11·7	16	96·9	18·8	32
−20·1 to −30·0	101·8	14·9	26	102·0	10·4	41	100·6	9·9	31	97·5	17·0	29
−30·1 to −40·0	—	—	—	105·6	9·1	16	102·5	8·8	5	102·8	14·3	5

* At 1.25 m. above ice level.   † At 8 m. above ice level.

in Table V. Nevertheless, it is felt that the calculated values of surface evaporation are an indication of the magnitude, corroborated from energy balance considerations. Because of the uncertainties, the summer evaporation will be assumed to be between 3 and 5 g. of water, say about 4 g. This amounts to about 10 cm. of surface snow. The evaporation throughout the remainder of the year was negligible, and in fact condensation was frequent in the form of hoarfrost, which was removed by moderate winds.

Drift Snow Losses

Drift snow losses can occur in two ways: wastage by removal out to sea, and by evaporation whilst airborne. The latter suggestion by MacDowall (1960, p. 156–57) has been partly confirmed by work at Halley Bay in 1959. This work is summarized in Table VII, where the mean relative humidities with respect to ice and the wind speeds at a height of 8 m. are compared at different temperatures and visibility ranges occurring with drifting snow. The data were extracted from all cases of drift snow when no snow was falling. The instrument for measuring the humidity was an ice and dry bulb ventilated thermocouple psychrometer designed by Tribble (MacDowall and Ellis, in press). The output of the psychrometer was 900 $\mu V/° C$, so that a $0 \cdot 01° C$ elevation or depression of the ice bulb temperature could be easily determined. The estimated error was $\pm 3\ \mu V$. Great care was taken to ensure that the ice layer on the ice bulb was thin and freshly formed, and that the readings were consistent over a period of some minutes.

The following features may be inferred from Table VII:

i. At constant visibility the mean relative humidity (*RH*) increases and the wind speed decreases as the temperature decreases.
ii. At constant temperature, relative humidity and wind speed decrease as the visibility range increases.
iii. The "saturation" wind velocity, i.e. the wind speed when the relative humidity reaches 100 per cent, is lower for low temperatures.
iv. There is a level above the surface where the air is just saturated, and this level rises with increasing wind strength. Above this level evaporation is likely to occur.
v. At low wind speeds and temperatures the relative humidity may exceed 100 per cent, indicating that mixing is taking place between the warmer air at the top of the surface temperature inversion and the colder supersaturated air close to the surface. The effect is greater at low temperatures when the temperature inversions are greatest.
vi. It follows from (v) that some of the so-called "drift" in supersaturated conditions could be in crystals formed by direct sublimation on to nuclei to form an ice fog. Such a fog should exhibit different optical effects to that observed in true drift snow.
vii. No quantitative values for humidity lapse can be inferred, but it is evident that evaporation may take place at some level above the surface in the majority of cases.

It should be noted that the temperatures are at $1 \cdot 25$ m. approximately, while the visibilities refer to eye-level—about $1 \cdot 7$ m. Therefore, there should be some reduction in the values of relative humidity to match the visibilities at $1 \cdot 7$ m.

Mason (1957, p. 186–90) gives an analytical expression for the growth rate of a crystal in a supersaturated environment. It may also be applied to evaporation and thus to find the time required to evaporate a drift particle in conditions indicated in Table VII. Lister (1960, p. 14–17) has shown that drift snow particles varied from $0 \cdot 0004$ to $0 \cdot 04$ mm.^{2} in cross-sectional area with a median size of approximately $0 \cdot 004$ mm.^{2}, and that there is no significant change in particle size for varying drift intensities. The dominant form was that of subrounded polished particles. Assuming a spherical crystal with a cross-sectional area of $0 \cdot 004$ mm.^{2}, which gives a radius of $0 \cdot 036$ mm., evaporation rates may be calculated using Mason's formula

$$dm/dt = 4\pi C\sigma \Big/ \left(\frac{JL_s^2 M}{KRT^2} + \frac{RT}{DM_{e_s}}\right), \qquad (16)$$

where $\sigma = (e/e_s)-1$, the supersaturation,
e = Ambient vapour pressure,
e_s = Saturated vapour pressure over ice,
J = Joule's constant $= 4 \cdot 18 \times 10^7$ erg/cal.,
D = Diffusion coefficient of water vapour in air,
K = Thermal conductivity of air,
C = A constant, dependent on the shape of the crystal. For a sphere $C = r$, the radius.
L_s, M, R, T have the same significance as in equation (8a).

The results of the above calculation are tabulated in Table VIII. The time for evaporation

TABLE VIII. EVAPORATION RATES FOR A SPHERICAL DRIFT PARTICLE 0·07 mm. IN DIAMETER

Temperature (° C)	*Saturation Vapour Pressure* (mb.)	*Thermal Conductivity of Air* (cal./cm./degree/sec.)	*Coefficient of Diffusion of Water Vapour in Air* (cm.²/sec.)	*Relative Humidity* (per cent)			
				98		96	94
				Evaporation Rate (dm/dt) (g./sec.)	*Evaporation Time* (sec.)	*Evaporation Time* (sec.)	*Evaporation Time* (sec.)
−5	4·015	$5 \cdot 71 \times 10^{-5}$	0·218	$3 \cdot 7 \times 10^{-10}$	528	264	176
−15	1·652	$5 \cdot 54 \times 10^{-5}$	0·204	$1 \cdot 93 \times 10^{-10}$	1,010	505	357
−25	0·6323	$5 \cdot 36 \times 10^{-5}$	0·190	$0 \cdot 828 \times 10^{-10}$	2,355	1,178	785
−35	0·2233	$5 \cdot 17 \times 10^{-5}$	0·176	$0 \cdot 307 \times 10^{-10}$	6,350	3,175	2,115

of the specified particle is well within the feasible flight time of the particle. Lister (1960, p. 28) gave a mean free-fall velocity of 40 cm./sec. as compared with the theoretical value of 200 cm./sec. for a sphere. The reduction in fall velocity is the effect of the angularity and non-spherical shape of many particles. Assuming that a particle reaches the top of the drift layer, say 200 m., the time for free fall would be 500 sec. During this time evaporation would be taking place, thus reducing the free-fall velocity still further and prolonging the flight time. In addition, the turbulent air motion would tend to maintain flight until the particle has been completely destroyed.

It has been inferred (Table VII) that evaporation is likely to take place at any height above the "saturation" level. This level does not appear to be too far removed from the surface and probably rarely exceeds 3 or 4 m. in height. This is small compared to the total drift layer of 100–200 m. No doubt, too, the humidity lapse is some function of wind strength and temperature—somewhat less with high wind speeds than with moderate wind speeds, greater with warm temperatures than with cold temperatures. If this is true, then for similar wind strengths not only will evaporation rates be greater at high temperatures but the relative humidities may be less, enhancing evaporation. However, evaporation rates for constant relative humidity still give a relative measure of evaporation at varying temperatures and can be used to estimate the magnitude of drift losses by evaporation.

Boundary conditions must be established first. At 0° C it is reasonable to assume that the total drift loss is by evaporation. At −60° C evaporation is so small as to be negligible. At this temperature it will be assumed that all drift losses are by wastage. Fig. 7 represents the evaporation of a "standard" drift particle for different temperatures and humidities. By assuming 100 per cent evaporation at 0° C, the evaporation rate at any temperature is proportional to the percentage of the drift loss by evaporation,

i.e.
$$\left(\frac{dm}{dt}\right)_T \alpha D_{(e)} \text{ per cent.}$$

Hence,
$$D_{(e)} = \frac{(dm/dt)_{T^\circ C}}{(dm/dt)_{0^\circ C}} \times 100 \text{ per cent.} \qquad (17)$$

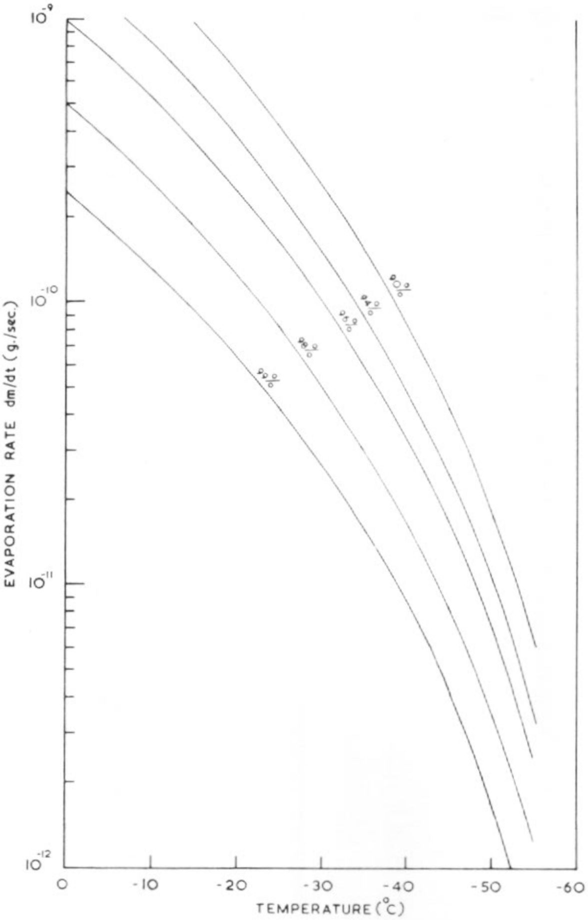

Fig. 7. Evaporation rate of a spherical drift particle of diameter 0·07 mm. for various relative humidities (with respect to ice).

By applying equation (17) to seasonal data from Halley Bay, an estimate of drift loss by evaporation is obtained (Table IX) and this amounts to 38 per cent of the total drift loss for the year. The above discussion is only true when there is no snow falling. When snow

TABLE IX. (ESTIMATED) DRIFT LOSS BY EVAPORATION, $D_{(e)}$, AND BY WASTAGE, $D_{(w)}$

Season	*Number of Cases* (n)	*Mean Air Temperature* (° C)	*Evaporation* $D_{(e)}$ (per cent)	*Wastage* $D_{(w)}$ (per cent)	$nD_{(e)}$
Summer (D, J, F)	114	−4·4	77	13	8,778
Autumn (M, A, M)	169	−15·6	34	66	5,746
Winter (J, J, A)	108	−23·7	18·5	81·5	1,998
Spring (S, O, N)	138	−19·8	26·5	26·5	3,682
Mean Value for Year	$\left(\frac{\Sigma n D_{(e)}}{\Sigma n}\right)$		38	62	

is falling, the air in the majority of cases will be at saturation level and the drift losses will be in the form of wastage. But the accumulation records only take account of net snowfall during the period of falling snow, so that drift losses during snowfall will be completely masked by the net increase recorded. Thus, the evaporation drift losses derived in Table IX from cases of drift with no snowfall will be a valid measure of the loss when the accumulation records indicate that ablation is taking place.

It is probable that at the warmer temperatures the evaporation drift loss is greater than calculated, and less at low temperatures. Bearing this proviso in mind, the drift losses by evaporation are probably not less than 38 per cent of the total recorded drift loss. The remainder is wastage out to sea.

Farther inland all drift losses from a flat surface will be evaporation losses, and for temperature and wind conditions similar to those at the coast the annual loss will be reduced by about 50 per cent.

The Annual Accumulation Budget

The annual budget may now be evaluated.

Gross accumulation (G_0)	=	153 cm. at $\rho_0 = 0\cdot31$	47·5 g. water
Surface settling (S_1)	=	-17 cm.	
Gross accumulation (G_1)	=	136 cm. at $\rho_1 = 0\cdot35$	47·5 g. water
Bulk settling (S_2)	=	$-(16+3)$ cm.	
Gross accumulation (G_2)	=	117 cm. at $\rho_2 = 0\cdot41$	48·0 g. water
Net accumulation (N) (from daily stakes)	=	89 cm. at $\rho_2 = 0\cdot41$	36·5 g. water
Direct surface evaporation (E)	=	10 cm. at $\rho_2 = 0\cdot41$	4·1 g. water
Drift losses (D) Evaporation $D_{(e)}$	=	7 cm. at ρ_2	2·9 g. water
Drift losses (D) Wastage $D_{(w)}$	=	11 cm. at ρ_2	4·5 g. water

The rounding of accumulation figures to the nearest centimetre and the use of the mean value of density introduces an error that amounts to 0·5 g. water difference between G_0, G_1 and G_2 water equivalents. This is less than 2 per cent of the gross accumulation, or about 2 cm. snow. The latter figure is well within the standard deviation of the actual measurements of net accumulation.

The important features of the budget are:

i. The total settling was twice the actual recorded ablation.
ii. Drift losses were twice the losses by direct evaporation from the surface.
iii. Nearly 40 per cent of the drift losses were by evaporation of drift snow in flight. This may be an underestimate.
iv. The total true ablation was about one-quarter of the recorded gross accumulation, and one-third of the net annual accumulation.
v. The estimate of total evaporation ($E+D_{(e)}$) agrees remarkably well with the larger amount derived from energy balance considerations (Table IV and previous section), and adds strength to the argument in favour of evaporation of drift snow.

Item (iv) is considered to be representative of Antarctic ice shelves and may be used to estimate the gross annual accumulation for any ice shelf station whose net annual accumulation is known. For instance, the net annual precipitation at Maudheim was about 37 g. of water (Swithinbank, 1957, fig. 9) and so the gross precipitation would be expected to be of the same order as Halley Bay, i.e. 48 g. water. For an inland station with no drift loss by wastage the net accumulation should be proportionately greater than for a coastal station, though the total amounts of precipitation may be vastly different, mainly on account of the maritime influence on the coastal station.

A consequence of the size of the gross accumulation is that the moisture content of the troposphere in Antarctic regions is probably greater than has been estimated previously, and will make a significant difference in energy balance computations.

Because accumulation readings were either daily or on alternate days, some of the gross precipitation and subsequent ablation would remain unrecorded. Just as the monthly accumulation totals are an underestimate of the actual precipitation, then so may the daily values be an underestimate of the actual gross accumulation. But it is doubtful if such variations could be measured with the number of daily measuring points that were used, because the results will be within the limits of errors imposed by the standard deviation of the surface roughness (σ_s), which is about ± 6 cm. snow (or ± 2 g. water). A more serious objection is that σ_s is also the order of variation occurring at a stake within the space of a few hours in drift snow conditions, and such variations would tend to increase the level of significant change and to exclude the majority of stake readings at, say three-hourly or even twelve-hourly intervals. Thus, with present measuring methods the daily measurements give a close approximation to the actual precipitation, and increased frequency of measurement will not show much, if any, gain in information.

Summary

Two datum levels in the form of "goal posts" set at the half wave-length of the more prominent sastrugi, and at right angles to the main direction of sastrugi may be used to measure daily changes of snow level. The accuracy of any stake measurement varies from $\pm 0 \cdot 1$ to $\pm 2 \cdot 5$ cm., according to the prevailing wind and surface conditions. The standard deviation of the two sets of readings (σ_s) is a measure of surface irregularities and is comparable to that of a larger accumulation stake network.

Daily readings of the "goal posts" and of the surface densities led to an estimate of surface settling (S_1) of 17 cm./yr., and a comparison of pit densities with surface densities revealed a bulk settling of 19 cm./yr. Rough estimates of total settling ($S_1 + S_2$) were also obtained by scrutiny of daily accumulation and ablation rates in conjunction with meteorological records.

Losses by direct evaporation are estimated to be as much as 4 g. of water, whilst drift losses by wastage, i.e. blown out to sea, and the evaporation of drift snow in flight were estimated to be 20 per cent of the net accumulation of 37–38 g. water. The evaporation of drift in flight amounted to not less than 38 per cent of the total drift-snow loss. In flat inland areas it is suggested that all drift losses are by evaporation.

The gross accumulation at Halley Bay for 1959 was 48 g. water, and a similar amount is suggested for Maudheim, which has the same net annual accumulation. The budget is considered to be representative of the ice shelf at Halley Bay except within 0·25 miles (0·4 km.) of the ice front.

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