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On the Mass and Heat Budget of Arctic Sea Ice[1]

By

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With 12 Figures

Summary. Measurements during the drift of "US Drifting Station A" show an annual mass increase of old ice consisting of 12.5 g/cm^2 snow and 52 g/cm^2 bottom accretion. During the summer seasons 1957 and 1958 an amount of 19.2 and 41.4 g/cm^2 respectively, was lost by surface ablation. The ratio of ablation on elevated "dry" surface and in meltwater ponds is 1:2.5. The average pond area was about 30%. Bottom ablation by heat transfer from the ocean was found to be 22 cm (July to Aug./Sept.).

Methods of measuring mass changes are described. In view of their importance as a means of checking the computed heat budget their accuracy is discussed in detail.

The heat budget is computed for a selected period during the height of the melt season. The average daily totals are, in cal/cm^2: + 142 from net short wave radiation, — 8 from net long wave radiation, + 9 from turbulent heat transfer, and — 11 from evaporation. The mean daily surface ablation is 0.8 cm. About 90% of it is due to the absorption of short wave radiation.

Only 62% of the total heat supply are transformed at the surface. 38% are transmitted into the ice and mainly used to increase the brine volume. The vertical distribution of this energy was used to compute the extinction coefficient for short wave radiation. From 40 to 150 cm depth it is 0.015 cm^{-1}, somewhat smaller than that of glacier ice.

The heat used during the summer to increase the brine volume in the ice acts as a reserve of latent heat during the cooling season. By the time an ice sheet of 300 cm thickness reaches its minimum temperature in March, 3000 cal/cm^2 have been removed to freeze the brine in the interior of the ice and the meltwater ponds, and 1700 cal/cm^2 to lower the ice temperature. Based upon the observed mass and temperature changes the total heat exchange at the upper and lower boundary is estimated. During the period

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May–August the upper boundary received 8.3 kcal/cm^2, while during the period September–April 12.8 kcal/cm^2 were given off to the atmosphere. The results are compared with those of YAKOVLEV, and considerable disagreement is found with respect to the amounts of heat involved in evaporation and in changes of ice temperature ("heat reserve").

Zusammenfassung. Beobachtungen während der Drift von "US Drifting Station A" zeigen an altem Eis einen jährlichen Massenzuwachs von 12,5 g/cm^2 Schnee und 52 g/cm^2 Eis an der Unterseite. Während der Schmelzperioden 1957 und 1958 betrug der Massenverlust an der Oberseite 19,2 bzw. 41,4 g/cm^2. Das Verhältnis der Ablation auf „trockenen" Eisflächen zu der in Wassertümpeln beträgt etwa 1 : 2,5. Etwa 30% der Gesamtfläche werden im Sommer von den Wassertümpeln eingenommen. Die Ablation an der Unterseite durch Wärmezufuhr vom Meer betrug etwa 22 cm (Juli bis August/September).

Die Methoden der Messung des Massenhaushalts werden beschrieben. In Anbetracht ihrer Bedeutung als Kontrolle des berechneten Wärmehaushalts wird ihre Genauigkeit näher untersucht.

Die Wärmebilanz der Eisoberfläche wird für einen ausgewählten Zeitraum während des Maximums der Ablationsperiode berechnet. Es ergeben sich folgende mittlere Tagessummen in cal/cm^2: + 142 kurzwellige Strahlungsbilanz, — 8 langwellige Strahlungsbilanz, + 9 Konvektionswärmestrom. — 11 Verdunstung. Die mittlere tägliche Oberflächen-Ablation betrug in dieser Zeit 0,8 cm. Etwa 90% davon werden durch Absorption kurzwelliger Strahlung verursacht.

Nur 62% des gesamten Wärmeangebotes werden an der Oberfläche umgesetzt. 38% gelangen in tiefere Schichten und werden dort hauptsächlich zur Vergrößerung des Volumens der Salzlösung verwendet. Die vertikale Verteilung dieser Energie wird zur Berechnung des Extinktionskoeffizienten für kurzwellige Strahlung herangezogen. In einer Tiefe von 40 bis 150 cm ergibt sich ein Wert von 0,015 cm^{-1}, etwas weniger als in Gletschereis.

Die Wärmemenge, welche im Sommer zur Erhöhung der Eistemperatur und der damit verbundenen Vergrößerung des Volumens der Salzlösung aufgewendet wurde, dient während der Abkühlungsperiode als Wärmereserve. Von ihrem Beginn bis zur Erreichung minimaler Eistemperaturen in März werden einer 3 m dicken Eisdecke 3000 cal/cm^2 an latenter Wärme (Verkleinerung des Volumens der Salzlösung und Gefrieren der Schmelzwassertümpel) und 1700 cal/cm^2 mit der reinen Temperaturerniedrigung entzogen. Auf Grund der beobachteten Massen- und Temperaturänderungen der Eisdecke wird der gesamte Wärmeumsatz an ihren Grenzflächen abgeschätzt. Während der Periode Mai bis August erhält die Oberfläche des Eises 8,3 kcal/cm^2 während in der Periode September bis April 12,8 kcal/cm^2 an die Atmosphäre abgegeben werden. Die Resultate werden mit denen von YAKOVLEV verglichen, wobei sich beträchtliche Unterschiede in den Beträgen der Verdunstung und der Wärmereserve der Eisdecke ergeben. Im Zusammenhang mit den unterschiedlichen Beträgen der Wärmereserve wird die spezifische Wärme des Meereises näher diskutiert.

Résumé. Les observations faites lors de la dérive du "US drifting station A" font apparaître un accroissement annuel de masse de la banquise de 12,5 g/cm^2 sous forme de neige et de 52 g/cm^2 par congélation à la base. Pendant les périodes de fonte de 1957 et de 1958, la perte de masse à la surface fut de 19,2 et 41,4 g/cm² respectivement. Le rapport de l'ablation sur la glace "sèche" à celle des flaques est de 1 : 2,5. Les flaques occupent en été env. 30% de la surface totale. L'ablation à la face inférieure de la banquise par la chaleur de l'eau fut d'environ 22 cm (Juillet à août/septembre).

On décrit les méthodes de mesure du bilan de masse et leur précision. On calcule ce bilan de chaleur à la surface au moment du maximum d'ablation et on en donne les composantes suivantes pour les sommes journalières moyennes en cal/cm²: bilan radiatif de courte longueur d'onde + 142, bilan radiatif de grande longueur d'onde — 8, flux de convection + 9, évaporation — 11. L'ablation superficielle moyenne est de 0,8 cm par jour dont 90% résulte de l'absorption du rayonnement à courte longueur d'onde.

Le 62% seulement de l'apport de chaleur est transformé à la surface de la glace ; le 38% pénètre en profondeur et sert surtout à accroître le volume du mélange salin. A une profondeur de 40 à 150 cm le coefficient d'exctinction pour le rayonnement court est de 0,015 cm^{-1}, plus faible que dans le glacier terrestre.

La quantité de chaleur accumulée en été sert de réserve pendant la période froide pour élever la température de la glace et pour augmenter le volume de la solution saline. Du début de celle-ci jusqu'au minimum des températures en mars, une couche de 3 m d'épaisseur perd 3000 cal/cm² en chaleur latente et 1700 cal/cm² par chute de température. Il est possible d'estimer le bilan total de chaleur des surfaces de la glace à l'aide des variations observées de masse et de température. Pedant la période de mai à août, la surface de la glace reçoit 8,3 kcal/cm², tandis qu'elle cède à l'air 12,8 kcal/cm² de septembre à avril. Les résultats obtenus diffèrent de ceux de Yakovlev dans les quantités de l'évaporation et de la réserve calorifique de la glace. Discussion au sujet de la chaleur spécifique de la glace de banquise.

Introduction

By far the greatest ice-covered area on the northern hemisphere is the Arctic pack ice. Compared to the Greenland ice cap, its mass is small, however, its great influence on the heat exchange between earth and atmosphere and its high sensitivity to climatic fluctuations make it an important factor in the distribution of heat over the hemisphere. The conditions necessary for the formation and preservation of sea ice are of equal interest to meteorology and oceanography.

The present report contains part of the results of heat and mass budget studies of sea ice carried out on "US Drifting Station A" during the International Geophysical Year 1957/58. A brief description of the expedition is given in [*34*]. Remarks on the methods of investigation as well as some preliminary results have been published in [*33*].

The camp built on a large ice floe in April, 1957, was damaged by ice cracks and pressure ridges in May, 1958, and had to be transported to another floe where the observations were continued until the camp had to be finally abandoned on 6 November, 1958. In the following, the two periods are called "Station 1" and "Station 2."

1. Mass Budget

The total mass of ice present on the Arctic Ocean is in general controlled by two processes: a) by large-scale horizontal drift, and b) by local melting and freezing. In some parts of the Arctic Ocean the horizontal drift is rapid (e. g. SSSR NP-I, May, 1937: 90° N, February, 1938: 70° N), and the ice may be transported to climatologically different regions within one year. Ablation and accretion of the ice take place at two surfaces. Considerations of the horizontal transport are complicated by the fact that frequently the condition of continuity is not valid. For example, in the case a flow divergence, large amounts of new ice may be formed where the ocean surface becomes exposed to the atmosphere by the parting of a continuous ice cover.

The general pattern of ice drift in the Arctic ocean is known primarily by the drifts of the ships "Fram," "Maud," and "Sedov," as well as by the Soviet and American drifting Stations (see [*9*, *15*]). The tangential forces exerted on the ice by air and water are also basically known [*9*, *15*]. The physical relationships found between air and ice movement represent average values with good accuracy but large deviations are observed in individual cases.

Their cause may be largely the fact that the pack ice cover can accumulate tensions so that the movement at one place is not only a function of the locally applied forces but influenced by forces transmitted over longer distances. The irregular distribution of such forces due to the irregular shape of the ice floes make the mechanical properties of large sheets of sea ice different from those of small samples. Investigations of the deformation of large fields of sea ice in relation to wind fields are highly desirable for an understanding of the ice budget in the Arctic ocean.

The following considerations of the mass budget refer to an individual ice floe. A simplified plot of its drift during the existence of "Station A" is given in Fig. 1. The total distance covered from May, 1957, to November, 1958, was 3200 km.

1. 1. Methods of Observation

Ablation and accumulation on ice surface are commonly observed by means of stakes. The upper surface of sea ice is in most cases irregular (hummocks, pressure ridges, meltwater ponds, snow-drifts, sastrugi), and the mass changes by ablation and accumulation are comparatively small. Therefore, accurate measurements can only be obtained when a sufficiently great number of stakes is used. At Station A, 15 stakes were used from July to September, 1957, 20 stakes from October, 1957, to May, 1958, and 30 stakes from May to October, 1958. During the two latter periods they were placed in two perpendicular lines, at equal distances.

When observing the mass changes of sea ice it is to be noted that not all the meltwater formed at the surface runs off. A certain part remains on the ice in ponds and re-freezes in the fall. During summer, ablation takes place both at the upper and lower surface.

After some experiments, the measuring arrangement shown in Fig. 2 was finally used. In addition to the stakes ($Z_1 - Z_3$), a water level recorder (tide gauge) was installed. It permits the observation of the hydrostatic equilibrium of the ice floe. The recordings of this instrument are particularly valuable since they are representative of a relatively large area. The support carrying the recording drum is anchored at a level L which remains fixed relative to the main body of the ice. The

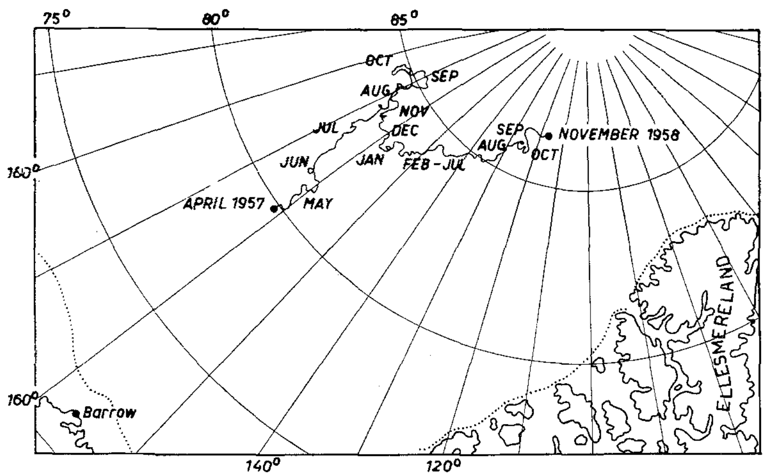

Fig. 1. Drifting Station A during the International Geophysical Year 1957/58

vertical movements of an ice plate (density ρ_i) floating in water (density ρ_w) are recorded as changes of l. Ablation or accretion at the upper or lower surface correspond to changes of l_1 and l_2. Assuming constant density, the state of equilibrium is given by

$$\rho_i (l_1 + l_2) = \rho_w (l_2 + l).$$

In the case of ice and sea water, mass changes at the upper and lower surface are recorded with a sensitivity ratio of about 7 : 1,

$$\Delta l = \frac{\rho_i}{\rho_w} \Delta l_1 - \frac{\rho_w - \rho_i}{\rho_w} \Delta l_2. \tag{1}$$

In the recordings of Δl, the contributions of Δl_1 and Δl_2 cannot be distinguished. Therefore, separate measurements of bottom ablation and accretion are necessary. In order to avoid the tedious and inaccurate procedure of measuring ice thickness through bore holes another device has been used (see Fig. 2). The "measuring wire" which is normally frozen into the ice, is released by electrical heating and pulled up so that the cross bar at its lower end rests against the ice bottom. The upper end of the wire is read against a fixed scale. Since the lower ice surface of old ice floes is comparatively smooth, a number of 3 to 4 of such measuring points was considered sufficient within the accuracy of all other observations.

In the water level recordings, two sources of error should be noted:

a) *Ice density.* Old sea ice has an average salinity of about $2^0/_{00}$. The size of the brine pockets is variable with temperature and leads to changes of the average density. Between — 1 and — 10° C the latter is of the order of 1%. According to eq. (1), the corresponding variation of l is of the same magnitude. In order to keep its absolute magnitude small it is advisable,

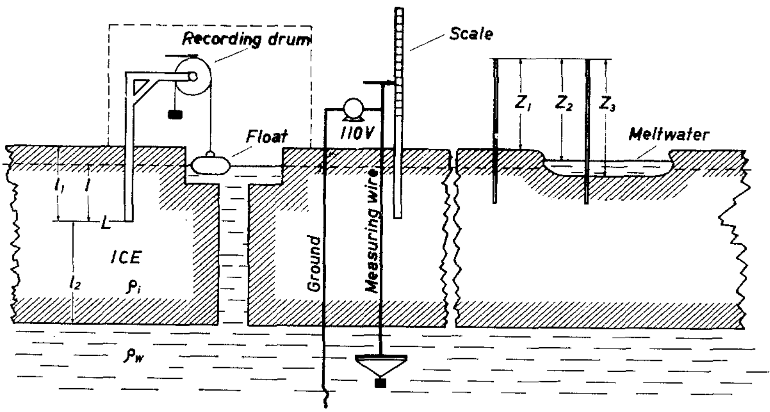

Fig. 2. Means of measuring mass changes of the ice cover

to make l also small, i. e. to anchor the support just deep enough to avoid difficulties with surface melting. Strictly speaking, it would be necessary to allow for the vertical distribution of salinity as well. Furthermore, the air content near the surface may not be constant during the ablation season. Pertinent observations are not available. However, for the present purpose it seems permissible to neglect changes of ice density.

The average ice density as measured by weighing cores of known volume was found to be 0.89 g/cm³ [*21*]. In April 1958, a crack through the scientific area of Station A afforded a welcome possibility to take a profile of snow height, free-board, and ice thickness. The results are given in Table 1. With the densities of snow ($\rho_s = 0.33$ g/cm³) and water ($\rho_w = 1.03$ g/cm³) the average density of the ice is found to be 0.91 g/cm³. For all further computations, a mean value of 0.90 g/cm³ has been used.

b) *Fresh water.* Most of the fresh water formed during the melt season runs off over the edges of the floes. If the relative area of open leads is not too great and if mixing by wind not too intensive, the layer of fresh water floating on the denser sea water may become thicker than the floes and expand along the ice bottom. In this case, the ice floats partially or entirely in fresh or low salinity water. With an ice thickness of 300 cm, a maximum hydrostatic adjustment of 9 cm may result. The present data do not allow an evaluation of this effect. Especially during the summer of 1958 there were large areas of open water in the vicinity of the station in which the thickness of the fresh water layer was only a fraction of the ice thickness [13]. On the other hand, it was necessary for reasons of camp and runway maintainance to drill a large number of drainage holes. The fresh water spreading at the underside of the ice leads to the formation of a sub-layer of fresh ice, as described in [33]. It seems, therefore, not possible to correct the water level recordings for density variations of the water. The estimated maximum

Table 1. *Mean Ice Density, April 1958, under the Assumption of* $\rho_s = 0.33$ *and* $\rho_w = 1.03$ *g/cm*³

Measuring points, distance ~5 m	Snow	Ice above water	Ice under water
1	42 cm	22 cm	315 cm
2	41	20	282
3	29	31	288
4	43	27	304
5	30	19	270
6	37	22	274
7	42	25	277
8	53	13	259
9	90	8	270
10	46	18	296
11	29	18	274
12	33	22	291
13	66	17	299
14	19	44	355
15	35	23	303
16	32	25	277
17	31	26	279
18	26	24	315
19	35	17	289
20	79	19	292
21	59	14	296
mean	42.7	22.1	291

$\overline{\rho_i} = 0{,}913$ g/cm³

error is 3 cm. The hydrostatic rise of the ice floe during the ablation season may be too small by this amount. (In maintaining the water level recorder it is particularly important to keep the water in the vertical tube communicating with the ocean either always salty, or always fresh. Here, the error may amount to 9 cm.)

By means of the measuring devices shown in Fig. 2 it is possible to determine the mass budget with reasonable accuracy. If A_T (cm) and A_W (cm) denote the mass losses on the "dry" surface and in ponds, respectively, A_U the loss at the bottom, and F the relative area of ponds, then the total loss of mass, ΔM, (in cm of water) during a certain time interval can be written

$$\Delta M = \rho_i A_T (1 - F) + \rho_i A_w F + \rho_i A_U - F \cdot \Delta d \qquad (2)$$

Table 2. *Examples of Ablation Measurements on Sea and Glacier Ice. By Means of their Standard Deviation and the Graph Shown in Fig. 4 the Number of Observations Necessary to Obtain an Arithmetic Mean of Given Accuracy Can Be Determined*

Station "A", 83°N, 165°W — Ablation on elevated surfaces, stakes

1958		A	N	σ	$\frac{\sigma}{\sqrt{N}}$	%
July	6/7	3.1	15	1.10	0.28	9
	7/8	0.8	15	0.83	0.21	26
	8/10	2.4	15	1.02	0.26	11
	10/11	1.2	15	1.08	0.28	23
	12/13	1.4	15	0.70	0.18	13
	13/14	0.9	15	0.62	0.16	18
	14/15	0.2	15	1.00	0.26	130
	15/17	1.7	15	0.85	0.22	13
	17/19	1.4	15	0.66	0.17	12
	22/23	1.0	15	0.63	0.16	16
	23/24	3.4	15	0.80	0.21	6
	24/25	0.6	15	0.45	0.12	20
	25/26	0.4	15	0.54	0.14	35
	26/27	1.0	15	0.94	0.24	24
	27/28	0.4	15	0.86	0.22	55
	28/29	2.5	15	0.77	0.20	8
	29/30	2.1	15	1.01	0.30	14
	30/31	1.1	15	1.22	0.32	29
Aug.	31/1	0.4	15	1.17	0.30	75
	2/3	0.2	15	0.93	0.24	120
	3/4	0.1	15	1.02	0.26	260
	4/5	0.6	15	1.12	0.29	48
	5/6	1.0	15	0.80	0.21	21
	6/7	0.2	15	0.87	0.22	110

Station "A", 83°N, 165°W — Ablation in ponds, stakes

		A	N	σ	$\frac{\sigma}{\sqrt{N}}$	%
July	6/7	2.9	15	0.87	0.22	8
	7/8	2.2	15	1.62	0.31	14
	8/9	2.2	10	1.80	0.57	24
	9/10	1.4	9	0.72	0.24	17
	10/11	2.4	13	2.66	0.74	31
	11/12	2.0	13	2.52	0.70	35
	12/13	2.0	13	1.75	0.48	24
	13/14	2.3	13	2.11	0.58	25
	14/15	2.4	13	0.83	0.23	10
	15/16	2.1	8	0.95	0.34	16
	16/17	1.9	8	1.48	0.52	27
	17/18	2.0	9	0.68	0.23	12
	18/19	1.8	9	1.20	0.40	22
	22/23	1.0	13	1.28	0.36	36
	23/24	3.0	14	2.39	0.64	21
	24/25	2.6	14	2.19	0.59	23
	25/26	2.0	14	1.99	0.53	26
	26/27	2.3	14	1.29	0.34	15
	27/28	1.3	14	1.36	0.36	28
	28/29	2.7	14	1.51	0.40	15
	29/30	2.9	14	2.00	0.54	19
	30/31	1.2	14	0.88	0.24	20
Aug.	31/1	3.4	14	1.64	0.44	13
	1/2	2.2	14	1.94	0.52	24

Chogo-Lungma-Gl., Karakorum, 4300 m — Ice ablation, bore holes

		A	N	σ	$\frac{\sigma}{\sqrt{N}}$	%
July	15	7.9	7	2.5	0.95	12
	16	4,7	7	1.4	0.53	11
	17	6,4	7	1.0	0.38	6
	19	4.8	7	0.9	0.34	7
	23	5.3	7	2.0	0.76	14
	25	6.4	10	1.4	0.44	7
	26	5.8	10	0.8	0.25	4
	27	4.8	10	0.4	0.13	3
	28	7.3	10	0.8	0.25	3
	29	7.8	10	0.8	0.25	3
	30	7.7	10	1.4	0.44	6
	31	8.2	10	1.8	0.57	7
Aug.	1	7.5	10	1.1	0.35	5
	2	7.5	10	1.4	0.44	6

A = Ablation in cm, arith. mean
N = Number of stakes or bore holes
σ = Standard deviation
$\sigma/\sqrt{N}$ = Mean error of the arith. mean

$$\% = 100\,\frac{\sigma/\sqrt{N}}{A}$$

where Δd is the increase of pond depth. This expression holds only for a sufficiently short period in which $F = \text{const}$. Since A_T, A_W, A_U, ΔM, and d are given by the observations (Fig. 2), eq. (2) could be used to determine F whose direct measurement is laborious. Experience showed that, during the first part of the ablation season the pond area increased rapidly, and since the necessary observations are not available it is impossible to evaluate eq. (2) under such circumstances.

1. 2. Accuracy

The observations of accumulation and ablation are an important and frequently applied means of checking observations of the heat budget. A closer examination of their accuracy seems of interest.

Apart from various experiments with ablatographs (e. g. [*12*]) and the occasional application of photogrammetry for special purposes [*38*] stake measurements are most commonly used. On temperate glacier ice, the measurement of the depth of bore holes has some advantages. In both cases the distance from a fixed level to the average ice level is measured. Therefore, the scattering of individual readings should be about equal. A selection of such observations from a subtropical glacier (Chogo-Lungma-glacier [*32*]) and from sea ice is given in Table 2. It is seen that, on days with little ablation (e. g. Station A, 14/15 July or 2/3 and 3/4 August), the average error of the arithmetic mean can become greater than the mean itself. An observation of such low accuracy is hardly usable to check other observations.

Bochow, Höhne and Raeuber [*8*] have given a handy graph to determine the number of observations necessary to obtain a representative mean. With an additional curve for 90% probability, it is reproduced in Fig. 3. First, the standard deviation of a series of observations, σ, is computed. The arithmetic mean is considered to be representative if it differs, with a predetermined probability P, by not more than x from the true mean. x and P are fixed according to any particular requirements. N is the number of observations. If the point $(x/\sigma, N)$ falls to the right of the chosen P-curve then the arithmetic mean is representative. If it falls to the left, the number of observations necessary for an arithmetic mean of the desired accuracy is established by finding this point of the respective p-curve whose ordinate is x/σ. The abscissa in this point is the required N.

According to Table 2 the standard deviation of ablation measurements on glacier and sea ice is approximately 1.0 cm. It seems reasonable to postulate that the computed mean should be within $\pm$ 10% of the true mean, with a probability of at least 90%. If the required accuracy of the ablation measurement is $\pm$ 0.1 cm, this would require 340 measuring points according to Fig. 3. On the other hand, a given number of observations, $N = 15$, renders an error $x' = 0.45$. According to the requirement stated above, this should be not more than 10% of the true mean. In other words, by using 15 stakes, ablation can only be measured with sufficient accuracy if it is not less than 4.5 cm. The daily amounts of ablation at Station A (Table 2) are all much below this magnitude and of a correspondingly little accuracy. When using ablation data to check observations of the heat budget, longer periods with sufficiently great amounts of ablation will have to be considered. On the other hand, the daily amounts of ablation on Chogo-Lungma glacier (Table 2) were large enough to render representative means with only 7 to 10 single values ($\varkappa'/\sigma = 0.7$).

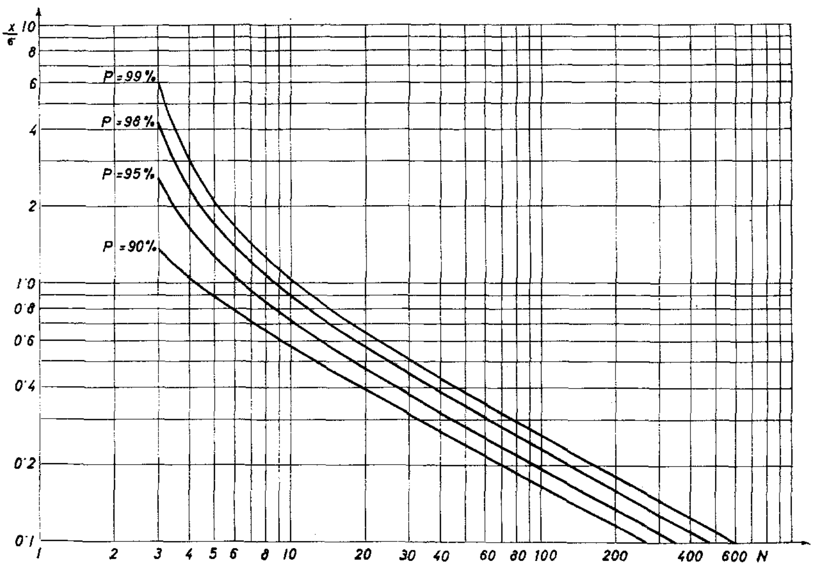

Fig. 3. Graph for the determination of the number of single values (N) necessary to obtain a representative arithmetic mean (after [8])

Much more unfavourable are such considerations for the measurement of accumulation. Due to the surficial irregularities and the transport of snow by wind (formation of sastrugi) the standard deviation becomes very large. For the measurement of accumulation the water level recordings are particularly valuable.

1.3. Results of Observations

Detailed observations of the mass budget began at Station 1, on 10 July, 1957 (Fig. 4). At this date, some ice ablation had already taken place. A few measurements of snow height and density had been made before the onset of melting (early June, first part of curves a and b in Fig. 4). The dashed parts of curves a and b have been extrapolated from our notes on the state of the surface. The data obtained at Station 2 permit a more detailed representation (Fig. 5). While the stake and

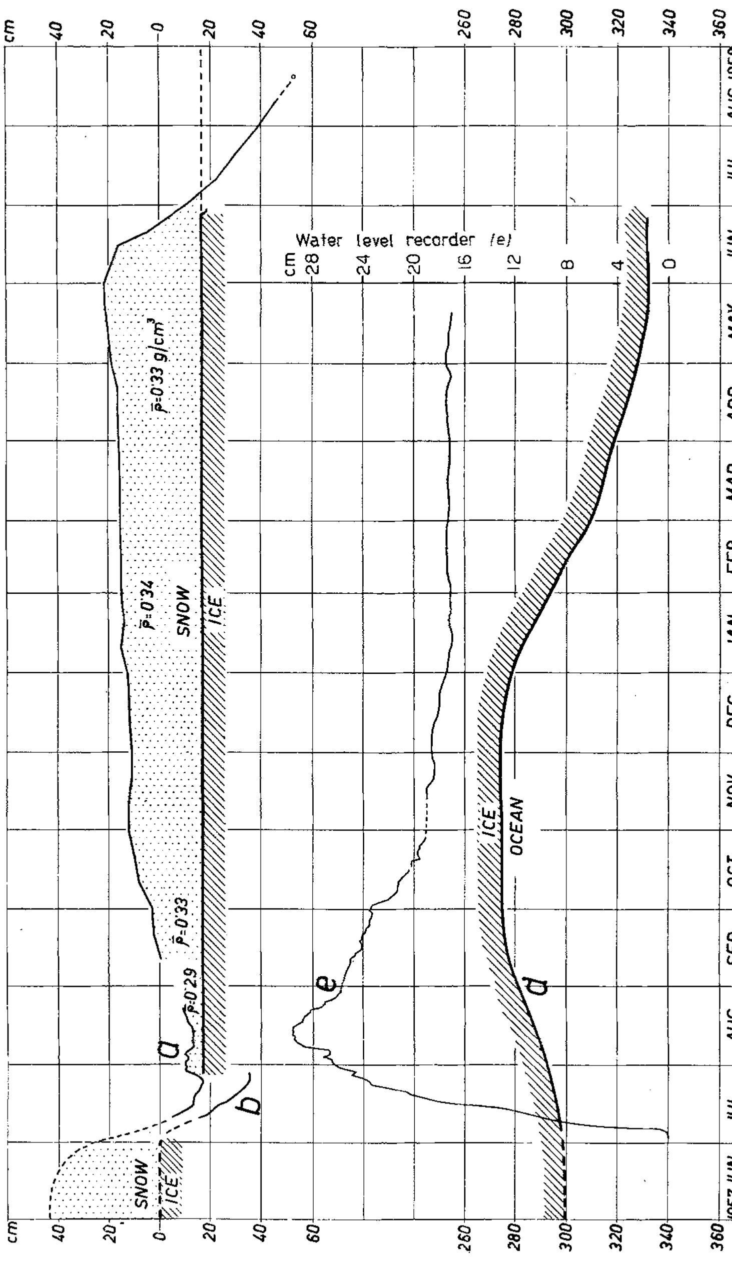

Fig. 4. Mass changes at the upper and lower surface of the ice, station 1. Curve *e* shows the hydrostatic adjustment of the floe

thickness readings could be continued for some time after the evacuation on Station 1, the water level recorder was taken down on 19 May and re-installed at Station 2 on 27 May, 1958.

1.3.1. Ablation

By comparison of Fig. 4 and 5 it is seen that the ablation periods 1957 and 1958 began at approximately the same time, however, in 1958

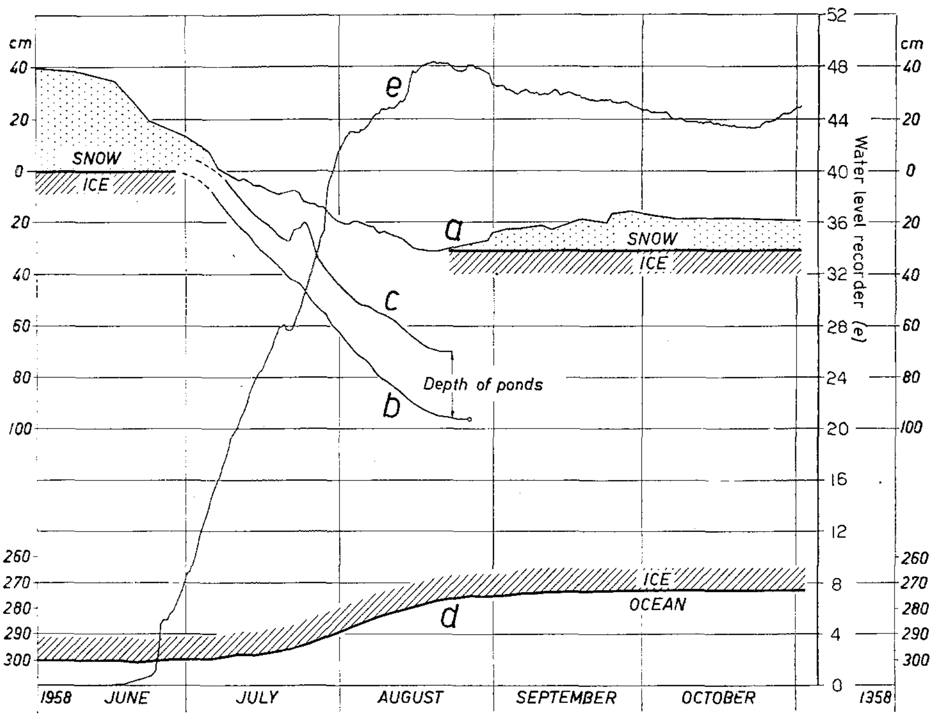

Fig. 5. Mass changes at the upper and lower surface of the ice (see also Fig. 2 and 4)

it lasted about one month longer and led to correspondingly greater ablation. The mean geographical position of the Station in July, 1957, was 82°N, 165° W, and in July, 1958, 84°N, 145° W (distance ~ 300 km.).

Eq. (1) can be used to ascertain the total loss of mass at the surface during the period of ablation. The values of Δl_1 and Δl_2 are taken from curves e and d in Fig. 4 and 5. For 1957, $\Delta l_1 = 32.2$ cm, and for 1958, $\Delta l_1 = 53.1$ cm (all values in water equivalent). According to the observations of snow height and density, 13.0 and 12.1, respectively, of the above totals come from the melting of snow. The remaining 18.0 and 39.9, respectively, represent the loss of ice at the surface. It is composed of the ablation on "dry" surfaces and in ponds, reduced by the pond volume at the end of the ablation period. The latter is an important factor in the heat budget. The latent heat stored in the pond water retards cooling in the fall. As mentioned earlier, observations of the pond area are not available. The observations of pond depths, too, were only in 1958 sufficiently numerous. In this case, eq. (2) may serve to estimate the pond area F. The values of A_T, A_W, and d are taken from Fig. 5 ($\rho_i = 0.90$ g/cm^3). A pond area of $F = 42\%$ is obtained. If the ponded meltwater were spread out it would form a layer of water 10.5 cm thick. The latent heat stored in this layer is 840 cal/cm^2. If one assumes a similar pond area for 1957, the heat reserve would be 200 cal/cm^2. Due to the shorter ablation season in 1957, the ponds were not as deep as in 1958. Because of the lower reliability of the 1957 observations too much emphasis should not be placed on the above result.

The amount of solid precipitation during June–August is generally small. A relatively strong snow-fall on 20 July, 1958, is clearly shown by the water level recording (Fig. 5, curve e). A small depression on 21 July is followed by an especially steep rise on 28 July which represents the run-off of the excess melt-water. A large part of the freshly fallen snow was blown into the ponds, as shown by the temporary increase of their depth (curve c).

As it will be explained later, the measurement of ablation of sea ice by means of stakes is of limited value. A large proportion of the total turnover of heat takes place below the surface and does not manifest itself as surficial ablation. The higher the temperature and salinity of the ice, the greater becomes internal melting and the more important it becomes to ascertain the true loss of mass (run-off) by a cumulative method as provided by the water level recorder.

1.3.2. Accumulation

The present observations provide further evidence that the annual amount of precipitation in the central Arctic is small, of the order of 150 mm. It takes only a short time for the freshly deposited snow to attain an approximately constant density of 0.33 g/cm^3. Similar results were obtained by the Soviet drifting Station NP-II [*36*]. Some of the average densities observed have been entered in Fig. 4. As mentioned earlier, the water level recordings can be used for the determination of a cumulative snow density. For the time from the beginning depression of the ice floe (Fig. 4, maximum of curve e) until the beginning of ice accretion at the bottom, eq. (1) renders $l_1 = 9{,}5$ cm of water column or 29 cm of snow with a density of 0.33 g/cm^3, in excellent agreement with the direct observations (curve a). However, for the time of accretion at the bottom, December–April, eq. (1) renders incorrect values. To a small extent this may be caused by a greater density of the freshly accreted bottom ice but it is easily seen by means of eq. (1) that even an extremely high density of the new ice cannot explain the floe's lack of lift. The actual cause is most likely to be sought in the fact that the 80 cm high box covering the water level recorder collected large snow drifts in all directions with a length of up to 20 metres. During clear weather and falling temperatures the surface of recently deposited snow drifts frequently become hardened by sublimation within the uppermost layer so that the drift is not easily removed by wind from another direction. It is, therefore, inevitable that any aerodynamical obstacle collects more or less permanent snow-drifts on all sides which cause an additional depression of the ice. Similar conditions prevailed during the deposition of snow in the fall, 1958 (Fig. 5). At the end of the melting season, when the

Table 3. *Observed Amounts of Mass Loss and Gain at the Upper and Lower Surface of the Ice. The Meltwater Retained in Ponds has been Deducted from the Loss Figures. A_T and A_W are the Amounts of Ablation on Elevated ("Dry") Surfaces and in Ponds, Respectively*

Boundary	Gain	Loss	Ice ablation	
			A_T	A_W
air/ice (cm water)	1956/57 : 13,0	1957 : 32,2 (13,0 snow; 19,2 ice)	15,5	31,0
	1957/58 : 12,1	1958 : 53,5 (12,1 snow; 41,4 ice)	27,6	85,3
	1958 : (5,2) (until Oct.)			
ice/ocean (cm ice)	1957/58 : 57	1957 : 22 1958 : 24		

ponds begin to freeze their level is about 20 to 50 cm below the average "dry" level. Most of the new snow is blown into these depressions and not accurately represented by the stake readings (curve *a*). In this case, the water level recordings are more reliable. The amount of accumulation from the middle of August to the end of October, 1958, is found to be 5.2 cm of water or 16 cm of snow with a density of 0.33 g/cm^3. When the ice-cover of the ponds is still thin it is frequently broken by load of the overlying snow. The filling of the ponds with water-saturated snow accelerates the process of freezing. In this manner, atmospheric precipitation which otherwise takes part only indirectly in the mass budget, can contribute a small amount to the formation of ice at the surface.

1.4. Summary

The constituents of the mass budget are summarized in Table 3. Ice accretion during the winter 1957/58 was greater by 16 cm than the ice loss during the previous summer. This was compensated by a relatively great loss during the following summer. Assuming an ice thickness of 300 cm in May, 1957, it was 316 cm in May, 1958, while the September thicknesses decreased from 259 cm in 1957 to 251 cm in 1958. The findings given in Table 3 provide further evidence that the average annual accretion of old ice with a thickness of 3 m is close to 50 cm.

From observations of the thickness of annual layers SCHWARZACHER [*29*] has derived a numerical relation between September ice thickness and annual accretion (Fig. 6, dotted line). Between thicknesses of 150 and 350 cm it is based on observations but its extrapolation to zero thickness (in September) renders a very high value of 340 cm for accretion

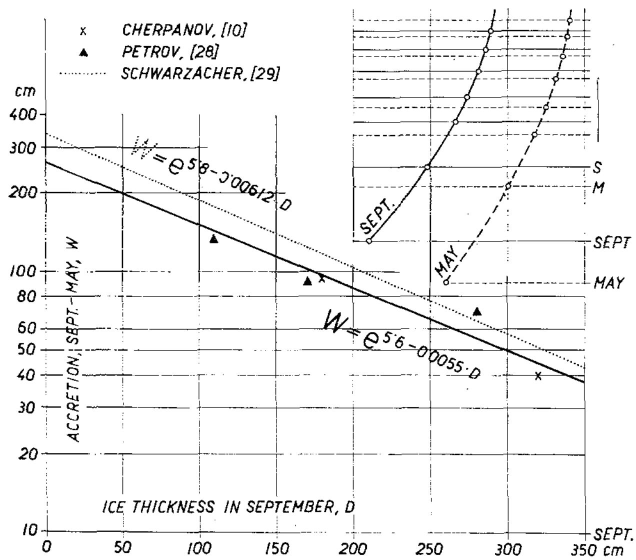

Fig. 6. Relation between ice thickness in September, D, and winter accretion, W (straight lines). Assuming a constant ablation of 50 cm/year and $W = e^{5.6-0.0055D}$, the ice thickness in September and May are indicated by open circles (log time scale on the right). The curved lines connecting them represent the margins of annual variations of thickness

during the first winter. By allowing for the measurements of PETROV [*28*] and CHERPANOV [*10*] it seems more favourable to chose somewhat smaller constants (full line). Accretion during the first winter is still 260 m, however, this amount is in better agreement with PETROV's observations of the increase of thickness with time. By using this relation and assuming a constant ablation of 50 cm (which seems justified for ice thicknesses $>$ 200 cm) the increase of thickness in May and September, beginning at $D = 0$, is also shown in Fig. 6.

Under the assumption that the annual layers are identical with winter accretion, SCHWARZACHER [*29*] uses the above relationship to reconstruct ice thickness and amount of ablation from 1957 back to 1951. The results are interesting and show great variations of ablation which SCHWARZACHER ascribes to a possible variation with ice thickness and to the effect of the variable snow cover on accretion. It seems, however, that even greater errors result from neglecting bottom ablation during

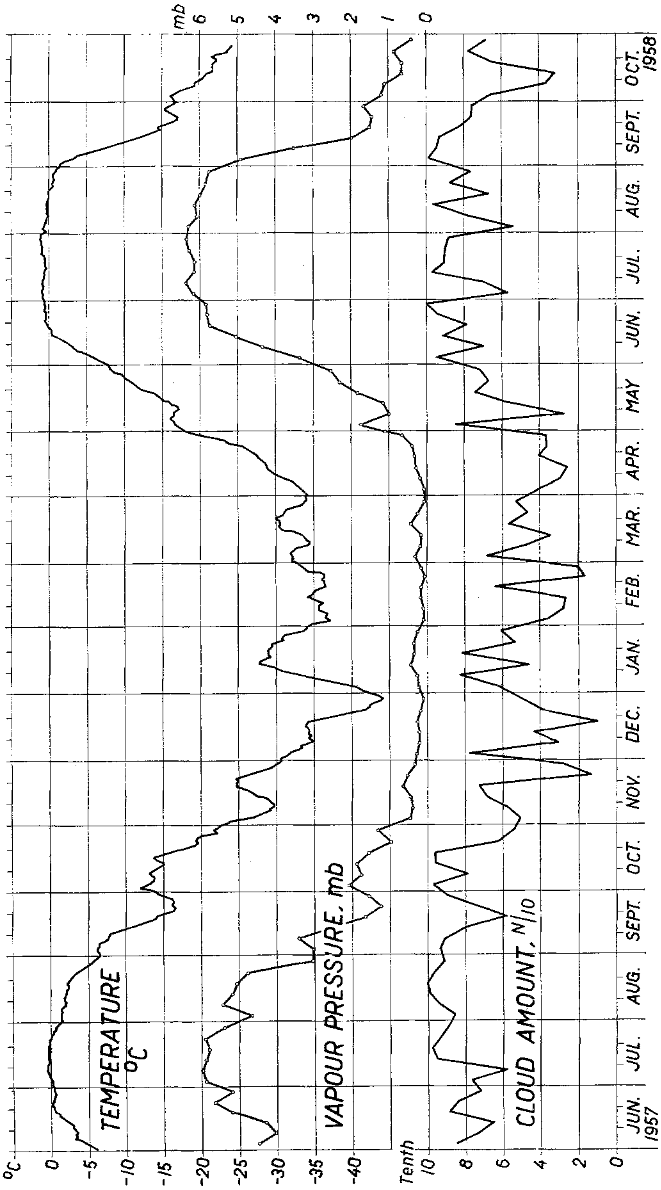

Fig. 7. Temperature (11-day running means), vapour pressure (5-day means), and cloudiness (5-day means) at Drifting Station A, according to observations by the US Weather Bureau

summer. Its amount would have to be added to the annual layers in order to obtain the true amount of winter accretion.

2. Climate

Fig. 7 and Table 4 show the variations of some climatological elements during the time of observation at Drifting Station A. During the melting period the 11-day running means of air temperature remain close to zero (max. 1.2 °C). The lowest temperatures were observed during the last week of December and followed by a relatively warm and cloudy period, as it is typical for the central Arctic [*24*]. Although the prevalence of low inversions, particularly during the cold season, causes

Table 4. *Monthly Means of Some Climatological Elements at "Drifting Station A"*

	Temperature, °C			Vapour pr. mb	Cloud amt. $N/10$	Wind, km/h (hourly means)	
	Mean	Daily var.	Interd. var.	Mean	Mean	Mean	Max.
1957 June	— 1.5	1.5	0.9	4.8	7.5	—	—
July	+ 0.1	0.7	0.6	5.8	8.5	14.5	32
Aug.	— 2.8	1.0	1.2	4.4	9.4	17.1	48
Sept.	— 11.6	1.1	2.5	2.2	8.1	10.6	40
Oct.	— 16.0	0.6	3.4	1.6	8.2	18.4	47
Nov.	— 27.7	1.0	2.7	0.5	5.0	10.8	39
Dec.	— 36.7	0.7	3.1	0.2	4.2	6.8	34
1958 Jan.	— 32.1	1.3	3.7	0.3	6.3	13.7	43
Feb.	— 36.1	0.9	3.4	0.1	3.2	10.3	40
Mar.	— 32.0	2.0	2.7	0.2	5.2	10.6	32
Apr.	— 28.4	4.0	1.4	0.3	3.5	6.6	29
May	— 13.4	2.5	2.0	1.8	6.4	10.6	26
June	— 1.8	0.8	0.9	5.0	8.9	11.7	32
July	+ 0.8	0.6	0.5	6.2	8.2	11.1	32
Aug.	0.0	0.7	0.7	6.0	7.7	9.3	29
Sept.	— 11.3	1.0	2.8	2.5	8.7	13.2	34
Oct.	— 22.1	1.5	4.2	0.8	5.8	13.7	42

great non-periodic variations of temperature the mean daily range remains small (Table 4). Daily variations of temperature exist only from March to September, with a maximum amplitude of 4° C in April. During the months without sun no daily period could be detected [*18*].

Vapour pressure was measured by means of psychrometers, in the instrument shelter about 1.5 m above the ground. The inherent inaccuracy of this method at low temperatures makes the results during the winter months somewhat unreliable. During July and August, vapour pressure reaches its maximum values, close to the saturation point at 0° C.

The amount of clouds also shows a great annual variation. The high degree of cloudiness during the summer months is due to the advection of warm air from the sub-arctic continents. During winter, the frequency distribution of cloudiness (Fig. 8) shows two maxima at 0 and 100% as is common at lower latitudes. The distribution during summer has only one dominating maximum at 100%. (The secondary maximum at

20–30% in winter would most likely disappear when using a greater number of observations.) A regular daily variation of the cloud amount is not apparent in the available data.

3. Heat Budget of Pond-free Ice

3.1. State of Surface

Since the evaluation of all data from Drifting Station A is not yet completed, the heat budget during a selected period of particular importance is treated in the following. According to Fig. 4 ablation between 10 and 23 July, 1957, was 11.0 cm on „dry" surfaces and 17.8 cm in ponds. A consideration of the heat budget during that time will be preceded by some general remarks.

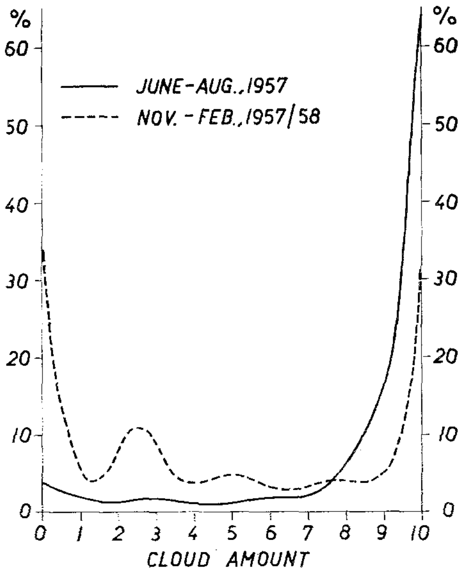

Fig. 8. Frequency of cloud amount, according to 3-hourly observations by the US Weather Bureau

It was pointed out by HOINKES [*19*] that, when comparing ablation with the heat budget it is necessary to compute the latter for periods with melting and frozen surface separately. Otherwise the heat sums cannot be related to the observed ablation because a loss of mass caused, during a period of positive heat budget, once it has occurred, cannot be restored by a following negative heat budget, that is to say, ablation may have occurred even though the total net heat during both periods may be negative. In the case of sea ice this consideration holds only to a limited extent since the meltwater formed is not entirely removed (ponds). For the area covered by ponds the above restriction is not valid. The accurate measurement of the thickness of temporary ice covers of the ponds, as well as the exact determination of the beginning and end of the freezing and thawing processes in general, is difficult. Especially towards the onset and termination of the ablation period the simultaneously existing snow, ice, and water-surfaces with their greatly varying heat absorption (albedo) make the "state of the surface" a poorly defined quality. The state of the various types of surface was noted several times daily. Based upon these observations it can be estimated that during about 7% of the period presently considered the surface was frozen. It was observed in many instances that, under an overlying frozen crust, melting proceeded at a depth of several centimetres. It seems, therefore, justified to treat the entire period as one with a melting surface.

In connection with this, the particular manner of disintegration of sea ice under the action of solar radiation may be pointed out. When pure glacier ice is exposed to short wave solar radiation an extremely porous surficial layer of a maximum depth of 15 to 20 cm is formed by internal melting. The coherence of the individual grains is not destroyed in this process, and the layer as a whole remains mechanically solid. On the other hand, one frequently finds on glacier tongues spindle-shaped areas of fine-grained ice whose surface appears similar to old firn. These areas represent sections of former crevasses which are filled with frozen water. The grain is columnar and perpendicular to the former walls of the crevasses. In contrast to the metamorphous regular glacier ice this ice disintegrates preferentially into small loose grains. A similar effect is seen on sea ice, although in that case the grain size seems to be greater.

The influence of this "firnification" of ice on the radiation balance is demonstrated by the following experiment. On 3 July, 1958, an area of about 5 m^2 was scraped with a bulldozer down to the compact ice. The albedo of this surface was 0.38. During the following two days of clear weather with an average insolation of about 20 cal/cm^2 hr the albedo increased to 0.65 as firnification progressed. The maximum thickness of the "firnified" layer at Station A was 15 cm. Its average density (end of July, 1958) was found to be 0.4 g/cm^3. The effect of this firnification is a substantial reduction of ablation rate by an increase of the reflectivity of the ice.

3.2. Net Radiation

3.2.1. Short-Wave

Global radiation was recorded by two solarimeters MOLL-GORCZYNSKI, one pointing upwards and one downwards and a self-balancing potentiometer. The calibration factor given by the manufacturing company was corrected for temperature, according to BENER [7]. The mean cloudiness from 10 to 23 July, 1957, was 94%. In view of such dominating importance of diffuse radiation it is justified to neglect the dependence of the calibration factor on solar elevation [*14*].

Daily averages of the albedo at the place of radiation measurements varied between 0.56 and 0.68. Further observations of the albedo were made with a portable device consisting of two solarimeters and a potentiometer. It was found that the mean albedo of clean, melting sea ice with a disintegrated surface lies between 0.62 and 0.70. Since the following considerations refer to that type of surface, an albedo of 0.66 has been assumed for calculations of short wave net radiation.

3.2.2. Long-Wave

Total radiation was recorded by means of a Schulze-radiometer [*30*] with hemispheres consisting of "Lupolen." This instrument requires separate calibration in the short-wave and long-wave range. Since a suitable standard instrument was not available at Station A, the short-wave calibration was done with the solarimeters. Calibration in the long-wave range by means of a "black body" (without ventilation) rendered unsatisfactory results. The proper method of calibrating the Schulze-radiometer is currently under discussion [*20*]. Awaiting a final

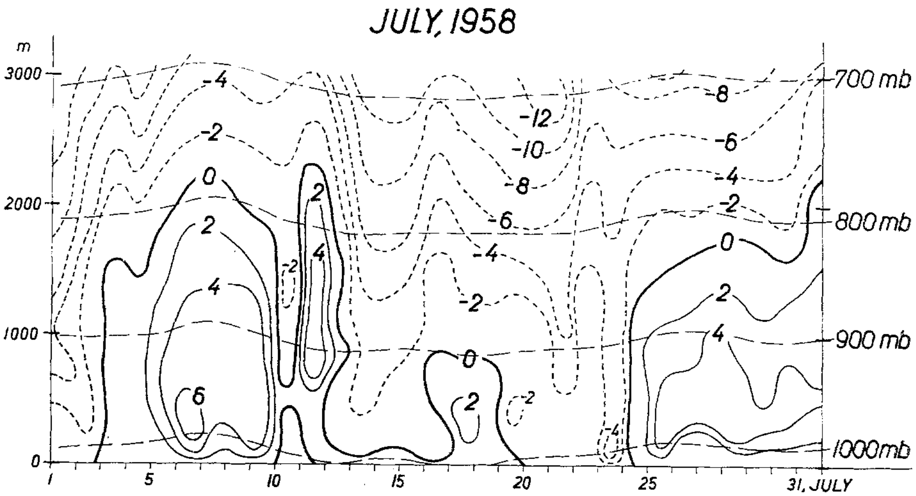

Fig. 9. Temperature in the first 3 km of the atmosphere over Drifting Station A, according to 12-hourly radiosonde observations by the US Weather Bureau

settlement of this question, long-wave radiation has been determined from radiosonde data.

The radiosonde observations of July, 1957, were not available at the time of writing, however, it seems possible to adapt the results of July, 1958, in an appropriate manner. In July, 1958, 8 ascents were made during clear weather (cloudiness $\leqq {}^1/_8$). Evaluation of the Elsasser-diagram (for the use of this diagram in clear and overcast conditions, see [*16*]) renders intensities of atmospheric radiation between 0.355 and 0.386, with an average of 0.370 cal/cm^2 min or 533 cal/cm^2 day. 15 ascents took place during overcast conditions (cloudiness $\geqq {}^7/_8$). The corresponding intensities lay between 0.436 and 0.463, with an average of 0.448 cal/cm^2 min or 645 cal/cm^2 day. While air temperature near the ground shows extremely small variations, great fluctuations were observed in the free atmosphere as a result of advection and vertical motion of the air (Fig. 9). Frequently, the temperature at cloud base is higher than at the ground, causing a positive net long-wave radiation.

Other means of determining atmospheric radiation are the empirical formulae of ÅNGSTRÖM and ASKLÖF (see [*17*]):

$$G_0 = \sigma T^4 (0.806 - 0.236 \cdot 10^{-0.052\, e}) \text{ and}$$
$$G_w = G_0 + k \cdot w (\sigma T^4 - G_0) \quad [\text{cal/cm}^2 \text{ min}]$$

(G = atmospheric radiation, e = vapour pressure in mb, w = cloudiness in tenths, k = factor depending on type of clouds and regional climatic conditions). By introducing for G_0 and G_{10} the values obtained from the

Table 5. *Daily Means of Cloudiness, Temperature, Vapour Pressure, and Daily Sums of Global Radiation ($S + D$), Atmospheric Radiation (G), Net Short-wave Radiation (NR_S), Net Long-wave Radiation (NR_e), and Net Total Radiation (NR). The Mean Albedo of the Pond-free Ice was 0.66. The Long-wave Radiation of the Melting Ice Surface was Taken to be 0.459 cal/cm². min*

1957 July	10	11	12	13	14	15	16	17	18	19	20	21	22	23		10th — 23rd
$\overline{N}/10$	9.0	7.4	9.7	10.0	9.9	10.0	10.0	9.9	9.4	9.3	10.0	9.8	7.4	9.8	means	9,4
$\overline{T}$, °C	−0.4	−0.2	0.8	0.6	0.7	0.8	0.3	0.1	0.0	−0.1	0.5	−0.1	−0.3	0.6	means	0,24
$\bar{e}$, mb	5.7	5.8	6.0	6.1	6.0	5.9	5.9	5.8	5.8	5.8	6.0	5.9	5.7	5.7	means	5,9
$S + D$ cal/cm²	522	605	322	302	328	428	399	402	559	400	320	425	532	280	totals	5824
G cal/cm²	643	633	660	662	660	665	660	641	642	648	662	657	639	660	totals	9132
NR_S	177	206	110	103	112	146	136	137	190	136	109	145	181	95	totals	1983
NR_e	−17	−27	0	2	0	5	0	−19	−18	−12	2	−3	−21	0	totals	−108
NR	160	179	110	105	112	151	136	118	172	124	111	142	160	95	totals	1875

ELSASSER-diagram, the second equation can be used to ascertain the factor k. It is 0.088, somewhat greater than in middle latitudes, due to the prevalence of inversions. Evaluation of the two equations with the actual meteorological data shows that during clear and overcast conditions they render an atmospheric radiation which is low by 17 and 3%, respectively. Under the assumption of a linear variation of this correction, daily sums of atmospheric radiation have been computed with the above formulae for the period from 10 to 23 July, 1957. The results are given in Table 5, together with the daily sums of global radiation ($S + D$), net short-wave radiation (NR_S) as well as mean daily cloud amounts ($\overline{N}$), air temperature ($\overline{T}$), and vapour pressure ($\bar{e}$). In computing the total net radiation the long-wave radiation of the melting ice surface was taken to be 660 cal/cm² day. The sum of net radiation over the entire period (NR) is 1975 cal/cm². This amount of energy is sufficient to melt 25.8 cm of pure ice. The actually observed amount of ablation was only 12.3 cm. This apparent discrepancy will be investigated in the following.

3.3. Sensible and Latent Heat

As mentioned in the previous chapter, the mean values of temperature and vapour pressure between 10 and 23 July, 1957, were close to the melting point of ice and to saturation at 0° C, respectively. It can a priori be assumed that the corresponding fluxes of heat and vapour will also be small. For the present purpose, an approximate calculation may suffice.

The mean wind velocity in July, 1957, at a height of 160 cm was 4.0 m/sec. Assuming a roughness parameter of $z_0 = 0.1$ cm and using PRANDTL's relation

$$\bar{u}_z = 5{,}75 \cdot u_* \cdot \log \frac{z - z_0}{z_0}$$

the friction velocity u_* becomes 21.7 cm/sec, and the eddy diffusivity $K_{160} = \rho\, u^* \varkappa\, (z + z_0) = 1{,}8$ or $K_{100} = 1.0$ g/cm sec. If temperature and vapour pressure follow a logarithmic distribution

$$T_z = T_0 + k_T \ln \frac{z + z_0}{z_0}, \quad e_z = e_0 + k_e \ln \frac{z + z_0}{z_0},$$

the mean vertical fluxes of sensible and latent heat

$$Q_S = 0.24\, K \frac{d\,T}{d\,z} \cdot t \quad \text{and} \quad Q_L = 600 \cdot K \cdot \frac{0.623}{p} \cdot \frac{d\,e}{d\,z} \cdot t \quad [\text{cal/cm}^2]$$

attain the following magnitude: $Q_S = 8.6$ cal/cm²day
$Q_L = -10.8$ cal/cm²day.

This means that the supply of sensible heat by dynamic convection is approximately compensated by the loss of heat by evaporation. Both amounts of heat are smaller by one order of magnitude than net radiation. Although this result represents only a rough estimate it cannot greatly deviate from the true values. If one chooses, instead of $z_0 = 0.1$, other values within the range of experience on glaciers [*19*, *32*] viz. $z_0 = 0.3$ and $z_0 = 0.05$, K_{160} becomes 2.1 and 1.65, respectively. The fluxes of heat become greater by 14% or smaller by 9% which amount to only a few calories per day. The error in the calculation of K introduced by neglecting stability should be even smaller.

In an investigation of the heat budget of arctic pack-ice based upon observations by the Soviet Drifting Station NP-II YAKOVLEV [*37*] comes to the conclusion that from May to August the very large amount of 5700 cal/cm² was lost by evaporation. The period of melting at NP-II lasted from about 10 July to 20 August, 1950 [*36*]. During a great part of that period the mean vapour pressure was close to the values observed at Drifting Station A. In the available condensed version [*37*] of YAKOVLEV's original publication the methods used in calculating the heat budget are, unfortunately, not communicated. It is, therefore, at present not possible to investigate the cause of the great difference in our findings. A possible explanation is offered in the following chapter.

3.4. Internal Melting, "Heat Reserve"

The ice of glacier tongues is normally contaminated to some degree by debris and cryokonite. In that case, the error committed by allocating the entire absorption of short-wave radiation to the surface is comparatively small. In the case of sea ice the amount of energy penetrating below the surface cannot be neglected.

The mean vertical distribution of the salinity of old sea ice according to SCHWARZACHER [*29*] is shown in Fig. 10, curve S. It is generally assumed that the brine is always in phase equilibrium with the surrounding ice so that the salt concentration in the cavities is a function of temperature alone. The calculation of the conduction of heat is complicated by the fact that thermal conductivity, specific heat, and density are functions of salinity and temperature. Our basic knowledge of the physical constants of sea ice is due to MALMGREN [*25*]. Important progress has been made by the recent work of WEEKS, ANDERSON, and ASSUR [*3, 35, 2, 5*].

Of particular importance in the present considerations is the specific heat of sea ice, c_S. It is composed of the specific heat of pure ice increased by the heat required for melting a certain amount of ice to lower the brine concentration. The values of c_S given by MALMGREN can be represented with good approximation by an analytical expression

$$c_S = c + \beta \frac{S}{T^2} \quad [\text{cal/g}\ {}^0\text{C}], \tag{3}$$

c being the specific heat of pure ice (0.5), S the salinity of the ice in $^0/_{00}$, T the temperature in negative degrees Celsius, and β a constant $= 4.1$. The heat Q required to raise the temperature of 1 cm^3 of ice is

$$d\,Q = c \cdot \rho \cdot d\,T. \tag{4}$$

By introducing (3) in (4) and integrating over the temperature interval $T_2 - T_1$ we obtain, under the assumption of constant salinity and density

$$Q = c \cdot \rho\,(T_2 - T_1) + \rho \cdot \beta \left(\frac{S}{T_1} - \frac{S}{T_2}\right)$$

or in a more practical form

$$Q = \rho\,\Delta T \left(0.5 + \frac{4.1\,S}{T_1\,T_2}\right) \tag{5}$$

Fig. 10 shows the variation of ice temperature from July, 1957, to May, 1958, at Station 1. The vertical profiles of temperature at the beginning and end of the period whose heat budget is considered are given as curves T in Fig. 11. Between 10 and 23 July, 1957, the average heating was 0.8° C. At both dates, the level of minimum temperature is about 220 cm below the surface. Conduction from deeper levels does not contribute to the heating of the ice above that level. Curves T and S in Fig. 11 provide the necessary data for an evaluation of eq. (5). The energy required for the observed rise of temperature from 10 to 23 July, 1957, was computed for layers of 10 cm thickness between the surface and a depth of 220 cm. The result is presented as curve E in Fig. 11. Summation of this curve, beginning at the level of minimum temperature upwards to a level 10 cm below the surface renders a total of 700 cal/cm². 12 cm of ice were melted at the surface, and it is likely that a certain amount of brine drained from the porous upper layer. Therefore, the present calculation must be inaccurate at shallow depths. Apart from the energy used for surficial ablation, the energy consumed for internal melting and heating was drawn from the positive radiation balance as observed at the surface.

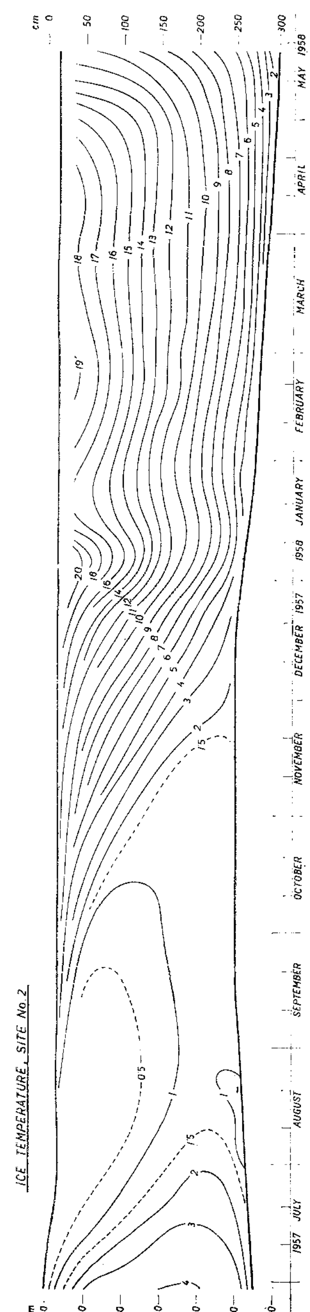

Fig. 10. Isopleths of ice temperature, Station 1, site No. 2

In this connection it seems of interest to estimate the entire amount of latent heat stored in the ice at the time of maximum temperature. At the end of August the total brine volume in 300 cm thick ice is 32.9 cm³/cm². By the time of minimum temperature in March (Fig. 10) it is reduced to 3.4 cm³/cm², whereby 2400 cal/cm² of latent heat are released. Another 600 cal/cm² of latent heat are stored in the meltwater ponds, and 1700 cal/cm² must be extracted from the ice to lower its temperature from August to March. The heat reserve of an ice sheet of 300 cm thickness in August, with reference to its temperature in March, is then 4700 cal/cm² (Table 7). This amount is about three times as great as the one given by YAKOVLEV [*37*].

3.4.1. Extinction of Solar Radiation

It is evident that the comparatively large amount of energy of 700 cal/cm² which reached the deeper levels between 10 and 23 July could not do so by conduction alone. With a thermal conductivity of the ice of 0.0045 [CGS] the heat flux by conduction becomes of the order of 70 cal/cm² in 14 days. It is further seen from curves T in Fig. 11

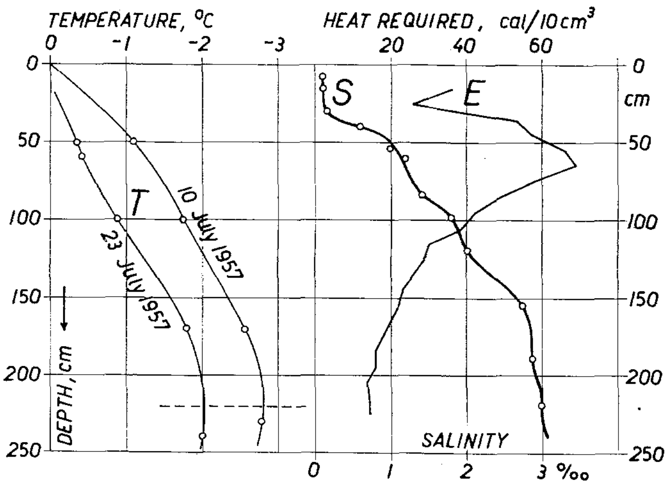

Fig. 11. The distribution of energy (E) required to raise the temperature (T) of ice with a given salinity (S) is used to compute the extinction coefficient for short wave radiation (see Table 6)

that, at depths between 50 and 150 cm, the temperature gradients and therefore the heat fluxes by conduction on 10 and 23 July were approximately the same which means that conduction did not contribute to the heating of this layer. Under these circumstances it seems justified to ascribe the heating of this layer entirely to the absorption of short-wave solar radiation. In any case, the error by neglecting the effect of conduction will be small. From the ratio of the amounts of energy transmitted through each 10-cm-layer the coefficient of extinction may be computed. The results (Table 6) are surprisingly similar to the ones obtained by AMBACH [*1*] by direct measurements on glacier ice. For depths $>$ 15 cm AMBACH finds a mean value of 0.018 cm^{-1} while the present indirect method leads to an average extinction coefficient between 50 and 150 cm depth of 0.015 cm^{-1}. In view of the apparently great importance of penetrating radiation for the total heat budget, more accurate measurements, under inclusion of selective absorption, are desirable. If absorption at deeper levels does not differ greatly from the one in the intermediate layer, the amount of energy penetrating an old ice floe of 3 m thickness (without snow cover) in July would be of the order of 1 cal/cm^2 per day.

3.5. Components of the Heat Budget

The total heat budget during the period 10 to 23 July, 1957, is summarized in Table 7. Approximately 0.5 cm of solid precipitation fell during that time. If this amount is added to the observed 11 cm of surface ablation (water equivalent) the difference between observed and computed ablation becomes 19%. The limited accuracy of ablation measurement as well as small variations in the distribution of salinity may easily account for such a difference.

Table 6. *Absorption of Short-wave Radiation by Sea Ice, as Computed from the Energy Required for an Observed Increase of Temperature and Brine Volume (see Fig. 11)*

Depth cm	Energy 10-23 July, 1957 cal/cm^2	Extinction coefficient 10^{-3} cm^{-1}
surface	1844	..
10	685	..
20	650	..
30	625	..
40	576	..
50	522	14
60	457	15
70	390	16
80	334	15
90	288	15
100	247	16
110	211	14
120	184	13
130	161	15
140	138	18
150	116	..
160	96	..
170	77	..
180	59	..
190	43	..
200	28	..
210	14	..

In accordance with the results of YAKOVLEV [*37*], although to an even higher degree, solar radiation is shown to be the dominating factor in the ablation of ice during the summer. The smallness of the convective heat fluxes is undoubtedly caused by the fact that, along the great distance between the station and the nearest coast, heat and water vapour of the warm continental air have been given off to the ice. The remaining gradients of temperature and vapour pressure are too small to render appreciable fluxes. The warm air at higher levels manifests itself primarily by the high degree of cloudiness and its influence on radiation. During overcast conditions in summer practically the entire short-wave radiative energy received at the surface is available for ablation. Under clear conditions the drop of atmospheric radiation may, even in July, result in a negative total heat balance (freezing). Evaporation between 10 and 23 July, 1957, was very small (154 cal/cm^2). Its contribution to ablation was less than 0.3 mm. The same goes for ablation by rain [*19*]. It seems important to note (Table 7) that 38% of the positive heat budget are transformed in the interior of the ice and contribute to the build-up of the great heat reserve previously discussed.

In a study of radiation data from the ice island T-3 in the Arctic ocean (1953: 85° N, 90° W) FRITZ [*14*] points out difficulties in reconciling the high values of net radiation with the fact that practically no surface ablation was observed. The short-wave net radiation in July, 1953, was about 100 cal/cm^2 day. According to FRITZ, 15% of this energy were used to heat the ice, and the rest was supposedly removed by convection. FRITZ bases his explanation on the fact that the top of the cloud

Table 7. *Energy Balance of the Ice During Maximum Melting Conditions, and Total Heat Reserve of the Ice Sheet*

10—23 July, 1957, cal/cm^2			mean, cal/cm^2 day
net radiation	short-wave	+ 1983	+142
	long-wave	— 108	— 8
sensible heat		+ 120	+ 9
latent heat		— 151	— 11
	total	+ 1844	From this, the following
amounts are used for:			
surface ablalion		1144	= 62%
heating and internal melting		700	= 38%

Heat reserve in August, with reference to the ice temperature in March, in cal/cm^2 (ice thickness ~ 300 cm).	ponds	600
	brine	2400
	cooling	1700
	total:	4700

cover continually gives off heat by long-wave radiation. This he assumes to result in super-adiabatic lapse rates at lower levels and a corresponding upward transport of heat. Indeed, the radiosonde observations from T-3 [*6*] frequently showed such lapse rates in the lowest 100 m of the atmosphere. In the previous year (July, 1952) this was, however, not the case [*6*]. At Drifting Station A, too, only few such observations were recorded. Above the shallow layer with unstable lapse rates there was in all cases an almost isothermal zone of 1000 m thickness [*6*] in which the convective heat flux must have been downwards. The loss of heat from the cloud top can, therefore, not be compensated from below. In view of the high percentage of radiation penetrating clean ice it seems possible that, during July, 1953, a good part of the net short-wave radiation on T-3 only increased the porosity of the ice and was not realized as surface ablation. The shorter the period of ice ablation the more likely becomes an effect of that nature. However, according to CRARY [*11*] an appreciable amount of ice ablation in summer seems to be the rule on T-3.

4. Yearly Totals of Heat. Comparisons

The present data may be used for an estimate of the yearly turnover of heat in the Arctic pack ice. Unfortunately, no measurements of the turbulent flow of heat in the ocean could be made at Drifting Station A. According to YAKOVLEV [37], the ice receives an amount of 5.5 kcal/cm^2 year from below, 0.9 from May to August, and 4.6 from September to April. From the data of the Soviet Drifting Station NP-II the following energy budget for the months May–August was derived [37]:

Net radiation	+ 11.0 kcal/cm^2
Latent heat..............	— 5.7
Sensible heat.............	— 2.2
Heat of melting..........	— 2.9
Heat reserve of the ice ...	— 1.5
Heat flow from the ocean.	+ 0.9

The total of heat received at the upper and lower surface was 11.9 kcal/cm^2. It compensates for the heat loss by evaporation and convection and provides the energy for ablation and the build-up of the heat reserve. The atmosphere looses 3.1 and the ocean 0.9 kcal/cm^2 in the process. The conspicuous features of this energy budget are the very large amount of evaporation and the small amount of the heat reserve. The smallness of the latter (1.5 kcal/cm^2) suggests that it may have been computed with the specific heat of pure ice, and that internal melting has not been allowed for. Since YAKOVLEV determined evaporation as a residual value its overestimation seems to be caused by an underestimation of the heat reserve. The ice temperatures at Station A were well within the range of previous observations [36], and the maximum total brine volume (end of August) as calculated from temperature and salinity is supported by the direct observations of SCHWARZACHER [29]. The value of the heat reserve given in chapter 3.4 should, therefore, be essentially correct. If we introduce our value of 4.7 kcal/cm^2 in YAKOVLEV's energy budget and deduct the increase of 3.2 kcal/cm^2 from evaporation, the latter becomes 3.8 kcal/cm^2 from May to August or, on an average, 31 cal/cm^{2}day, a much more probable value.

It is, at present, not possible to give a detailed analysis of the atmospheric and solar components of the heat budget at Drifting Station A. However, from the observation of the mass budget, ice temperature, and salinity a cumulative heat budget can be obtained. Its constituents are denoted as follows:

S = heat for melting of snow cover,
A_s = heat for run-off part of surface ice ablation,
A_b = heat for bottom ablation,
R = heat reserve of the ice sheet, as defined in Table 7,
B = heat released by bottom accretion,
Q_w = total heat drawn from solar and atmospheric sources during the "warm" season, May–August,
Q_c = total heat given off to the atmosphere during the "cold" season, September–April,

q_w = heat drawn from the ocean during the warm season,
q_c = heat drawn from the ocean during the cold season.

It must be emphasized that the division of the year in a period with a positive and one with a negative heat budget of the ice surface is only a crude approximation. During the warm season the ice sheet receives heat from both sides. Its cumulative heat budget can be written

$$S + A_s + A_b + R = Q_w + q_w. \qquad (6)$$

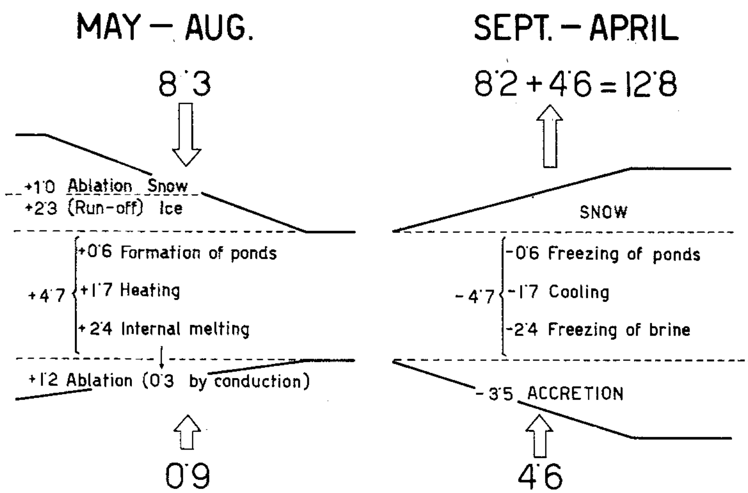

Fig. 12. Totals of heat exchange at the boundaries and in the interior of the ice, in kcal/cm²

During the cold season it receives heat at the lower and looses heat at the upper surface. The total upward heat flow is

$$R + B + q_c = Q_c. \qquad (7)$$

The condition of stationary ice thickness is

$$A_s + A_b = B. \qquad (8)$$

By combination of eqs. (6), (7), and (8) it is seen that

$$q_w + q_c - S = Q_c - Q_w. \qquad (9)$$

The deposition of snow on the ice is an increase of mass which is not associated with a transformation of heat at the earth's surface. Eq. (9) shows that the net annual amount of heat provided to the atmosphere is equal to the upward heat flux in the ocean less the heat required for melting of the snow cover.

Average values of S, A_s, A_b, and R are known from the observations previously discussed. Using the values for the heat flux in the ocean given by Yakovlev, a cumulative heat budget as presented in Fig. 12

is obtained. The temperature at the ice-ocean interface is normally — 1.7° C. During a short time in late summer the heat flux by conduction in the ice downward at all levels (see Fig. 11). An estimated amount of 0.3 kcal/cm² reaches the lower surface and causes, together with the upward heat flux in the ocean, bottom ablation [22]. Its observed amount seems to substantiate YAKOVLEV's value of the summer heat flux in the ocean. A new instrument ("turbulimeter") described by KOLESNIKOV et al. [23] should be particularly effective in measuring turbulent heat flow in the ocean.

According to Fig. 12 the ice receives during the warm season 8.3 kcal/cm² from solar and atmospheric sources. A total of 12.8 kcal/cm² is returned during the cold season. In the course of a year the atmosphere in the Central Arctic receives 4.5 kcal/cm² which originate from oceanic advection. Further evaluation of the data obtained at Drifting Station A should provide a more detailed understanding of the solar and atmospheric components of the heat budget and their relative importance, as well as clues to the possible effect of climatic variations on the maintenance of the ice cover.

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