HEAT BALANCE STUDIES ON SEA ICE NEAR SYOWA STATION, EAST ANTARCTICA

Nobuyoshi ISHIKAWA, Shun'ichi KOBAYASHI,

The Institute of Low Temperature Science, Hokkaido University, Kita-19, Nishi-8, Kita-ku, Sapporo 060

Tetsuo OHATA

Water Research Institute, Nagoya University, Furo-cho, Chikusa-ku, Nagoya 464

and

Sadao KAWAGUCHI

National Institute of Polar Research, 9–10, Kaga 1-chome, Itabashi-ku, Tokyo 173

***Abstract*:** Heat balance studies were carried out on the Antarctic sea ice surface in the austral spring and summer of 1980. The surface albedo of sea ice covered by a thin snow-layer was kept nearly constant (around 0.8) in spring, but in summer it was reduced to the same value as that of bare ice owing to the change of the surface properties with a great increase in the amount of absorbed net radiation. Variations of heat balance components were presented for every 10 days and two seasons in this paper.

1. Introduction

Sea ice surrounding the Antarctic Continent changes with time, and is considered to have some influence upon the Antarctic climate. Heat balance studies on sea ice were attempted under the program of the Polar Experiment in Antarctica (POLEX-South).

Research into the growth of sea ice and heat budget in the polar region has been widely pursued, *e.g.* by UNTERSTEINER (1961) and WELLER (1968), who conducted field investigations, and by MAYKUT and UNTERSTEINER (1969) and PARKINSON and WASHINGTON (1979), who did numerical analyses. Meteorological observations were made near Syowa Station by MAKI (1974), who described the characteristics of turbulence on sea ice.

This paper presents some preliminary results of the heat budget on sea ice.

2. Instrumentation

A fixed meteorological site (A) was set up on fast floating sea ice about 50 m from Ongul Islands on which Syowa Station (69.0°S, 39.6°E) is located, in Lützow-Holm Bay, East Antarctica. The sea ice in this region was almost flat and covered by snow or snow-ice a few centimeters thick. Snow is fairly different from underlying bare sea ice in optical properties such as surface albedo and extinction

Table 1. Observed quantities and instruments.

Site	Item	Instrument
A	Radiation	
	Global solar radiation (1.5 m)	Pyranometer (EKO MS-42)
	Reflected short-wave radiation (1.5 m)	〃 (〃)
	Net radiation (2.0 m)	Balance meter (EKO CN-11)
	Temperature	
	Air temperature (0.1, 0.5, 1.0, 1.7 m)	Platinum resistance thermometer (R900-12)
	Snow, ice and water temperature (0.0, −0.1, −0.2, −0.3, −0.4, −0.5, −0.6, −0.7, −0.8, −0.9, −1.0, −1.5 m)	Thermo-junction thermometer
	Wind speed and direction (6 m)	Three-cup anemometer with a wind vane (MAKINO GRS 4054)
	Dew point temperature (2.0 m)	Lithium chloride dew-point hygrometer (Dewcel)
	Sensible heat flux (1.2 m)	Ultrasonic anemothermometer (KAIJÔDENKI DAT-100)
	Ice thickness	Automatic sea ice thickness device (YAHAGI, 1980)
	Ablation amount	Snow stakes
	Surface condition	8 mm interval camera
B	Radiation	
	Reflected short-wave radiation (1.0 m)	Pyranometer (Moll-Gorczynski)
	Temperature	
	Snow, ice and water temperature (0.0, −0.1, −0.2, −0.3, −0.4, −0.5, −0.6, −0.7, −0.8, −0.9, −1.0, −1.5 m)	Thermo-junction thermometer
	Ablation amount	Snow stakes

coefficient; therefore, another experimental site (B) 5×5 m in area was established by removing a thin snow-ice layer from a stretch of bare sea ice for measuring surface albedo.

Instruments and observed quantities are listed in Table 1. Meteorological observations were also carried out at Syowa Station.

3. Observational Results

3.1. Variations of meteorological elements

Variations of meteorological elements are shown in Fig. 1, using mean values of 5 days from measurements at site (A) denoted by Nisi-no-ura and Syowa Station. It follows from the figure that in both sites incoming short-wave radiation, air temperature and water vapor pressure increase rapidly after mid-October, but relative humidity remains almost unchanged. Those elements taken at sea ice are compared with those at Syowa Station. On the average, water vapor pressure and relative humidity are higher over sea ice than at Syowa. Before late October the extreme and mean temperatures are lower over ice than at Syowa Station. This is expected since the meteorological site of Syowa Station is 21 m above sea level,

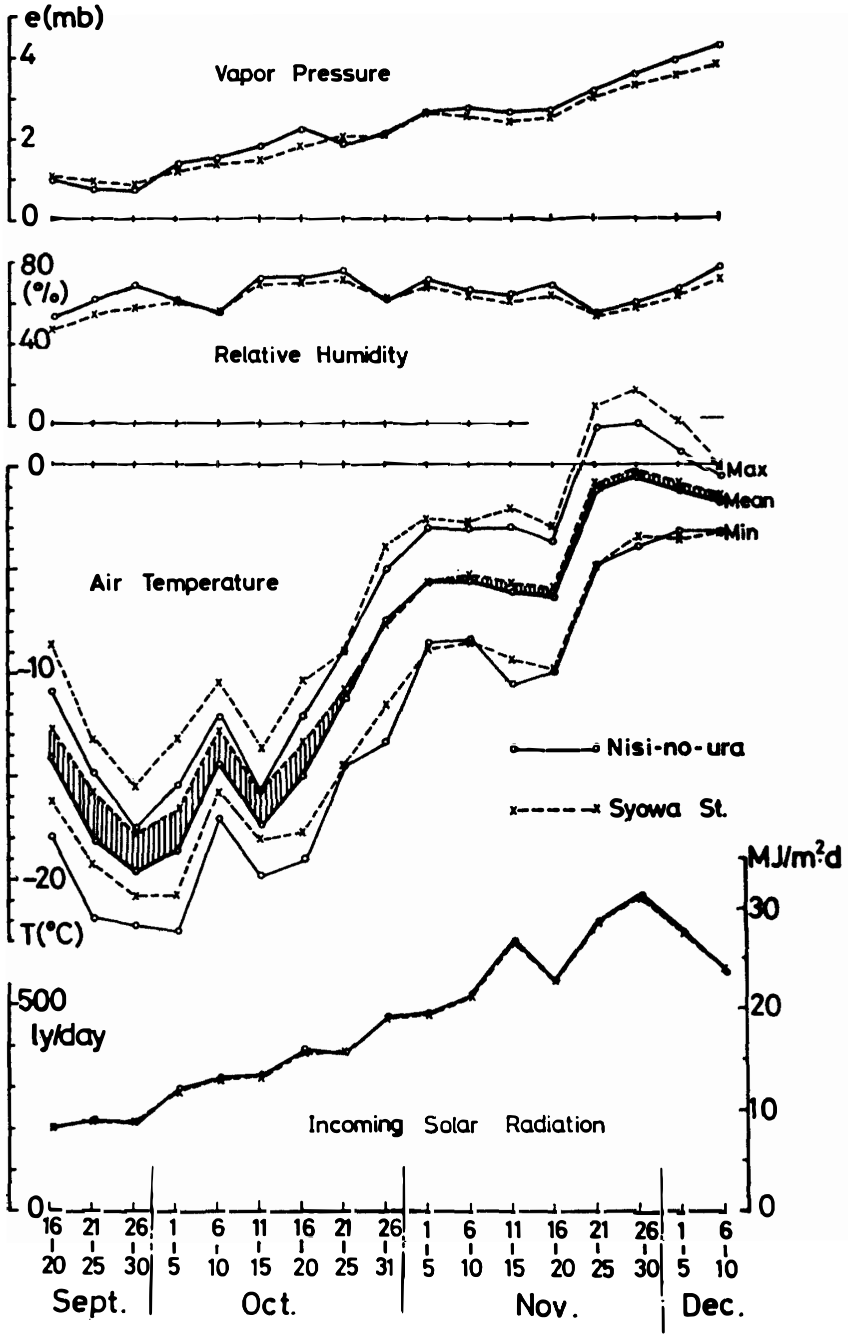

Fig. 1. Variations of meteorological elements at site (A) and Syowa Station in 1980.

and there was a surface temperature inversion during the observational period. After mid-November the minima and means of air temperature had nearly the same values at both sites, respectively, but the maxima show some discrepancies between two sites. This was because snow-cover which had been deposited over the island disappeared in November, and bare rocks were exposed. As a result, the bare rocks were heated by solar radiation, causing the temperature to rise. But the ice surface was not allowed to warm itself above the melting point even if the solar elevation was high. On the other hand the surface layer contained liquid water from internal melting. This water tended to freeze and prevented the surface temperature from dropping. A comparison of meteorological elements between different surfaces (*e.g.* moraine and ice) reported by WENDLER and ISHIKAWA (1973) gave similar results.

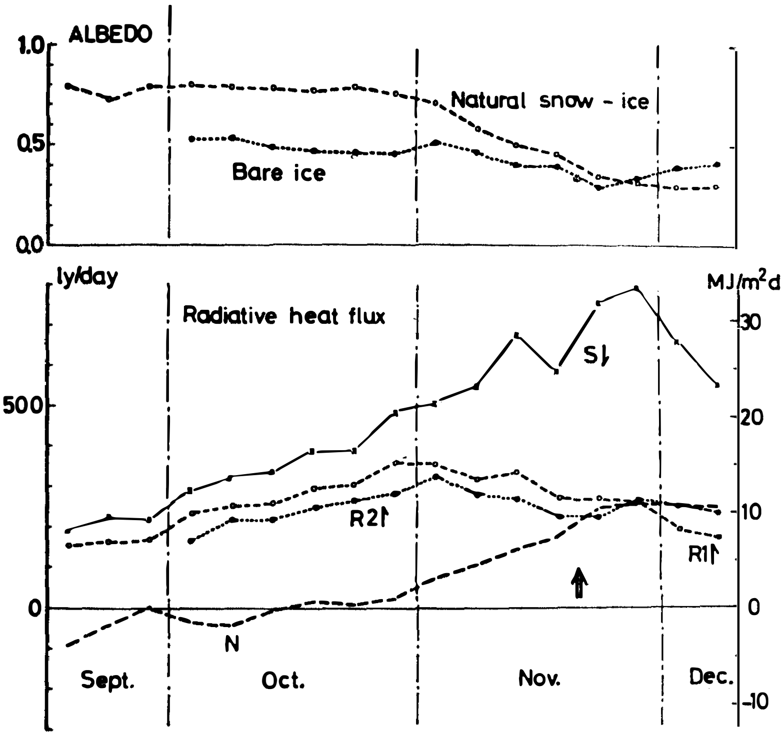

Fig. 2. Mean values of radiation components for every 5 days at sites (A) and (B) in 1980.

Figure 2 shows the variations of 5-day means of the short-wave (S↓ and R1↑) and net radiation (N) over the natural snow-ice surface at site (A). Values of reflected short-wave radiation (R2↑) and surface albedo measured at experimental site (B) are also presented in this figure. Incoming short-wave radiation increased monotonically from September. Temporary decreases of the solar radiation appeared in mid-November and early December, when cloudy weather lasted for several days. Surface albedo of the natural surface (snow-ice) was high (around 0.8) in spring and gradually decreased to the same value as that of bare ice in November owing to the reducing of thickness of the snow-layer on the ice. The albedo of bare ice was low (0.5–0.3) during this period, but in early December it increased owing to the formation of micropenitents of snow on the ice surface. The net radiation was negative before mid-October, but the value rapidly increased after November on account of the decrease of the surface albedo.

The maximum value of ice thickness was about 120 cm in mid-October, and the lower boundary temperature of sea ice was kept at the freezing temperature during the observational period. Figure 3 shows vertical profiles of sea ice temperature below the depth of 10 cm from the surface at four different times. After mid-October the sea ice temperature rose rapidly, and in late November an

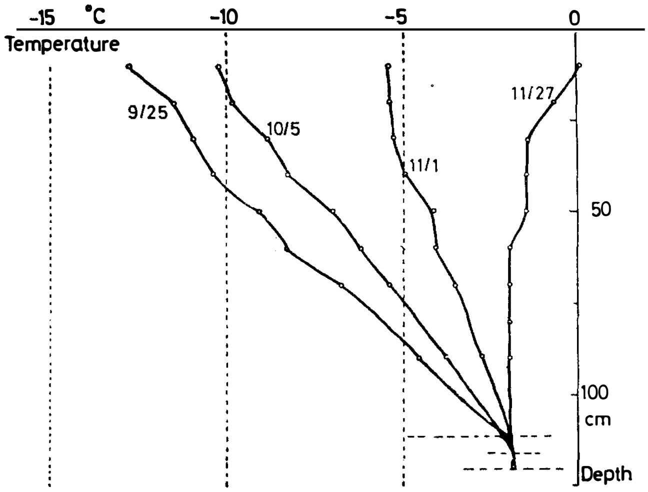

Fig. 3. Vertical profiles of sea ice temperature below the depth of 10 cm from the surface at four different times.

isothermal temperature profile was observed through the whole body of the ice sheet.

3.2. *Heat budget of sea ice*

The heat balance at the natural surface (A) was estimated over one hour periods and summed for every 10 days. Figure 4 shows the heat balance model used in the present investigation. Unfortunately, the exchange amounts of heat flow between the lower boundary of sea ice and underlying sea water cannot be estimated. Furthermore, an accurate surface temperature is difficult to obtain in the ablation season. Therefore, the volume-heat balance is considered in the layer between the upper boundary and some depth from the surface (ΔZ) instead of the heat balance at the surface or the entire ice sheet. Before the formation of puddles (internal melting pool) the depth of 10 cm was taken as ΔZ: after puddles form, the thickness from the surface to the upper boundary of puddles was taken as the layer. Snow and ice are considered as semi-transparent media to short-wave radiation.

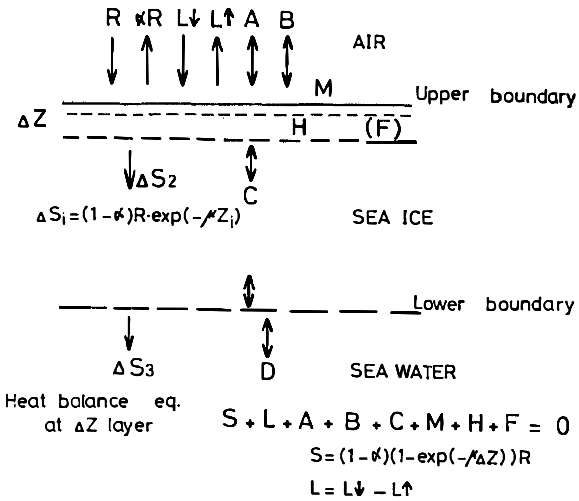

Fig. 4. Schematic model of heat balance.

Therefore, the absorbed amount of net short-wave radiation in the ΔZ layer is estimated instead of that at the surface. The heat budget equation of the ΔZ layer is considered to have the form:

$$S+L+A+B+C+M+H+F=0, \qquad (1)$$

where S is the extinction amount of net short-wave radiation in the layer, L is the net long-wave radiation (effective terrestrial radiation) at the surface, A and B are the sensible and the latent heat fluxes at the surface, respectively, C is the heat flux conducted through the lower boundary of the ΔZ layer, M is the heat used by surface melting, H is the change of heat storage in the ΔZ layer and F is the remainder term. Each component of the heat balance equation was obtained as follows:

1) Extinction amount of net short-wave radiation S:

$$S=\Delta S_1-\Delta S_2 : \Delta S_i=(1-\alpha)\exp(-\mu Z_i)R, \qquad (2)$$

were α is the surface albedo, μ is the extinction coefficient, Z_i is the depth and R is the amount of incoming short-wave radiation at the surface. Extinction coefficients μ 0.13 cm^{-1} for ice and 0.5 cm^{-1} for snow are used in the paper, which are derived from laboratory experiments at the Institute of Low Temperature Science by ISHIKAWA and ISHIDA (1970) and FUKAMI and KOJIMA (1980).

2) The net long-wave radiation L was estimated from the difference between the amount of all-wave radiation and the short-wave radiation balance at the surface.

3) Sensible heat flux A:

$$A=\rho_a C_p \overline{W'T'}, \qquad (3)$$

where ρ_a and C_p are the air density and specific heat capacity, respectively, W' and T' are fluctuations in vertical wind component and air temperature, which are obtained directly by the ultrasonic anemothermometer.

4) Latent heat flux B:

$$B=a(b+c\bar{V})\Delta e, \qquad (4)$$

where a is the latent heat of sublimation, $\bar{V}$ is the wind speed at 6 m height, Δe is the vapor pressure difference between 0 and 2 m heights, and constants (b and c) are adopted from KUZ'MIN's equation (1972).

5) Conductive heat flux C:

$$C=k\frac{\Delta\theta}{\Delta Z_b}, \qquad (5)$$

where k is the thermal conductivity of sea ice, and $\Delta\theta/\Delta Z_b$ is the temperature gradient of the sea ice between the bottom of sea ice and the lower boundary of the ΔZ layer.

6) Surface melting heat flux M:

$$M=79\rho_i\Delta h', \qquad (6)$$

where ρ_i is the sea ice density and the ablation amount $\Delta h'$ is measured by the stake method.

7) The change of heat storage H:

$$H=\rho_i C_i \int_0^{\Delta Z} \frac{\partial \bar{\theta}}{\partial t} \mathrm{d}Z, \tag{7}$$

where C_i is the specific heat capacity of sea ice and $\bar{\theta}$ is the mean temperature of the ΔZ layer.

Physical properties of sea ice (k, ρ_i, C_i) are quoted from ONO (1968). Fluxes toward ΔZ layer are regarded as positive, and those away from it as negative. The positive and the negative signs of M mean freezing and melting at the surface. The sign of the heat storage amount H is positive for heat loss and negative for heat gain. Variations of heat balance components are shown in Fig. 5 using 10 day

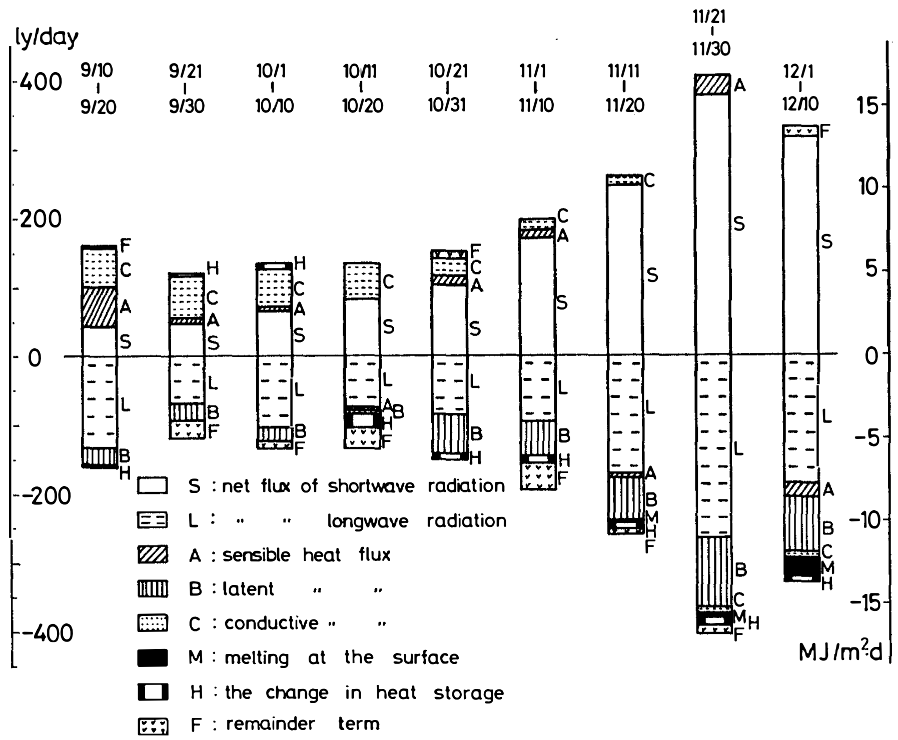

Fig. 5. Variations of heat balance components for every 10 days on sea ice in 1980.

means and in Fig. 6 for two seasons (austral spring and summer), and the percentage components are also presented in Fig. 6.

The main heat source is short-wave radiation in both seasons. In spring the negative heat sources are long-wave radiation and latent heat amount for sublimation, but absorbed solar radiation cannot compensate for the heat loss, therefore, the conductive heat from the bottom of the sea ice becomes positive, which implies a growth of sea ice thickness. In summer the heat source is only by short-wave radiation, and the sum of long-wave radiation and latent heat flux of sublimation are almost balanced with the heat source.

Sensible heat flux has both signs and is changeable in a short time period depending on the air temperature gradient. But for a long time period it has a small positive sign and does not contribute much to the heat balance. The sign of heat storage is positive but the amount is small before the end of September. After mid-October it turns negative, and which means the sea ice is heated up.

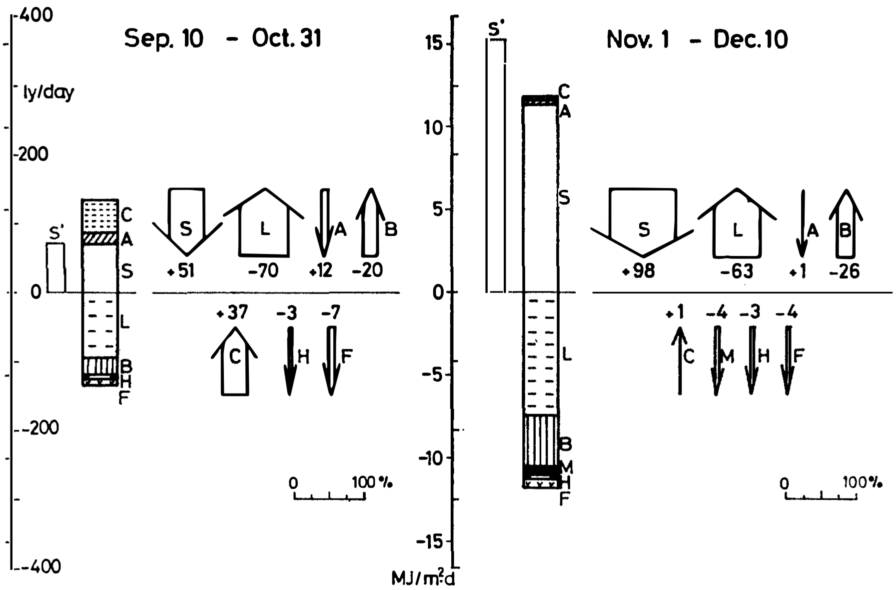

Fig. 6. Seasonal values and percentage components of the heat balance in the austral spring and summer in 1980.

The component F is calculated by eq.(1) as the remainder term, but it might be considered as internal-melting or refreezing of sea ice in the ΔZ layer.

In Fig. 6 the amount of the short-wave radiation balance at the surface is shown using the suffix S'. In spring sea ice is covered by a snow-layer several centimeters thick. Surface albedo is high, and the extinction coefficient of snow is very large. Therefore, the amount of short-wave radiation balance at the surface is nearly the same as the amount of short-wave radiation absorbed in the ΔZ layer. After bare sea ice is exposed, surface albedo becomes low, and the extinction coefficient of ice is small. The difference between S' and S is about 100 ly/day (4.2 MJ/m^2·day) in summer, which passes through the ΔZ layer. It is absorbed in a deeper layer and produces the internal melting layer (puddle). The authors observed the formation of puddles starting the first step. Puddles were formed at a depth of 12–15 cm below the surface only in the bare ice region in late November, when the air temperature was still below the freezing point. In mid-December the ice crust lying on the puddles melted away, leaving puddles exposed. Descriptions of puddles near Syowa Station were given by TAKAHASHI (1960) and ENDO (1970). The mechanisms of puddle formation will be presented in another paper by the authors.

Acknowledgments

The research was supported by the members of JARE-21 wintering team, especially by Messrs. R. KATO, Y. KATSUTA, K. SHIRAISHI and Dr. Y. NAITO. The authors would like to express their thanks to Mr. H. MOTOYAMA and Dr. N. ONO, of the Institute of Low Temperature Science, Hokkaido University, for reducing the data and providing useful advice. Prof. H. YAHAGI, of Kushiro College, Hokkaido University of Education, made to automatic recording device for sea ice thickness, Dr. E. AKITAYA, of the Institute of Low Temperature Science, Hokkaido University, made long-term temperature recorders and an 8 mm interval camera system and Mrs. N. OCHIAI typewrote the manuscript, for which the authors are grateful.

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(Received May 10, 1982; Revised manuscript received August 9, 1982)