Observations of Radiative Cooling at Basins in Midwinter and Snow-Melting Season

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Abstract

Meteorological observations were carried out at two places in mountainous regions in the midwinter and snow-melting seasons of 1971 and 72; each place was represented by two nearby stations, higher and lower, having an altitude difference of 250 m. The following two facts were found: firstly, there were differences in sensible heat flux between the two stations under temperature inversion; secondly, the intensity of inversion was much greater in the midwinter season than in the snow-melting season.

1. Introduction

There are several reports about temperature inversion at mountainous regions (Lake 1956, Oke 1970, Benson 1970), but there are few dealing with the heat balance under such a condition (Wendler 1971, Kojima *et al.* 1971). The present observations were undertaken for the purpose of investigating the conditions which give rise to temperature inversion in a mountainous region and to the differences in inversion strength between the winter and snow-melting seasons. The term 'temperature inversion' used in this paper means that the temperature itself is higher at an upper station than at a lower station.

2. Period of observations and instrumentation

Midwinter observations were made for 11 days from February 1 to 11, 1971, at the two stations at Toikanbetsu in the Teshio mountainous region in the northern part of Hokkaido (Fig. 1.). The lower station B is situated at the center of the basin at a height of 10 m above sea level and the higher station A on the top of the surrounding mountain nearby with a height of 235 m. The horizontal distance between them was about 5 km. The air temperature was measured with a ventilate thermocouple, which was sheltered against solar radiation; the wind speed was obtained with a thermister anemometer; the radiation was measured with a CN-1 type net radiometer. At the two stations these

meteorological elements were obtained at a height of 1.0 m above the snow surface, and the surface temperatures were also measured. More detailed measurements were carried out for the study of heat balance during the 11 days from January 14 to 24, 1972, at the same place in order to obtain the temperature profile, wind velocity profile, relative humidity, condensation (evaporation) amount and net radiation flux.

The observations in the snow-melting season were carried out for 8 days from April 15 to 22, 1971, at the two stations at Moshiri in the Uryu mountainous region in the central part of Hokkaido (Fig. 2.). Station B is located in the center of the basin with an altitude of 300 m

Fig. 1. Location map of the two heat balance stations of Toikanbetsu.

A: the higher station

B: the lower station

and station A was on the top of a mountain nearby at a height of 550 m above sea level, and measurements of air temperature, wind speed and net radiation flux were carried out at a height of 1.0 m above snow surface. In both seasons

the ground was covered with snow.

According to the data obtained in the past three years, the probability of occurrence of nocturnal inversion was once every two days during January and February at Toikanbetsu. Because even a stable inversion resolved itself in the daytime (Ishikawa and Ishida, 1971), the measurements were made only at night (from 6 p.m. to 6 a.m.).

3. Meteorological conditions during the observations period

Fig. 3 represents the variations in mean hourly values of meteorological elements in the midwinter at Toikanbetsu. Solid lines show the values at the lower station and broken lines those at the higher station; in addition to those, the air temperatures measured in the midway are shown by chain lines.

It may be seen in Fig. 3 that the meteorological conditions before February 6 are entirely different from those after that data. Before the 6th there was no inversion; the wind speed was relatively high, and the amount of nocturnal net radiation flux was nearly the same at the two stations. After the 6th a conspicuous inversion was noted, the maximum intensity was clear, the wind was nearly calm, and a net radiation flux was larger at the mountaintop than at the bottom.

The time variations in meteorological conditions

Fig. 2. The location of the meteorological stations of Moshiri.

A: the higher station

B: the lower station

Fig. 3. Variations in mean hourly values of air temperature (A.T.), wind speeds (W.S.) and net radiation fluxes (N.R.) at Toikanbetsu, 1971.

Fig. 4. Variations in mean hourly values of meteorological elements at Moshiri, 1971.

in the snow-melting season at Moshiri are shown in Fig. 4. An inversion also occurred, but its intensity was weak, compared with the intensity in midwinter, the maximum value being about 10°C . This inversion seems to have taken place under the condition of light wind, although this cannot be said definitely, because sufficient data of wind speed was not obtained on the mountaintop.

4. Results of observations

Relations between air temperature and wind speed at Toikanbetsu are expressed in Fig. 5 where circles and crosses show the data at the upper and the lower station, respectively. Sometimes at extremely low temperature, *i.e.* below -30°C , was observed at the low station under the light wind, but at the upper station the temperature was about -13°C at the time. At Toikanbetsu the air temperature atop the mountain was never below -15°C in the past three years. Because the height of the tropopause above Hokkaido in winter is approximately 8 km above sea level and its mean air temperature is -55°C to -65°C , the air temperature at the surface is estimated to be -3°C to -13°C considering the standard lapse rate of $-0.65^{\circ}\text{C}/100\text{ m}$. The air temperature at the higher station roughly agrees with this calculated value. From this result it might be said that the very low temperature at the lower station was not caused by advection of a cold air mass but by the local cooling of air near the surface.

The air temperature at the lower station T_B

Fig. 5. The relation between wind velocity (W) and air temperature (T) at Toikanbetsu, 1971.

showed a larger variation than that at the higher station T_A during the observation period. The difference ΔT between T_B and T_A being considered the intensity of inversion ΔT and T_B are plotted in Fig. 6a. The circle represents the value in midwinter and the cross in the snow-melting season. The temperature difference can be expressed by the linear equation of T_B :

$$T_B = a + 0.8 T_A \dots (a) \text{ in both seasons.}$$

Fig. 6a shows the relation between T_A and T_B . In this figure the line K-K' ($T_B = 2.0 + T_A$) corresponds to the dry adiabatic lapse rate. The temperature inversion is in a region below the line M-M' ($T_B = T_A$) in this paper. Inversion also occurred even if T_A was above 0°C , but in such a condition, the phase change in the snow deposit must be considered and this is a more complex problem. Then, except for the case of T_A being warmer than 0°C , the relation of T_B with T_A can be expressed by the following equation:

$T_B = a' + 5 T_A \dots (b) \text{ in both seasons, as the line B-B' in midwinter and the line C-C' in snow-melting season. Eq. (b) can be derived analytically from eq. (a). From this fact it might be said that, when radiative cooling takes place, the changes of air temperature at the bottom are five times as large as those at the mountaintop. Since this tendency is also noted in the data at Toikanbetsu in the past years (Ishikawa and Ishida, 1971), the results obtained in 1971 are not abnormal phenomena. At Toikanbetsu a remarkable inversion occurred under the condition that the wind speed at the higher station was above } 5 \text{ m}\cdot\text{s}^{-1}, \text{ but the wind speed at the lower station } V_B \text{ was below } 2\sim 3 \text{ m}\cdot\text{s}^{-1}. \text{ Therefore, the relation of the intensity of inversion with } V_B \text{ is shown in Fig. 7. It becomes stronger as } V_B \text{ decreases. In the snow-melting season,}$

Fig. 6a. Variations in the intensity of inversion (ΔT) with the air temperature (T_B) at the lower stations.

Fig. 6b. The relation between air temperatures at the two stations (T_A and T_B).

season ΔT was small, even if the wind speed was nearly equal to that of midwinter.

5. Short-term heat balance at the two stations in midwinter

Heat balance was calculated hourly and then summed up for half-day values (from 6 p.m. to 6 a.m.). The following fluxes can be distinguished:

- A) radiation balance Q_N
- B) sensible heat flux Q_T and latent heat flux Q_e
- C) heat flux in the snow Q_C and heat of snow-melting Q_M

Fig. 7. Variations in the intensity of inversion (ΔT) with the wind velocity (V_B) at the lower stations.

Fluxes toward the surface will be regarded as positive and those fluxes away from the surface as negative. On midwinter nights no snow melting was observed; therefore, Q_M can be assumed to be zero. Over the snow surface where there is no melting, the heat balance is given by the following equation (Munn, 1966):

$$Q_L\downarrow - Q_L\uparrow \pm Q_e \pm Q_C \pm Q_T = 0, \tag{1}$$

Where $Q_L\downarrow$ is downward long-wave radiation from the sky, and $Q_L\uparrow$ upward long-wave radiation from the snow surface.

A) Radiation balance

A snow deposit is concerned to be approximately a black body as far as long-wave radiation is concerned; therefore, $Q_L\uparrow$ is equal to σT^4 , where T is the temperature of the snow surface. The difference of net radiation flux $Q_N = Q_L\downarrow - Q_L\uparrow$ between the two stations was measured for each of the two seasons (Fig. 3 and 4). $Q_L\downarrow$ is not directly measured, but may be evaluated by using the values of Q_N and $Q_L\uparrow$. Table 1 shows these values at the two stations in 1971. In this table $\Delta Q\uparrow$ is difference in outgoing radiation between the two stations ($\Delta Q\uparrow = Q_L'\uparrow - Q_L\uparrow$), and $\Delta Q\downarrow$ is that of incoming radiation ($\Delta Q\downarrow = Q_L'\downarrow - Q_L\downarrow$). $\Delta Q\uparrow$ is much larger than $\Delta Q\downarrow$ during the inversion after February 6 in Table 1, and

Table 1.

	Higher Station			Lower Station			Difference		
Date	$Q_L \uparrow$	Q_N	$Q_L \downarrow$	$Q_L' \uparrow$	Q_N'	$Q_L' \downarrow$	$\Delta Q \uparrow$	$\Delta Q \downarrow$	ΔQ_N
2/2-3	-297.5	-5.1	292.4	-308.0	-8.5	299.5	-10.5	7.1	-3.4
2/3-4	-304.8	-6.3	298.5	-322.8	-5.4	317.4	-18.0	18.9	-0.1
2/4-5	-297.6	-6.0	291.6	-316.8	-7.0	309.8	-19.2	18.2	-1.0
2/5-6	-290.4	-34.8	255.6	-294.0	-16.8	277.2	-3.6	21.6	18.0
2/6-7	-276.0	-42.2	233.8	-235.2	-26.4	208.8	+40.8	-25.0	15.8
2/7-8	-275.5	-74.8	200.7	-210.1	-19.7	190.4	+65.4	-10.3	55.1
2/8-9	-280.8	-75.6	205.2	-222.3	-21.6	200.7	+58.5	-4.5	54.0
2/9-10	-280.2	-74.4	205.8	-220.8	-28.8	192.0	+59.4	-13.8	45.6
2/10-11	-277.2	-54.5	222.7	-225.6	-38.4	187.2	+51.6	-35.5	16.1

on that occasion ΔQ_N becomes larger. Namely, when T_A was higher than T_B' the absolute values of $Q_L \uparrow$ and Q_N became larger at the higher station. This means that the energy loss at the higher station is larger than at the lower station. More detailed measurements were carried out in 1972 and the same tendency was seen.

B) Sensible and latent fluxes

When the temperature and the wind velocity have logarithmic vertical profiles, sensible heat flux Q_T can be calculated by using the following equation (Lettau 1949):

$$Q_T = \rho_a C_p k^2 \frac{T_1 - T_0}{\ln(Z_1/Z_0)} \frac{V_1 - V_0}{\ln(Z_1/Z_0)} \quad (2)$$

where ρ_a is air density, C_p the specific heat of air under constant pressure, k the von Kármán constant, T_1 and V_1 the air temperature and wind velocity at the height of Z_1 respectively, and T_0 and V_0 are at Z_0 . A roughness parameter of 0.1 cm was obtained for our observation period. This value agrees with those by other investigators (e.g. Untersteiner 1958, Wendler and Streten 1969, Holmgren 1971). In calculation the following values were used: $\rho_a = 0.0013 \text{ g} \cdot \text{cm}^{-3}$, $C_p = 0.24 \text{ cal} \cdot \text{g}^{-1} \cdot ^\circ\text{C}^{-1}$, $k = 0.4$, $Z_1 = 100 \text{ cm}$ and $Z_0 = 0.1 \text{ cm}$. T_0 is the temperature at a height of 0.1 cm, obtained from extrapolation of the temperature profile, and $V_0 = 0.0 \text{ m} \cdot \text{s}^{-1}$. A water vapor flux is calculated by using the equation (Lettau 1949, Naruse et al. 1970):

$$E_v = \rho_a k^2 \frac{0.623}{P} \frac{V_1 - V_0}{\ln(Z_1/Z_0)} \frac{e_1 - e_0}{\ln(Z_1/Z_0)} \quad (3)$$

where E_v is the water vapor flux, P the atmospheric pressure, e_1 the water vapor pressure at Z_1 , and e_0 at Z_0 . In calculation, e_0 was assumed to be the saturation vapor pressure over ice at the surface temperature and P was 990 mb at the higher station and 1020 mb at the lower station.

C) Heat flux in the snow and heat balance as a whole

Q_c was not obtained because of insufficient temperature profiles in the snow layer. Therefore, Q_c was estimated as the remainder of heat balance equation (1).

Each term of the heat balance over the two stations is given diagrammatically on a daily basis in Fig. 8. Individual daily fluxes were summed up from 6 p.m. to 6 a.m. Inversions did not occur on January 15 to 16 and 20 to 21; on other days they did. The absolute value of net radiation at the higher station was much larger than that at the lower station except on 20 to 21. In general, Q_T at the higher station is larger than at the lower station. This tendency can be explained from the fact that it is more windy at the higher station. Especially during the period of inversion Q_T was nearly equal to Q_N at the higher station, but Q_T was smaller at the lower station. The latent heat flux Q_e was not so large (0~15% of the heat sink or source) during the period of the observations.

Direct measurements of the amount of condensation (evaporation) were carried out, the maximum value of the latent heat flux being found to be 12%. The percentage-components of both heat sources and sinks over the two

Fig. 8. Dialy values of the components of heat balance over the two stations, Toikanbetsu, January 1972.

Fig. 9. The percentage-components of heat balance during the observation period over the two stations, Toikanbetsu, January 1972.

stations are shown in Fig. 9. Each term was summed up for whole period of the observations. It can be seen that the cooling component at the two stations is only by radiative process. On the other hand, the main heating component at the two stations is a sensible heat flux, and the latent heat flux is of no significance. From these results it can be said that the sensible heat flux always compensated for the energy loss by

Table 2.

Height (cm)	0-5	10	20	50	100
Velocity (cm·s ⁻¹)	60-70	80-90	30	20	10

nocturnal radiation at the higher station but the sensible heat flux was not so large as compared with Q_N at the lower station under the inversion condition.

The stability of the atmosphere was not considered in this calculation. During the period of a strong inversion the atmosphere near the surface should be in a stable stratification (McVehil, 1964), so Q_T of the lower station might be overestimated.

The nocturnal heat balance was not measured in the snow-melting season. Among the studies which have been made hitherto, however, Kojima 1971 measured the heat balance in the daytime. In his paper the sensible heat flux at the higher station was reported to be 1.7 times that at the lower station for 6 days from April 16 to 21, 1971, at Moshiri.

6. Observations of downward airflow on slopes

There are some papers treating of the downward airflow on a slope in a mountainous region under the inversion condition (Bergen 1969, Imaoka 1964). Such an airflow was observed near the mountaintop (150 m above sea level) at Toikanbetsu in 1972 by using smoke and an anemomaster placed consecutively at five different heights (0~250 cm) on the slope. There was a very weak main flow (about 1 m·s⁻¹) upward at a height of 2.5 m above the slope, but near the surface (0~20 cm) there was a downward flow having a speed of 80~90 cm·s⁻¹ (Table 2). The temperature of this airflow (at a height of 10 cm) was not so low (about -13°C). At that time the temperature at the lower station was -25°C, and -12.5°C at the higher station. From this result one may notice that the air temperature at the lower station can hardly drop so low (e.g. below -25°C) even though such a downward airflow is standing within a basin.

7. Discussion remarks

It has been shown that radiation is the only heat sink and the sensible heat flux is the most important source for the heat balance at the surface during the period of nocturnal temperature

inversion in midwinter.

In the snow-melting season, the sensible heat flux was larger at the higher station than at the lower station, and the wind speed and net radiation flux were not so different between the two stations as compared with those in the midwinter; nevertheless, the intensity of inversion was not so strong. Therefore, the difference in the structure of snow deposit might be a cause of the weak inversion in the snow-melting season. Namely, snow layers in the midwinter do not contain liquid water, but in the snow-melting season the water content near the surface in the daytime amounts to as much as about 15~25% of the solid water per unit volume. Some portions of this water freeze owing to nocturnal cooling. It has been observed at Moshiri immediately after sunrise that a surface crust layer (freezing layer) was 20 cm thick near the lower station and 0~2 cm thick at the higher station. Therefore eq. (1) is not available for the calculation of wet snow. The study of heat balance should be pursued in future, considering the phase change (from water to ice) in snow layers.

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厳冬期及び融雪期における盆地冷却の観測

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1971年の冬期と融雪期に、北海道北部及び中央部のいずれも高度差 250 m の盆地状地形の山頂と麓において気象観測を行ない、いかなる条件のもとで盆地冷却が生じるかを調べた。気温逆転の大きさは、冬期の方が融雪期に比べて大であった。

1972年の冬期には、熱収支の計算のために、さらに詳しい観測を行なった。夜間の積雪表面からの熱損失は長波長放射によるものだけであった。一方積雪表面に供給される熱の中で重要なものは顕熱であり、逆転が生じている時、麓と山頂で大きな差がみられた。