SOME FEATURES OF THE TURBULENT TRANSFER ON THE BARE ICE FIELD NEAR THE YAMATO MOUNTAINS, EAST ANTARCTICA

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***Abstract*:** Measurements were made of profiles of wind speed and air temperature on the bare ice surface on the lee of the Yamato Mountains from 1 to 7 December 1973. The vertical profile of air temperature was found unstable in the daytime when the wind speed decreased, as the bare ice surface was heated by solar radiation. Consequently, a mean eddy heat flux transported from the surface was 50 ly/day. As for an energy budget, estimated latent heat flux was about 150 ly/day which indicates the evaporation of 0.25 cm/day in the summer. The observation of wind turbulence by the use of sonic anemometer showed a strong turbulence generated by nunataks. This strong turbulence contributes to the surface ablation of bare ice in the summer.

1. Introduction

Nine pieces of meteorites were discovered in December 1969 by the oversnow glaciology party of the 10th Japanese Antarctic Research Expedition (JARE-10) near the southeastern foot of the Yamato Mountains in East Antarctica (YOSHIDA *et al.*, 1971). In addition to them, 12 meteorites were found in December 1973 in the same area by the party of JARE-14 (SHIRAISHI *et al.*, 1976). From 25 October 1974 to 17 January 1975 the party of JARE-15 was sent to the same area for a systematic search after meteorites, whereby the party collected 663 meteorites (YANAI, 1976). Since the meteorites were found in the bare ice area, KUSUNOKI (1975) pointed out that ablation will aid the exposure of meteorites buried in the ice mass and it may contribute to the concentration of meteorites in this bare ice area. The average net ablation of 5.2 cm/year in terms of ice thickness was obtained between 1969 and 1973 in the area (YOKOYAMA, 1975). Consequently, the energy and mass exchanges between the air and the ice must play an important role in preservation of the bare ice surface.

Micrometeorological observations were carried out on the bare ice near the Yamato Mountains from 1 to 7 December 1973 (KOBAYASHI, 1978b). Fig. 1 shows a location map of the observation site (D0) in the bare ice area on the leeward of Massifs E and F of the Yamato Mountains. The prevailing wind direction near the surface was nearly easterly. The results disclosed that ablation and strong

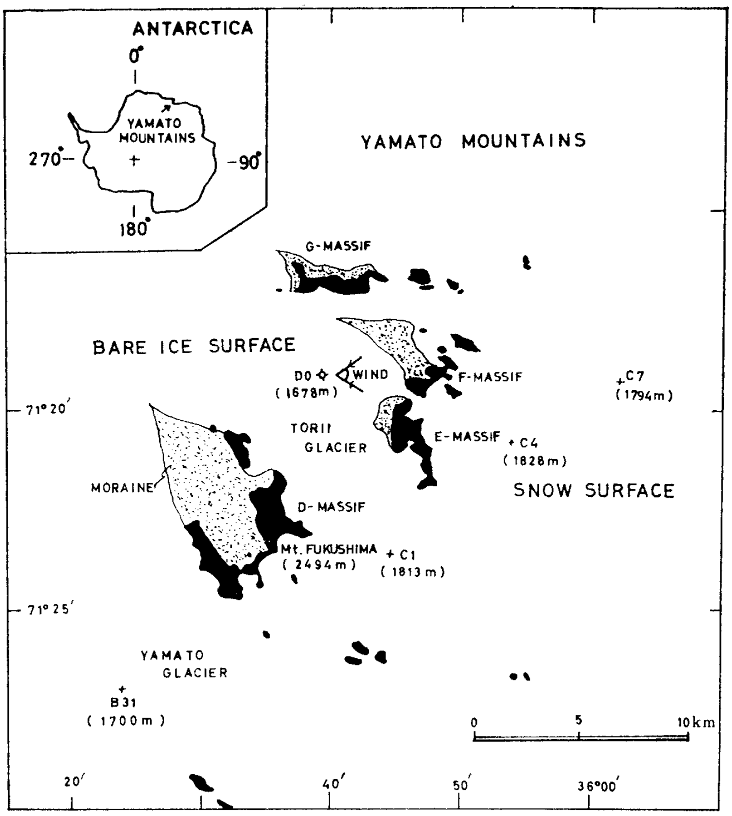

Fig. 1. Location map of observation site (D0) near the northern Yamato Mountains.

turbulence predominated in this bare ice area, particularly in the summer season.

2. Atmospheric Boundary Layer on the Leeward of Nunataks

A wind profile in an atmospheric boundary layer was obtained by means of a radiosonde above the bare ice field on the leeward of the Yamato Mountains. A wind shear hodograph obtained on 12 December 1973 is shown in Fig. 2. It shows that the wind veers clockwise with height, which is different from the results obtained at Mizuho Station on the vasts nowfield where the katabatic wind shifted counterclockwise with height (KOBAYASHI, 1978a). Changes of wind vectors on the snowfield near Mizuho Station are caused by both the Coriolis force and a thermal wind in the inversion layer. On the other hand, the Magnus effect is useful for the explanation of the structure of atmospheric boundary layer on the leeward of nunataks, *i.e.* the vorticity-velocity cross product generates a body force in a strong eddies field. Fig. 3 illustrates the geometry of the Magnus effect (TENNEKES and LUMLEY, 1972). In this case, a stream with a vorticity will be shifted by a body force to the clockwise direction, as shown in Fig. 3. In fact, observations of wind turbulence by the use of a sonic anemometer showed a strong turbulence

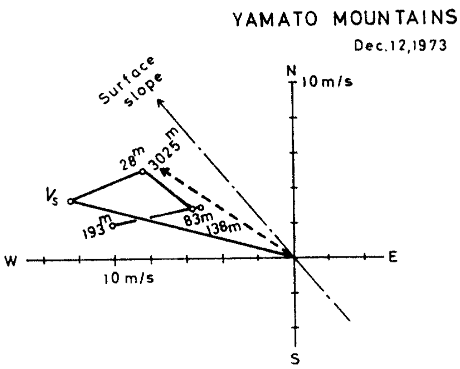

Fig. 2. *Wind shear hodograph in an atmospheric boundary layer obtained on 12 December 1973. The end point of a wind vector veers as function of height. A dashed line shows a wind vector at elevation of 3025 m above bare ice surface. V_S: Surface wind.*

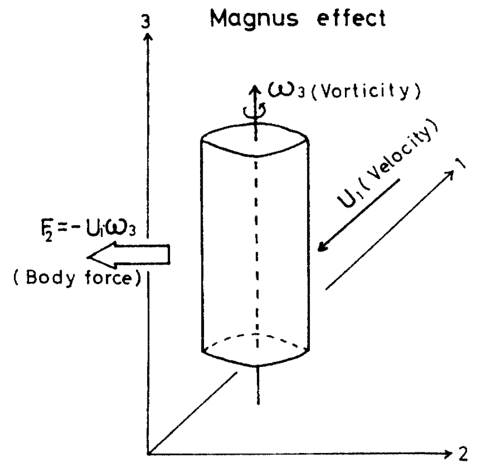

Fig. 3. *Schematic representation of Magnus effect (after* T*ENNEKES and* L*UMLEY, 1972). U_1: velocity in stream; ω_2: vorticity in a predominant eddies field; F_2: body force.*

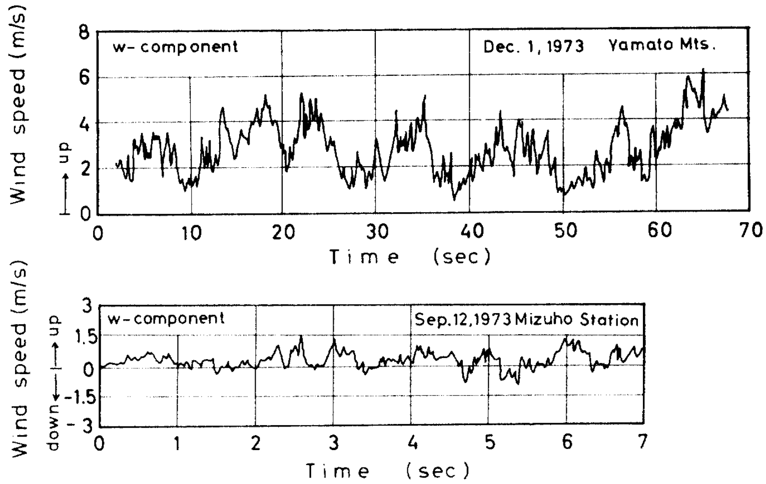

Fig. 4. *Variations of wind turbulence in vertical component on the leeward of the Yamato Mountains and at Mizuho Station.*

generated by nunataks compared with observation at Mizuho Station, as shown in Fig. 4. The vertical component of wind observed in the bare ice area always showed the upward velocity of 3 m/s, while the scale of turbulence at Mizuho Station was small. Although observations at the bare ice field are made in different season from observations at Mizuho Station, many large eddies are found in the field on the leeward of nunataks. These large and strong eddies will contribute to ablation on the ice surface, besides the field being free from snow accumulation by blown snow particles.

3. Atmospheric Surface Layer on the Bare Ice Surface

Vertical profiles of wind speed and air temperature up to 3.5 m were observed at point D0 on the bare ice surface from 1 to 7 December 1973 (see Fig. 1). Five sets of small three-cup electromagnetic anemometers were installed at the heights of 3.35, 1.85, 0.86, 0.55 and 0.35 m. Air temperatures were measured with thermister thermometers at the heights of 3.15, 1.65 and 0.15 m. Some results of data analysis were reported (KOBAYASHI, 1978b), and the present paper deals with energy budget. The atmospheric surface layer on the bare ice surface showed an unstable condition in the daytime when the wind speed decreased, because the bare ice surface was heated by solar radiation. In an adiabatic atmosphere (neutral condition), the wind profile is represented by the following logarithmic equation:

$$U(z)=(U_*/\kappa)\ln Z/Z_0, \tag{1}$$

where U_* is the friction velocity, $\kappa(=0.4)$ is von Karman's constant and Z_0 is the surface roughness length. However, when buoyancy forces are present, *i.e.* for a diabatic case, it is not possible to present a universal form for the diabatic wind profile. In an unstable condition the wind speed must decrease with height. However, since buoyancy forces near the surface are controlled by the surface, a wind profile can approximate to the logarithmic law in a neutral condition.

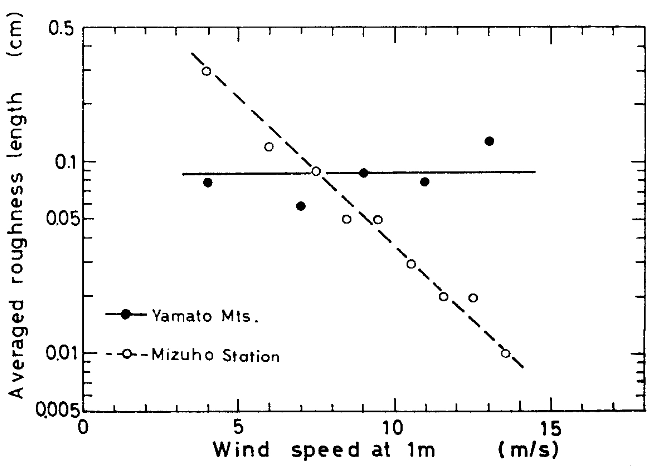

Fig. 5. Relations between averaged roughness length and wind speed at 1 m height on the bare ice (solid line) and snow surface (dashed line).

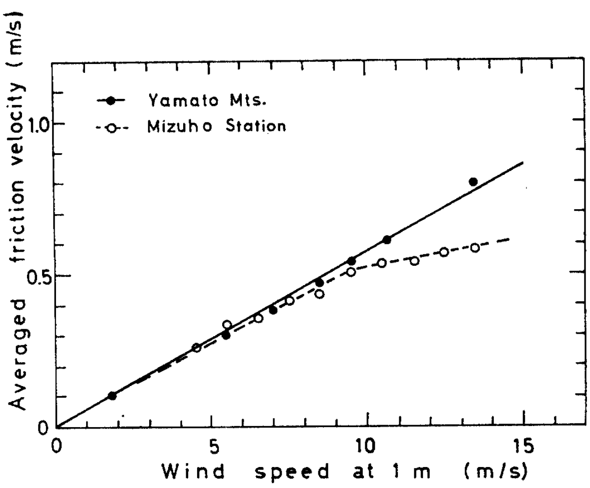

Fig. 6. Relations between averaged friction velocity and wind speed at 1 m height on the bare ice (solid line) and snow surface (dashed line).

Fig. 5 shows the averaged roughness lengths obtained on the bare ice surface near the Yamato Mountains and the snow surface at Mizuho Station in the region of cold katabatic flow. The former is a constant value of 0.085 cm, which is about 1/30 of the depth of small concaves (2–3 cm) on the bare ice surface (Nicruse's result; see MONIN and YAGOLOM, 1971). As for the latter, the roughness length decreases conspicuously with the increase in wind speed. Relations between averaged friction velocity and wind speed at 1 m in height in both areas are shown in Fig. 6. The friction velocity obtained on the bare ice surface near the Yamato Mountains was larger than that on the snow surface at Mizuho Station when the wind speed was more than 10 m/s.

4. Eddy Heat Flux on the Bare Ice Surface

An eddy heat flux can be estimated from friction velocity and vertical differences in air temperature and wind speed in the lowest layers. The aerodynamic method satisfies an assumption that eddy transfer coefficients for momentum, heat and water vapor are the same, that is, $r_m = r_h = r_e$. Then, the equation for an eddy heat flux may be written in an expression analogous to a molecular transfer, yielding the flux of heat in the form

$$H = \rho C_p [T(Z_1) - T(Z_2)]/r_h, \tag{2}$$

where ρ (=1.2 kg·m^{-3}) is the air density, C_p(=0.24 kcal·kg^{-1}·deg^{-1}) is the specific heat of air, $T(Z_1)$ and $T(Z_2)$ are the air temperatures at the heights of Z_1 and Z_2 respectively, and r_h is the drag coefficient. The drag coefficient is expressed by

$$r_h(Z_1, Z_2)=[U(Z_2)-U(Z_1)]/U_*^2, \tag{3}$$

where $U(Z_1)$ and $U(Z_2)$ are wind speeds at the heights of Z_1 and Z_2, respectively.

Eq. (2) represents an eddy heat flux between the heights of Z_1 and Z_2, and the positive sign of flux indicates its transport away from the interface.

Next, the surface temperature of the bare ice surface can be approximately

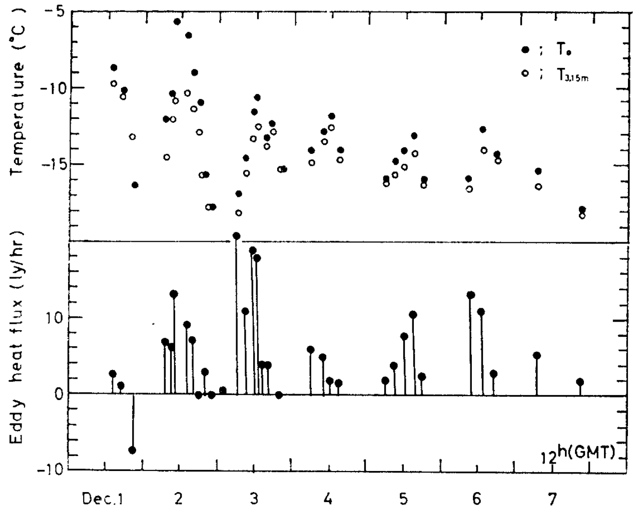

Fig. 7. Variations in temperature at surface and 3.15 m height and eddy heat fluxes on the bare ice surface.

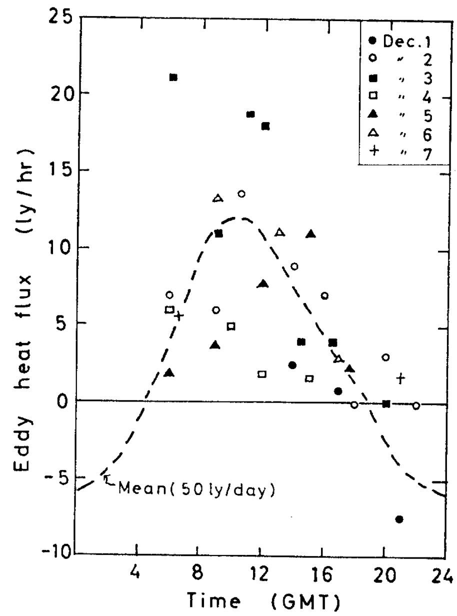

Fig. 8. Hourly variation in eddy heat fluxes on the bare ice surface. A dashed curve shows an averaged heat flux.

calculated by the following equation:

$$T_0=[r_h(Z_1)\cdot H/\rho C_p]+T(Z_1), \tag{4}$$

where the drag coefficient between the height of Z_1 and the surface is represented by

$$r_h(Z_1)=U(Z_1)/U_*^2. \tag{5}$$

Fig. 7 gives variations from 1 to 7 December 1973 in temperature at the surface and at 3.15 m and in eddy heat flux. Since the measurements of wind speed and air temperature were not made continuously, the figure lacks nighttime data. Eddy heat flux may have negative values at night. Fig. 8 illustrates hourly variations in eddy heat flux over the bare ice surface and also includes the averaged flux as shown by a dashed curve. It shows that the total flux obtained by integrating the dashed curve was about 50 ly/day.

5. Energy Budget at the Surface

Although observations were not made on the energy budget at the bare ice surface, a rough estimation of latent heat flux will be made. The energy budget at the ice-air interface has been determined from the following equation:

$$R_0=H_0+S_0+E_0, \tag{6}$$

where R_0 is the net radiation, H_0 is the eddy heat flux, S_0 is the snow heat flux computed from snow temperature curves at the surface on the assumption that no annual mean heat flux exists in the snow, and E_0 is the latent heat flux. The radiation balance at the surface may be expressed by

$$R_0=R_S+R_L=S_G(1-\alpha)+L_d+L_u, \tag{7}$$

where R_S is the absorbed radiation (shortwave balance), R_L is the effective radiation (long wave balance), S_G is the global radiation, α is the albedo of the ice surface for global radiation, L_d is the downward long wave radiation from the atmosphere and L_u is the upward long wave radiation from the surface. KAWAGUCHI and SASAKI (1975) reported that the value of the global radiation obtained at Mizuho Station (70°41.9′S, 44°19.9′E; elevation 2230 m) was 950 ly/day when the sky was clear around the summer solstice. According to KAWAGUCHI and FUJII (1977), the albedo of the bare ice surface in the vicinity of the Yamato Mountains was 0.5. Hence, the absorbed radiation was nearly equal to 475 ly/day. Then, the upward long wave flux at the surface may be calculated by

$$L_u=\sigma T_0^4, \tag{8}$$

where σ ($=0.8\times10^{-10}$ ly·min^{-1}·deg^{-1}) is Boltzman's constant. Since the surface temperature computed from eq. (4) had a mean value of 260 K during the period of observation, the upward long wave radiation at the surface was 530 ly/day. On the other hand, for the downward long wave radiation from the atmosphere: KONDO (1971) proposed the following empirical formula

$$L_d=\sigma T_h^4[1-(0.49-0.066\sqrt{e})C_L]-\Delta I_L, \tag{9}$$

where T_h and e are the daily mean value of absolute air temperature and the daily mean value of vapor pressure near the surface, respectively, C_L is the correction factor due to the cloud for the long wave radiation and ΔI_L is the correction of the downward long wave radiation due to the inversion. In this paper, the values at the top of a surface inversion layer are used for T_h and e, that is, $T_h=262$ K, $e=2.38$ mb, $C_L=1$ (fine weather) and $\Delta I_L=0$. From the result of calculation by eq. (9), the downward long wave radiation was about 330 ly/day. Therefore, the net radiation estimated at the bare ice surface was about 275 ly/day, *i.e.* the bare ice surface may be heated by the fraction of this net radiation. Since the mean value of eddy heat flux is 50 ly/day, the sum of snow heat flux and latent heat flux will be 225 ly/day. Although no data are available of ice temperature profiles in the bare ice layer, the snow heat flux is smaller than the latent heat flux. Acccordingly, the latent heat flux may be of the order of 150 ly/day, which is equivalent to the ablation rate of 0.25 cm/day.

6. Concluding Remarks

The rough ablation rate estimated in the above section may be larger at the ice-air interface than at the snow-air interface, because the albedo of the snow surface (*e.g.* 0.8 by KAWAGUCHI and SASAKI, 1975) in the vicinity of Mizuho Station is larger than that of the bare ice surface. Thus, the net radiation estimated at the snow surface was roughly -10 ly/day. This negative sign may be effective in bringing about a radiative cooling at the surface. According to NISHIO (1978), the average latent heat flux calculated by water vapor measurements at Mizuho Station in the summer season was 54 ly/day (evaporation: 0.2 cm/day) which is comparable to the value of 64 ly/day for the sum of snow heat flux and eddy heat flux.

Eddy and latent heat fluxes at continental stations in the summer season are summarized in Table 1. The ablation rate estimated on the bare ice is three or more times as greater than those measured at other stations covered with the snow surface. Net ablation observed in the bare ice area near the Yamato Mountains averages 5.2 cm/year (YOKOYAMA, 1975). Accordingly, since the ablation on the bare ice in summer was about 7.5 cm/month, in other seasons negative latent heat fluxes (condensation) may be predominant. Pursuance of this study calls for a prerequisite of making measurements of energy budgets at the

Table 1. Eddy heat flux and latent heat flux at continental stations in summer season, Antarctica.

Station	Season	Eddy heat flux (ly/day)	Latent heat flux (ly/day)	Surface	References
Pionerskaya	Jan. 1956 Jan. 1958	13.3	−3.9	Snow	RUSIN (1960)
Komsomolskaya	Jan. 1958	10	—	Snow	RUSIN (1960)
Vostok	Jan. 1958	16.7	—	Snow	RUSIN (1960)
Mizuho Station	Dec. 1976 ~ Jan. 1977	−30∼−50**	53.5 22.5 (0.75 mm/day)*	Glazed snow	NISHIO (1978)
Mizuho Plateau	Nov. 1969 ~ Jan. 1970	—	28 (0.9 mm/day)*	Hard sastrugi	AGETA (1972)
Yamato Mountains	Dec. 1973	50	100∼200**	Bare ice	KOBAYASHI

Pionerskaya (69°44′S, 95°30′E, elevation 2700 m)
Komsomolskaya (74°05′S, 97°29′E, elevation 3420 m)
Vostok (78°27′S, 106°52′E, elevation 3420 m)
* Value estimated by stake method.
** Estimated values by KOBAYASHI.

bare ice surface throughout the year.

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