VERTICAL STRUCTURE OF KATABATIC WINDS IN MIZUHO PLATEAU

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Abstract: Observation of vertical structure of katabatic winds blowing in Mizuho Plateau was made by means of radio sondes. It was found that the layer of the katabatic winds is roughly identical with the surface inversion layer. The surface inversion layer showed seasonal variations in thickness and intensity. The katabatic wind layer disappeared when a cyclone was passing over the ocean in the vicinity of Syowa Station, as noticed from the disarranged wind speed profiles.

Wind vectors in the lowest few hundred metres of the katabatic wind layer showed that a thermal wind persisted in the surface inversion layer. Therefore, the katabatic winds in Mizuho Plateau may be called the "inversion winds", because they are essentially controlled by the thermal wind due to the existence of a sloped inversion layer.

1. Introduction

Katabatic winds constitute a predominant phenomenon on the Antarctic coastal slope. It is the movement of cold air down along the slope by gravity, after the radiative cooling of the ice sheet at its surface.

Studies have been made of katabatic winds in Mizuho Plateau and Syowa Station and their vicinities by a number of investigators. Morita (1968), who described some characteristics of katabatic winds at Syowa Station, reported that their effect reaches Syowa Station only when the katabatic flow is of "type 2" in Ball's theory (Ball, 1956, 1960) in which a jump can occur inland. The flow patterns of the katabatic winds are well known around Mizuho Plateau; *i.e.*, Ageta (1971) and Watanabe (1978) deduced the flow lines of katabatic winds in this area from the orientations of sastrugi and snow dunes. Katabatic winds have a diurnal variation during the summer season. Yamada (1974) reported that the katabatic wind speed observed at Mizuho Camp in the summer season showed the minimum at 1800 LT and the maximum at 0600 LT both once a day, which corresponded respectively to the maximum and minimum of air temperature observed at 1500 LT and 0300 LT with 3 hours time lag. The annual mean speed of katabatic winds at Mizuho Camp is as high as 10–11 m/s (Sasaki, 1974; Inoue *et al.*, 1978). Kobayashi and Yokoyama (1976) measured vertical profiles of temperature as well as wind speed and direction by using radio sondes and pilot balloons in Mizuho Plateau in 1973. Analysing meteorological data at Mizuho Camp and Syowa Station, INOUE *et al.* (1978) showed that Mizuho Camp is located not only in a cold katabatic wind region (DALRYMPLE, 1966), but also under the influence of a synoptic scale disturbance near Syowa Station.

The katabatic wind layer is always accompanied by a surface inversion layer. It is the purpose of this paper to present the vertical profiles of temperature in the surface inversion layer and vertical profiles of wind speed and direction in the katabatic wind layer in Mizuho Plateau.

2. Temperature Profiles in the Katabatic Wind Layer in Mizuho Plateau

Four representative examples of profiles of surface temperature inversion under a clear-sky condition are shown in Fig. 1. Examples (1), (2) and (3) show the profiles at Mizuho Camp, and (4) at C98. This surface inversion above the ice sheet is closely connected with katabatic winds; *i.e.*, the katabatic wind layer roughly coincides with the surface inversion layer. The ordinate in this figure shows the height from the ice sheet at Mizuho Camp and C98, where the elevation was 2230 and 1824 m above sea level, respectively.

It is found from Fig. 1 that the thickness of the surface inversion layer is 600 m in winter, which is much thicker than 250 m in autumn and summer, and that the intensity of inversion (the difference in temperature between the top and bottom of the layer) in winter is 15–20°C which is much greater than 5°C in

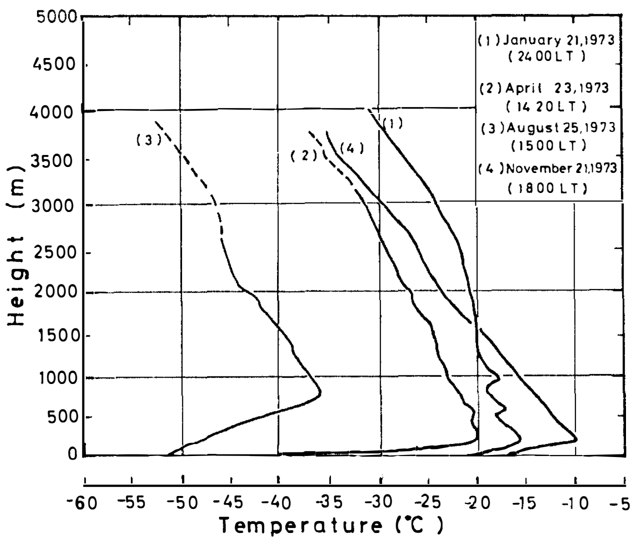

Fig. 1. Distribution of temperature with height from the ice sheet during katabatic winds.

summer. The largest intensity of surface inversion was observed in April, namely about 20°C, while the inversion layer recorded the maximum thickness of about 600 m in August.

MAHRT and SCHWERDTFEGER (1970) approximately simulated the vertical temperature profile above the snow surface on the Antarctic Plateau by an exponential equation

$$T(z)=T_h-\varDelta T\cdot\exp\left(-z/\sqrt{K/f}\right), \qquad (1)$$

where $T(z)$ is the temperature at height z from the snow surface, T_h is the temperature at the top of surface inversion layer, $\varDelta T$ is intensity of inversion, K is mean coefficient of eddy diffusivity and f is the Coriolis parameter.

Using the data of the temperature profiles in surface inversion layer, the values of K estimated by eq. (1) were in the range between 10^3 and $10^5\,cm^2\cdot s^{-1}$ as listed in Table 1.

Table 1. Mean coefficient of eddy diffusivity and speed of the thermal wind in surface inversion layer above the sloped terrain.

Run	1	2	3	4
Date	Jan. 21, 1973	Apr. 23, 1973	Aug. 25, 1973	Nov. 21, 1973
Local time	2400	1420	1500	1800
Temperature at surface, T_s (°C)	−21	−40	−52	−16.5
Temperature at top of inversion, T_h (°C)	−16	−20	−36	−10
Strength of inversion, $\varDelta T$ (°C)	5	20	16	6.5
Mean temperature in inversion layer, T_m (°C)	−18.5	−30	−44	−13.3
Eddy diffusivity, K (cm²/s)	1.0×10^4	5.7×10^3	5.0×10^5	2.0×10^5
Thermal wind, $\varDelta G$ (m/s)	4	16	14	5

The presence of a cold atmospheric layer of approximately constant thickness above sloped terrain implies the existence of a horizontal temperature gradient through the surface inversion layer. Such a horizontal temperature gradient causes a thermal wind, blowing parallel to the contour lines of the sloped terrain, with the high ground to the right hand side (DALRYMPLE *et al.*, 1966; DALRYMPLE, 1966; SCHWERDTFEGER and MAHRT, 1968; MAHRT and SCHWERDTFEGER, 1970). According to DALRYMPLE *et al.* (1966) the relationship between the slope of the terrain S and the thermal wind $\varDelta G$ (the difference in geostrophic wind speed between the top $\vec{G}_h$ and bottom $\vec{G}_0$ of the inversion layer) is represented by the equation

$$\varDelta G=\vec{G}_h-\vec{G}_0=\frac{g\cdot(T_h-T_s)\cdot S}{f\cdot T_m} \qquad (2)$$

where g is the acceleration of gravity, T_s is temperature at the height 10 m above surface and T_m is mean temperature in a surface inversion layer.

Therefore, the magnitude of the thermal wind as a function of the strength of the inversion and the slope of the ground is determined by eq. (2). The slope of the ice sheet surface in a region of 2100–3100 m in elevation of Mizuho Plateau is in the range of $(2.2–3.8)\times10^{-3}$, averaging 3×10^{-3} (SHIMIZU *et al.*, 1978). The value of Coriolis parameter is $1.4\times10^{-4}\mathrm{s}^{-1}$ at Mizuho Camp (70°41′53″S, 44°19′54″E, 2230 m). Using these values of f and S, the speed of the thermal wind estimated by eq. (2) was listed in Table 1. Compared with the thermal winds in January and November, those in April and August are quite strong.

As an example of measurement by a pilot balloon of the height of the bottom of an altostratus, the result of an observation on 12 September 1973 showed the height of about 2800 m above sea level, or 570 m above the ice sheet at Mizuho Camp. In this case it was clear that the bottom of the cloud height coincided with the top of the surface inversion layer. This height also coincided with the height of the bottom of an altocumulus that was within a range from 2 to 3 km according to the result of a stereoscopic observation by two fisheye lens cameras in the summer season at Syowa Station (KIKUCHI *et al.*, 1976). The thickness of a surface inversion layer at Syowa Station was 350 to 400 m in winter and 250 to 300 m in summer (MAKI, 1974). Therefore, the height of the bottom of the cloud at Syowa Station did not coincide with the top of the surface inversion layer.

3. Wind Structure in the Katabatic Wind Layer

Six examples given in Figs. 2–4 show wind shear hodograph in the katabatic wind layer. Figs. 2, 3 and 4 show the representative examples in the autumn,

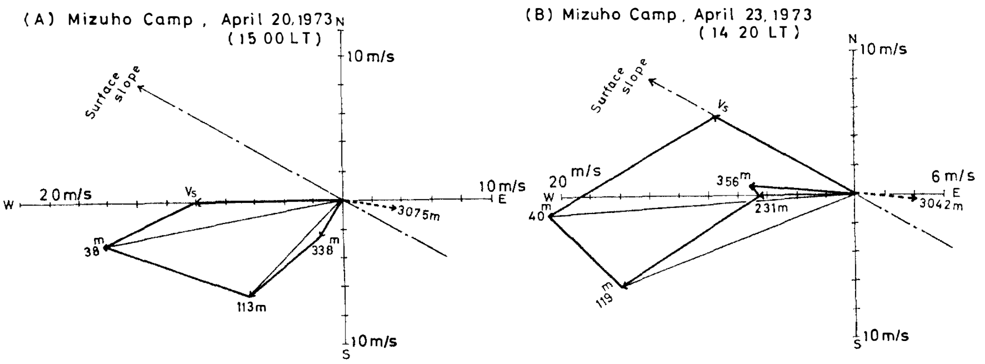

Fig. 2. Wind speed distribution in a katabatic wind layer in autumn. Numbers in metres are elevations above the ice surface. The end point of a wind vectors veers on the hodograph as function of height. V_S: Surface wind.

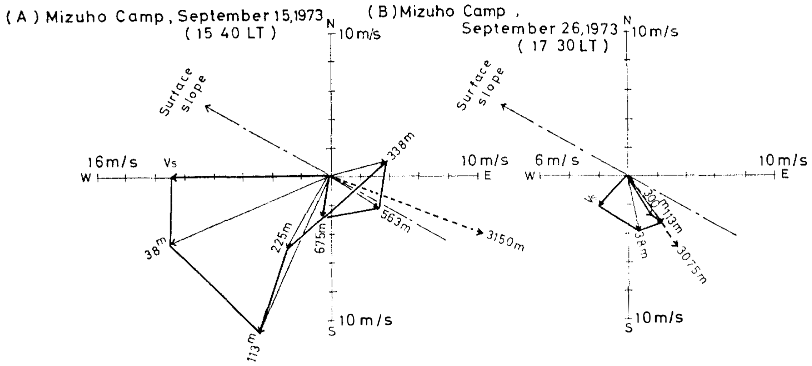

Fig. 3. Wind speed distribution in a katabatic wind layer in winter. Numbers in metres are elevations above the ice surface. V_s: Surface wind.

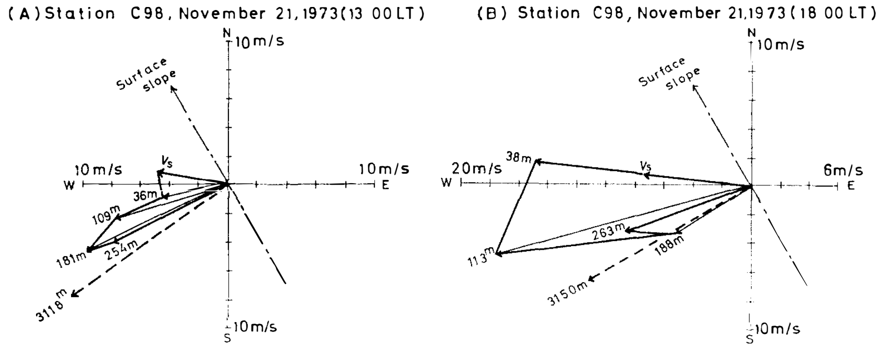

Fig. 4. Wind speed distribution in a katabatic wind layer in summer. Numbers in metres are elevations above the ice surface. V_s: Surface wind.

winter and summer seasons, respectively. As shown in these figures, the deviation of some surface wind (V_s) from the line of greatest slope caused by Coriolis' force showed a tendency of the counterclockwise direction. The speed of wind increased with height up to 40 to 100 m then decreased as the height approached the top of the katabatic wind layer, *i.e.*, the katabatic winds showed the maximum speed of 15 to 20 m/s between the heights of 40 and 100 m and the minimum speed at the height which coincided with the top of the surface inversion layer. Fig. 2 presents two examples obtained in April, in which the strongest wind is shown at a lower height in the katabatic layer, whereas a wind at a height of 3000 m above the ice sheet is relatively weak. The wind direction in the upper portion of the layer is roughly westerly with the southward component which is considered to be the compensating current against a surface wind which is easterly, flowing outward to the periphery of the continent.

Fig. 3(A) gives a typical example observed in September. Compared with the winds in the katabatic layer in April, those in September are not so strong any more, but an upper westerly wind, the compensating wind becomes stronger than that in April, because a meridional air circulation is enhanced in winter. Fig. 3(B) represents an example of the destruction of the katabatic wind layer by a strong turbulent wind caused by a cyclone passing over the ocean in the vicinity of Syowa Station. Namely, the shape of wind speed profiles of the katabatic wind layer similar to the shape of those of a wall jet is extinguished during a short period of about 15 hours, and the direction of the surface wind becomes northerly from easterly.

Fig. 4 gives two examples of katabatic winds on 21 November 1973, in which (A) shows vectors of a gentle wind observed at 1300 LT and (B) those of a strong wind developed five hours later at 1800 LT. This strong katabatic wind in summer has nearly the same structure of wind profiles as the katabatic wind in autumn, but the surface wind speed is much weaker in summer than in autumn. The direction of an upper wind in summer approaches toward the direction of a surface wind.

The wind vectors in the inversion layer as presented in these figures show that a thermal wind persists in the inversion layer. Winds veer with height; *i.e.*, they shift counterclockwise in the southern hemisphere. The angle between the wind vectors at the surface of the ice sheet and the top of the inversion layer lies in the range of 40 to 70 degrees.

The thermal wind also flows along the isotherms with cold air to the right in the sloped inversion layer; *i.e.*, the thermal wind in East Antarctic Plateau tends to be directed parallel to the contour lines of the terrain with the higher ground to the right hand side (SCHWERDTFEGER and MAHRT, 1968). It is assumed that the average winds are geostropic winds (G_h) at the top of the inversion layer

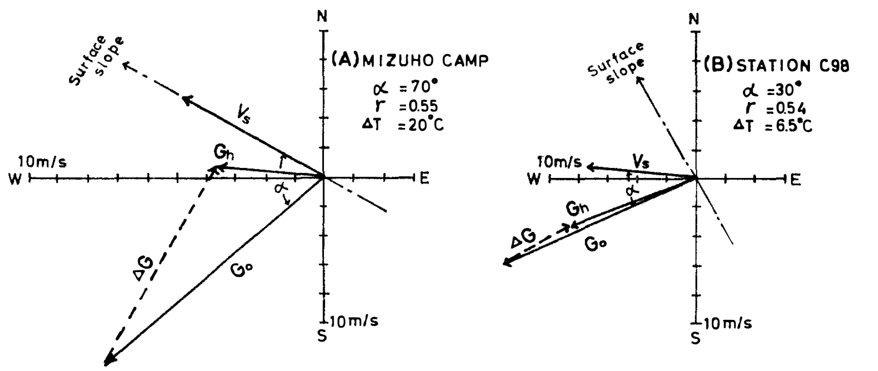

Fig. 5. Difference between the actual surface wind (V_s) and the geostrophic surface wind (G_0) depending on the strength of the inversion ΔT.
ΔG: Thermal wind in sloped inversion layer.
G_h: Wind vector at the top of the inversion layer.
α: Deflection angle between V_s and G_0.
r: V_s/G_0.

and the thermal winds (ΔG) calculated by eq. (2) as shown in Table 1 tends to be parallel to the contour lines of the terrain. Therefore, using the data of G_h and ΔG, the difference between the actual surface winds (V_s) and the geostrophic surface winds (G_0) was set out schematically as shown in Fig. 5, α represents the deflection angle between the actual surface wind and the geostrophic surface wind, and r shows the ratio of the actual surface wind to the geostrophic surface wind. This difference between V_s and G_0 is due to friction which depends on the intensity of the inversion and surface roughness; *i.e.*, the difference increases with increasing strength of the inversion ΔT as shown in Fig. 5.

4. Concluding Remarks

The study of the structure on katabatic winds at other Antarctic stations have been made since IGY by many investigators such as BALL (1956, 1960), SHAW (1960), TAUBER (1960), STRETEN (1963, 1968), LETTAU (1966) and DALRYMPLE (1966). But, observations by means of radio sondes in Mizuho Plateau were first made by the wintering party of the 14th Japanese Antarctic Research Expedition in 1973. The results obtained were limited only to a few examples. Under an Antarctic anticyclonic condition, the katabatic winds are almost in a steady state and the clear sky always dominates the weather. Namely, a thin cold anticyclone (the cold katabatic flow) exists at low levels on the sloped ice sheet surface, where the surface winds blow outward to the periphery of the ice sheet. The compensating current blows at high levels toward the inland of the ice sheet. This compensating current is likely to change to a sinking motion at a plateau higher than 3000 m above sea level and is connected to a thin anticyclone. This sinking motion forms a thick anticyclone so that the sky is always clear at such a high plateau. The results obtained in Mizuho Plateau support the atmospheric circulation over Antarctica in winter reported by WHITE and BRYSON (1967).

According to SHIMIZU *et al.* (1978), the surface slope in Mizuho Plateau is $(0.51–1.7)\times10^{-3}$ in the area higher than 3100 m, and $(2.2–3.8)\times10^{-3}$ between 2100 m and 3100 m. BALL's (1960) monogram for determining the katabatic force in terms of inversion strength and surface slope shows that the katabatic force exceeds the pressure gradient force when the surface slope exceeds 2×10^{-3}. Therefore, it is expected that the katabatic slope begins from around an altitude of 3100 m in Mizuho Plateau. The katabatic winds in Mizuho Plateau may be called "inversion winds" as discussed by SCHWERDTFEGER (1970), because they are essentially controlled by the thermal wind due to the existence of a sloped inversion layer.

On the other hand, the influences of a cyclone on the structure of katabatic winds are important in relation to snow accumulation and to exchange of energy between the ice sheet and the atmosphere. More extensive researches are called for to measure the katabatic winds by means of radio sondes under a cyclonic weather condition.

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