Gravity Wind on a Snow Patch

By Tetsuo Ohata* and Keiji Higuchi*

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Abstract

An occurrence of a gravity wind was found on a snow patch of smaller than 1.2 km in length, on a warm summer day. The fundamental properties of this wind obtained from the observation are as follows. The maximum wind speed (U_max) was 1.5 to 4.0 m/sec., and the height of the maximum wind speed (Z_max) was between 0.7 to 1.2 m above snow surface. The Z_max and U_max had a linear relation. The wind speed of the gravity wind at a certain height, which can be considered to be proportionate to U_max, was stronger when the difference of the air temperature between the free air and the cooled air layer above the snow surface was larger. The wind speed of the gravity wind was stronger at lower parts of the snow patch, having a relation of U * l where U is the wind speed and l is the distance from the upper end of the snow patch. The occurrence of the gravity wind was controlled by the strength of the general wind in the valley. It occurred only when the general wind was weak.

The vertical profile of the observed wind speed can not be explained by the classical theoretical profile obtained by Prandtl (1952), but can be explained by the empirical profile obtained by Martin (1975).

1. Introduction

There are local winds which occur as a result of an air temperature difference, that is, density difference between the air mass above an inclined surface and the free air around. One well known type of this wind is the mountain-valley wind. Another type of wind is the "Glacier Wind" which has been long known in the European Alps (Tollner, 1931; Hoinkes, 1954a, 1954b; Businger and Rao, 1965; Martin, 1975). This glacier wind is a type of katabatic wind which is strongest in the mid afternoon.

We studied the possibility of a wind system similar to the glacier wind on the snow fields and snow patches in the mountain regions of Japan, from the view point of snow mass-atmosphere interaction. In the high mountain ranges of Japan, snow cover remains until late in summer, and in some places perennial snow patches are formed (Higuchi and Iozawa, 1971). The surface temperatures of the snow patches do not exceed 0°C, but in those mountain ranges the daytime air temperature rises up to high as 10 to 15°C in the summer season. Under such conditions, the air layer near the snow surface is cooled considerably by snow, and there develops a strong temperature difference between the air layer near the snow surface and the surrounding air. This situation will induce a downslope gravity wind similar to the glacier wind. This type of wind was found in observations made on a snow patch at Tsurugisawa, Tateyama Region, Northern Japan Alps in the summer of 1975 and 1976. The results of these observations and some discussion on the characteristics of these winds will be described in this paper.

As this is the first report to show the existence of this type of gravity wind on snow fields and snow patches, a technical term for this wind should be established. In Japanese the term "SEKKEI-KAZE" meaning "snow patch wind" is being used, but for English an appropriate term has not yet been found. In this paper, gravity wind, or gravity wind on snow patch or snow field will be used. Other technical terms used to refer to this type of wind are katabatic wind, cold air drainage and drainage wind.

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Fig. 1a Map of the observational area. The black part of the small map is the position of the large map. The observational area is marked by an oblique line.

Fig. 1b Photograph of the valley, showing the snow cover at the beginning of July. This photograph was taken on July 19, 1968.

2. Observations

Observations were made on a snow field in the mountain ranges in Central Japan. The site was located in a cirque facing north, south of Mt. Tsurugi. The map of the area and a general view of the valley are shown in Figs. 1a and 1b. The period and item of observations, the place and the size of the snow field (snow patch) are shown in Table 1. The situation of the snow field (snow patch) is shown in Figs. 2a and 2b.

The snow cover in periods 1 and 3 should be called snow field instead of snow patch, but to eliminate confusion it will all be called snow patch in this paper.

Observations on periods 1 and 3 were made on the same snow patch L in Fig. 2a. The size of the snow patch L in Fig. 2a corresponds to period 1, but it was nearly the same in period 3. Observation in period 2 were made in September, two month later than period 1. At this time the snow patch L had melted away and the observations were made on the smaller snow patch R in Fig. 2b.

Points where the measurements were made are shown in Figs. 2a and 2b. Points A, C, D and E were on the snow patch, point B on the bare ground. In period 1, point C was used; in period 2, point A and B; in period 3, points C, D and E.

The size and the inclination of the snow patch is shown in Table 1.

The observed quantities are shown in Table 1. Mainly, the vertical profile of the wind speed and air temperature, and the wind direction were observed. In order to study the areal distribution of wind in period 3, wind at only one height was observed. Instruments used for the observation
Table 1. Periods, items and places of the observations. L, R is shown in Fig. 2.

<table>
<thead>
<tr>
<th>Period</th>
<th>Observation items</th>
<th>Place</th>
<th>Features of the snow patch</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>vertical profile of wind and air temperature</td>
<td>snow patch L point C</td>
<td>Length: 1.2 km, Width: 250 m, Inclination: upper part 15°, lower part 5°</td>
</tr>
<tr>
<td>2</td>
<td>same as above</td>
<td>snow patch R point A bare ground point B</td>
<td>Length: 0.25 km, Width: 60 m, Inclination: 7 to 10°</td>
</tr>
<tr>
<td>3</td>
<td>wind speed observation at 3 point</td>
<td>snow patch L point C, D, E</td>
<td>same as period 1</td>
</tr>
</tbody>
</table>

Fig. 2a The situation of the snow field and the observation points for period 1 and 3. The place with no contour line is the snow cover. The position of the snow patch in period 2 is shown by a broken line.

were a photo-electric cup type anemometer for wind speed, and a platinum resistance thermometer placed in a shelter for air temperature. The observations of wind direction were made by flag in period 1 and 2, and by the use of a photo-electric wind vane in period 3.

The averaging time for the observed meteorological elements were 10 minutes, if not specially cited.

3. Observational results

3.1 Basic properties of the wind

3.1.1 Vertical profile of the wind speed and air temperature

A typical profile of the wind speed and air temperature for a gravity wind on snow patches is shown in Fig. 3. Profile (a) was observed in period 1 and (b) in period 2. Both have the characteristics of the gravity wind, which has the maximum wind speed near the surface. The height of the maximum wind speed is 0.8 m in the case of (a) and 1.1 m in (b). This height is very low in comparison with those of other types of katabatic wind, such as the katabatic wind in the interior of Antarctica (Lettau, 1966), glacier wind (Hoinkes, 1954a), and mountain winds or cold air drainage (Defant, 1949; Imaoka, 1964). The case of (c) shows the profile of a general
Fig. 3 The vertical profile of wind and air temperature during the occurrence of the gravity wind on snow patches ((a) and (b)). For comparison, the profile of the general wind is shown (c).

wind (not gravity wind), actually the lowest layer of the valley wind which blew on to the snow patch. This wind shows a profile similar to the logarithmic one.

As seen in Fig. 3, the air temperature profile of the gravity wind was quite different in comparison with that of the general wind. The air temperature at 4.0 m level was about the same for (a) and (c) in Fig. 3, but the air temperature profile for the gravity wind (a) has a larger air temperature gradient, approximately 3 times that of the general wind between 0.4 m and 4.0 m levels. The strong temperature gradient is the cause of the occurrence of this type of wind.

Another interesting feature is the existence of a "thermocline", a layer of strong temperature gradient which exists around the height of the maximum wind speed. This appears to indicate the existence of two independent air layers, namely, a cool air layer underneath a warm layer. But this thermocline becomes obscure as the gravity wind gets stronger. The existence of a thermocline in a gravity wind is reported in an observation on an ice cap by Holmgren (1971).

Since it changes the turbulent structure of the wind, the thermocline will have an effect in reducing the vertical eddy diffusion between layers, and will have an important role in the development of the gravity wind.

The gravity wind on a snow patch will be restricted to gravity winds which has the maximum wind speed at the height lower than 2 meters. This restriction is needed to discriminate the wind that originated on the snow patch which is under discussion in this paper, and other types of gravity wind such as mountain wind, which blows in the whole valley.

3.1.2 Steadiness of wind direction
Another characteristics of this gravity wind is the steadiness of the wind direction. Such a tendency can be seen in Table 2, which shows the mean wind direction and its standard deviation at the height of 1.0 meter in the gravity wind layer. It was observed every 5 sec. for the period of 10 minutes. For comparison, Table 2 shows also the case of a general wind with almost constant wind direction and wind speed in the same range of 3.0 to 3.5 m/sec. This range of wind speed was chosen, for the strongest gravity wind occurred in this range.

As seen in Table 2, the fluctuation of the wind direction of the gravity wind is one order of magnitude smaller than that of the general wind. This can be explained as follows. Since this gravity wind is confined to the lowest few meters, at most 10 meters of the atmosphere, and is driven

<table>
<thead>
<tr>
<th>Date time</th>
<th>Mean wind Speed</th>
<th>Wind direction</th>
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<tr>
<td></td>
<td>Mean</td>
<td>Standard deviation</td>
</tr>
<tr>
<td>7, July 17 : 20</td>
<td>3.1 m/sec</td>
<td>177° (S)</td>
</tr>
<tr>
<td>8, July 15 : 30</td>
<td>3.2</td>
<td>195° (SSW)</td>
</tr>
<tr>
<td>8, July 17 : 20</td>
<td>3.4</td>
<td>196° (S)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Date time</th>
<th>Mean wind Speed</th>
<th>Wind direction</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean</td>
<td>Standard deviation</td>
</tr>
<tr>
<td>6, July 8 : 50</td>
<td>3.1 m/sec</td>
<td>190° (S)</td>
</tr>
<tr>
<td>6, July 11 : 40</td>
<td>3.5</td>
<td>188° (S)</td>
</tr>
<tr>
<td>6, July 11 : 50</td>
<td>3.3</td>
<td>186° (S)</td>
</tr>
</tbody>
</table>

Table 2. Steadiness of the wind direction of the gravity wind on a snow patch. In comparison, the analogous data for a non-gravity wind in the same wind speed range is also shown. Each sampling period is 10 minutes. In the column of mean wind direction, magnetic North is taken as 0°.
by the gravity force, it would be reasonable to consider that the wind direction is determined by the direction of the slope.

3.1.3 Relation between the height and intensity of the maximum wind speed

The relation between the height of the maximum wind speed and its intensity is shown in Fig. 4 for the gravity wind observed in periods 1 and 2. This relation is nearly linear. Data in this figure are taken from periods when the gravity wind was blowing steadily for at least 10 minutes. The case of period 1 is shown by white circle and of period 2 by dark circle. The gravity wind was generally stronger during period 1, when the snow patch was larger than in period 2. This fact will be discussed later in section 3.2.

In spite of the difference in the range of wind speed in the two periods, the points plotted in Fig. 4 seem to fit one regression line. The regression line in the figure is given as

\[ Z_{\text{max}} = 0.32U_{\text{max}} + 0.21 \]  

where \( Z_{\text{max}} \) is the height of the maximum wind speed in meters, and \( U_{\text{max}} \) is the maximum wind speed in m/sec. This result is quite interesting when considering that there were differences between the two periods in the factors which contribute to the development of the gravity wind, such as the roughness length of the snow surface, the inclination, the air temperature and the fetch length. The roughness length of the snow surface obtained from the wind profile during the occurrence of the general wind was less than 0.1 cm in the case of period 1 and 0.5 to 0.8 cm for period 2. The roughness length of the bare ground was around 0.5 to 2.0 cm. The difference of the roughness length in the two periods is due to the difference in the development of the ablation hollow. The inclination of the slope was 5° for period 1 and 8° for period 2. The free air temperatures were nearly same in both observation periods, namely, in the range of 10 to 13°C, but that in period 1 was slightly lower or the average.

In Fig. 4, a similar relation obtained by Hoinkes (1954a) for glacier winds on several different glaciers is also plotted. The height of the maximum wind speed is rather high in comparison with our observations. A similar tendency can be seen in the observation by Martin (1975). It is quite difficult to explain such difference, since many factors effect the development of the gravity wind.

3.1.4 Relation between air temperature and wind speed

The driving force of this type of wind is the difference in the air temperature between the free air and the air layer near the snow surface. Therefore, the relation between such temperature difference and the intensity of the wind was obtained. The air temperature at 2.4 m level (\( T_{2.4} \)) is taken as the free air temperature, and that at 0.4 m level (\( T_{0.4} \)) as the cool air layer. \( T_{2.4} - T_{0.4} \) was correlated to the wind speed at 0.4 m level (\( U_{0.4} \)) which will show the strength of the gravity wind, as shown in Fig. 5. There is a linear relation expressed as
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\[ U_{0.4} = 0.275(T_{2.4} - T_{0.4}) + 0.20 \quad (2) \]

where \( T_{2.4} \) and \( T_{0.4} \) are in °C and \( U_{0.4} \) in m/sec. The corresponding relation obtained by Martin (1975) for a glacier wind is shown by A in the same figure. It can be noticed from the position of line A and our regression line that the wind speed is larger in the case of Martin than our observations under the same temperature difference. This is probably due to the difference of the area of snow above the observation site in the two cases. Martin's observation site was in the ablation area of a medium size glacier (3.1 km in length; Lliboutry, 1965), and ours on a small snow patch. As will be seen in the next section this factor has a great effect on the development of gravity winds.

3.2 Dependence of wind speed on the position at the snow patch

As shown in Fig. 4, there is a difference in the range of wind speeds between the case in period 1 on a large snow patch and in period 2 on a small snow patch. In these two cases, there were difference in the factors which affect the development of the gravity wind, such as roughness length, inclination of the slope, and free air temperature, and the size of the snow patch, in other word fetch length. There are empirical equations which relates the wind speed of gravity wind along the stream line with the contributing factors such as air temperature, inclination of the slope and fetch length. When considering these equations such as ones obtained by Reiher (1936) and Bergen (1969), fetch length seems to be the only factor that could explain the difference in the wind speed in period 1 and 2.

In period 3, on observation was made in order to investigate the dependence of wind speed on the position within the snow patch. The sites of the measurements were C, D and E in Fig. 2a, which were selected to be located at different distance from the top end of the snow patch along the center line. The wind speed was observed at a height of 1.0 m during periods with strong steady gravity wind. The relation between wind speed and the distance from the top end is shown in Fig. 6. The data was taken simultaneously at the three points.

It is evident that the gravity wind is stronger at lower positions that is further down valley. The range of wind speeds measured on the small snow patch in period 2 is also shown by A, in Fig. 6. These data fit well with those from C, D and E. In those two periods the two free air temperatures were nearly the same, in the range of 10 to 13°C as mentioned before, so the relation of the data in period 2 to those in period 3 can be explained by the dependence of wind speed on the distance of the observation point from the upper end of the snow patch.

An simple relation which fits our result seems to be

\[ U = 0.117 \sqrt{l} \quad (3) \]

where \( U \) is the wind speed at the height of 1.0 m in m/sec, and \( l \) is the distance along the slope from the upper end of the snow patch in meters. The result is shown in Fig. 7. The value of the wind speed used in this figure is the average value of the 5 cases in Fig. 6 when the wind speed at all 3 observation points were available. With the average value the range of the wind speed in the 5 cases is also shown. Simple relations between the wind speed and topographical elements such as height difference, fetch length inclination and air temperature, obtained by Reiher (1936) and Bergen (1969), were not applicable to our results.
Fig. 7 Relation between the wind speed (U) and the square-root of the distance from the upper end of the snow patch. From the observation of wind at different places on the snow patch, the origin of this wind system can be discussed. In many cases, gravity wind on a snow patch was occurring simultaneously at three observation points C, D and E. Therefore, from a statistical point of view, it can be stated that on a snow patch of this size, the origin of the wind is the upper part of the snow patch.

3.3 The effect of a general wind on the occurrence of the gravity wind

The gravity wind on the snow patch is not continuously blowing, but controlled by the general wind in the valley. “General wind” in this case means both the upper wind blowing into the valley from outside, and also the mountain and valley wind system which is predominant inside the valley on a calm clear day. When the general wind is very strong, the cool air layer near the snow surface is diffused, and no local gravity wind can occur.

The effect of the general wind on the occurrence of the gravity wind was studied on the basis of the observations during Sept. 5–8, 1975 in period 2. During the observation period, Japan was under the influence of the Pacific high pressure system. The observation site was in calm fair weather, and so a typical mountain-valley wind was observed in the whole valley, except on the snow patches. In this period, the gravity winds on snow patches were observed clearly in relation to these wind system which is a larger-scale wind system. Fig. 8 shows the wind at point B on the bare ground and at point A on a snow patch (Fig. 2), in the daytime of Sept. 6. The wind data used were taken at the height of 2.0 m at B and 1.4 m at A. The wind at point B can be considered as the general tendency of the valley. During the whole daytime, the wind at point B was an upslope valley wind, which showed a typical cyclic change. A gravity wind blew steadily at point A on the snow patch in the morning and evening hours, when the valley wind was weak. In the midday, the gravity wind occurred only when the valley wind was weak. The value 1.2 m/sec was the critical value of the valley wind speed at point B to change the wind at point A from gravity wind to valley wind. In the case of less than this wind speed, the gravity wind predominated at point A on the snow patch, and in the case of larger than this value the valley wind blew at point A.

Fig. 9 indicates the relation between the wind at point A (U_A) and B (U_B). The broken line shows the case when the wind speed at point A and B are the same. This figure contains not only the daytime situation when the valley wind occurred in the valley, but the nighttime situation when the mountain wind was occurring, for the period Sept. 5–8. The open circles indicate cases in which the height of the maximum wind speed of the gravity wind was lower than 2.0 m, and the dark circles higher than 2.0 m.

The reason for such classification is as follows. In this paper, the gravity wind on a snow patch is defined as the wind which has its maximum wind speed at a height lower than 2.0 m above the snow surface. According to such a definition, we can separate this wind from the mountain wind, for the mountain wind observed at point
B always had its maximum wind speed at a height higher than 2.0 m.

On the basis of Fig. 9, the range for the occurrence of gravity wind on the snow patch can be given by the wind speed of the general wind at B($U_B$) as

$$-2.0 < U_B < 1.1$$

when the wind direction is indicated as + for the northerly wind which is the upvalley wind, and for the southerly downvalley wind. For the range of $1.1 < U_B < 1.5$, both gravity wind and valley wind blow on the snow patch. For the range of $1.5 < U_B$, the wind on the snow patch is always a valley wind. For $U_B < -2.0$, the wind on the snow patch is a downvalley, in most case a pure mountain wind.

3.4 Characteristics of the gravity wind on a snow patch

The most important character of the wind observed here is the shallowness of the wind system. The height of the maximum wind speed was lower than 1.2 m above the snow surface for most of the cases, and its intensity was 1.5–4.0 m/sec, sometimes more than 4.0 m/sec. This wind system is quite shallow, in spite of its strong wind speed. This is closely related with the weak turbulence in the air layer, which is very stable. As seen in Fig. 3, the air temperature gradient is large and a thermocline exists. The turbulent diffusion within it will be suppressed, so the momentum and heat will not be easily diffused through these layers. This is the main reason for the shallowness. The above noted character is primarily the result of the strong contrast between the temperature of the snow surface and the free air temperature.

Another interesting feature of this wind is its areal development. It was shown in 3.2 that the wind was stronger over the lower parts of the snow patch. Two processes can be conceived. One is the acceleration of wind along the flow line due to the cooling by the snow patch. Another is the convergence of the wind due to the concavity of the surface of the snow patch. But it is not clear which process is the dominant factor.

4. Discussion

The characteristics and occurrence of the gravity wind on snow patches have been described in this paper. On the basis of the physical processes controlling the development of the wind system, this wind can be classified into the same category as the nocturnal drainage wind, mountain wind, katabatic wind on ice caps and ice sheets and glacier winds.

The controlling factors for the development of these winds can be considered as follows:

(a) temperature difference between the air near the surface and the free air
(b) inclination of the slope, and the topography of the area
(c) lapse rate of the free atmosphere
(d) roughness of the surface
(e) diffusion coefficient of the air
(f) upper air wind.
But the main factors can be different in each case.

Prandtl (1952) has developed a theory of mountain and valley winds and derived a theoretical profile of wind speed and air temperature for these winds in general. Defant (1951), Lettau (1966) and Imaoka (1964) have succeeded in applying this theory to their observations of katabatic winds. But, Martin (1975) showed the non-applicability of this theory to the observations of glacier wind.

Therefore, it is interesting to check the applicability of Prandtl’s theory to the gravity wind over a snow patch. Prandtl’s formulation is given as

\[ U_z = C \sin \left( \frac{z}{l} \right) \exp \left( -\frac{z}{l} \right) \]  (4)

where \( z \) is height perpendicular to the slope; \( C \) and \( l \) are coefficients determined from physical factors such as (a), (b), (c) and (e); \( U_z \) is the wind speed at height \( z \). It is quite difficult to determine each parameter from observations of the gravity wind on snow patches. But a wind profile can be obtained from the equation when we assume the value of maximum wind speed and its height.

One case of comparison of our observational data with a theoretical profile is shown in Fig. 10. A typical profile of the gravity wind on a snow patch is shown by open circles. The profile derived from Eq. (4) is shown by a broken line. These two do not fit well.

The reason for such discrepancy can be considered as follows. The assumption of \( U \) depending only on \( z \) in Prandtl’s theory is unrealistic in the case of the gravity wind on snow patches. We saw in section 3.3 that the wind speed of the gravity wind at two places few hundred meters apart had a difference of 1.0 m/sec or more. So a term of subsidence or convergence of flow has to be considered, and this will invalidate the assumption of \( U \) depending only on \( z \). Another is the assumption of the constant vertical diffusion coefficient. Such an assumption is not realistic, since the air layer near the snow surface is quite stable. Perov et al. (1967) have done a numerical calculation of the slope wind, and have shown that when the eddy diffusion coefficient changed from constant to exponentially changing value through the air layer, the height of the maximum wind lowered considerably. This effect should be considered in the case of the gravity wind on snow patches.

Martin (1975) obtained an empirical equation, when the Prandtl’s equation did not fit to his observation of glacier wind. It is given as

\[ U_z = A \cdot \ln \left( \frac{z}{a} \right) \cdot \exp \left( -\frac{z}{b} \right) \]  (5)

where \( A \) (m/sec), \( a(m) \) and \( b(m) \) are coefficients determined from the observed wind profile. When \( b \to \infty \), this equation (5) becomes the equation giving the logarithmic profile. As shown by the full line in Fig. 10, it is possible by this equation to obtain a profile fitting the observed values of the gravity wind on a snow patch. Therefore it will be important to find a physical basis for Eq. (5).

Such investigations and also a development for an improved theoretical model which incorporates the needed parameters are essential in understanding the mechanism of the gravity wind reported in this paper and comparing it with other observed gravity winds.

The radiative flux divergence is one process which has not yet been mentioned, but may be quite important in determining the vertical heat transfer of the gravity wind which has a large air temperature gradient.

Finally, a comment will be made on the effect of this wind on the ablation of the snow patch. This arises from the fact that the gravity wind is generated by the sensible heat loss of the lowest air layer, which mean the heat gain of the snow patch. As the gravity wind blows when the general winds are weak, this local wind might have a great effect on the heat transport at the snow surface. Hoinkes (1954b) has studied the glacier wind and concluded that these local winds will

Fig. 10 Applicability of the theoretical and empirical formulas to the observational results.
increase the sensible heat transfer to the glacier surface, on account of the strong air temperature gradient. It would be important to study this point in the case of the snow patches, since the gravity wind on snow patches has an extreme profile. Moreover, an investigation of the effect of this wind system to the ablation of snow patches as a whole is important, especially on large snow patches where this wind system is strong.

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References