The Influence of Seasonally Varying Atmospheric Characteristics on the Intensity of Nocturnal Cooling in a High Mountain Hollow

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(Manuscript received 9 August 2001, in final form 28 January 2002)

ABSTRACT

Seasonal differences in the longwave radiation balance, diurnal cycles of humidity and cloudiness, and ambient winds during the warm season were evaluated to determine their effect on the nocturnal cooling processes of a high mountain hollow in central Japan. This hollow is located at an elevation of 2230 m at the bottom, having a depth of 50–200 m with a diameter of approximately 1 km. One of the most marked seasonal changes in the atmosphere is a reduction in the downward longwave radiation from summer to autumn due to reduced water vapor. This reduction is larger than that of the upward longwave radiation due to a reduced surface temperature. This radiation balance resulted in larger (smaller) potential intensity of radiative cooling during autumn (summer). A composite analysis of diurnal variation for days in summer and autumn showed that the humidity and cloud conditions, differing between the seasons, determined the initial timing and development of nocturnal cooling. During summer (July and August), increased downward longwave radiation in the evening caused by a combination of increased water vapor and cloud cover over the mountains, suppresses and delays cooling. In contrast, dry and fair weather conditions throughout early autumn (September and October) cause strong and continuous cooling. Despite the high potential for radiative cooling during late autumn (after mid-October), nocturnal cooling was frequently disturbed by strong ambient winds exceeding 3.6 m s\(^{-1}\). The weak winds during summer and early autumn have a minor effect in determining the actual cooling intensity.

1. Introduction

Distinctive climates due to topography, such as those with cold-air pools and drainage flows, develop during the nocturnal cooling of the land surface in mountainous areas. In recent decades, comprehensive approaches coupling intensive observational results with theoretical analyses, such as summarized by Kondo (1985), Whitman (1990), and Barry (1992), have provided an understanding of the processes and structures that depend largely on the surrounding terrain and the ambient atmosphere. However, those studies were generally conducted when the conditions were optimal—that is, under clear and calm ambient weather, based mostly on a short period not exceeding one month. With this background in mind, based on interseasonal observations made at a high mountain hollow in central Japan during the warm season, Iijima and Shinoda (1998, 2000) investigated the seasonal differences in the processes of cold-air pool formation between summer and autumn, and related the differences to water vapor content. The results showed that during summer the large water vapor content in the evening delays and weakens cold-air pool formation, whereas strong cold-air pools were formed under dry air conditions during autumn. These studies, however, did not provide conclusive observational evidence that the downward longwave radiation is strongly related to the water vapor content.

The primary driving force of nocturnal cooling within a basin is the net loss of radiation from the bottom of the hollow and from the sidewall surface rather than the radiative flux divergence of the basin atmosphere (Maki et al. 1986; Maki and Harimaya 1988). The large contribution of downward longwave radiation to cold-air pool formation was quantitatively estimated by Mori and Kondo (1984) and Maki and Harimaya (1988). The seasonally varying effects of water vapor on nocturnal cooling through the effect of downward longwave radiation was suggested based on the theoretical estimation of the potential intensity of nocturnal cooling (Kondo 1982). An observational attempt should, therefore, be made to confirm this theoretical consideration. The initial focus is on the direct measurement of downward longwave radiation, to evaluate the seasonal changes in the intensity of nocturnal cooling.

The second focus is on the ambient winds that act to erode the nocturnal cooling layer within the hollow. The ambient winds exhibit marked seasonal variation with rather strong westerly winds developing during winter, and may thus have a significant contribution to seasonal

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variation. It is a well-known fact that the strength of the ambient winds on the ridge top of a valley or a basin is closely related to the nocturnal drainage wind speed and depth, depending on the topographical structure of the valley (Yasuda et al. 1986; Barr and Orgill 1989; Orgill et al. 1992; Petkovsek 1992). The reduction in nighttime air temperature near the ground is also related to the synoptic winds at a height of 1000 m AGL (Kondo and Mori 1982). However, because most valley field experiments have focused on limited periods when the ambient winds were weak, little data has been accumulated concerning the effects of variable ambient winds (Whiteman 1990).

To investigate these matters, interseasonal field observations were made during the warm season of 1999 in a high mountain hollow of central Japan. Attention was focused on the longwave radiation balance, diurnal cycles of humidity and cloudiness, and ambient wind fields.

2. Observational description

Field observations were conducted in a remote subalpine hollow (Fig. 1) in the north Yatsugatake Range of central Japan (36.1°N, 138.2°E). The bottom of the hollow is at an elevation of 2230 m and has a near-flat surface composed of gravel and sand. The enclosed area of the hollow has a diameter of approximately 1 km. The difference in elevation between the bottom of the hollow and the surrounding ridges ranges from 50 m at the lowest point at the south of the ridge to 200 m at the highest point at the west. One of the remarkable characteristics of the landscape in the hollow is the vegetation inversion. That is, alpine tundra plants and Japanese alpine pine (*Pinus pumila*) communities are distributed around the bottom, whereas a subalpine conifer forest covers the surrounding slopes. The snow-free season is from mid-May to early December.

Meteorological observations were carried out from 21 July to 4 December 1999 (136 days). The main observation system was set up at the bottom of the hollow (P.A, shown in Fig. 1). Direct measurement of downward longwave irradiance was made using a precision infrared radiometer (PIR; The Eppley Laboratory, Inc.). The output from the PIR was converted to the downward longwave radiation using the compensation formula (Shiobara and Asano 1992) using sensor and dome temperature measurements. Other climatological elements, such as air temperature and relative humidity at two heights (0.5 and 1.8 m), incoming and reflected solar radiation, and net radiation and ground temperatures at two depths (5 and 10 cm) were observed at the same site. Observational data were recorded using a data logger (CR-10X, Campbell Scientific, Inc.) at 10-min intervals.
3. Methods of analysis

a. Potential nocturnal cooling

The amount of surface radiative cooling can be measured in terms of the nighttime radiation balance. The potential intensity of nocturnal cooling can be calculated using the ground surface temperature and downward longwave radiation data, assuming calm conditions in which there are no sensible and latent heat fluxes (Kondo 1994).

The radiation budget at the surface under calm conditions around sunset is given by

$$\begin{align*}
R_n &= \sigma T_0^4 - L_0↓,
\end{align*}$$

(1)

where $R_n$ is the net radiative flux at the surface (the positive value denotes upward flux), $L_0↓$ is the downward longwave radiation (positive in the downward direction), $\sigma$ is the Stefan–Boltzmann constant ($5.67 \times 10^{-8}$ W m$^{-2}$ K$^{-4}$), and $T_0$ is the surface temperature. The $0$ subscript denotes the time at which nocturnal cooling begins, which is generally around local sunset (Kondo and Mori 1982). Under normal conditions in which $R_n$ is an upward (cooling) flux, the radiative energy at the ground surface continues to decrease because of surface radiative cooling. Hence, the surface temperature decreases with $R_n$. Therefore, if we assume that $L_0↓$ is constant throughout the night, then the net radiative flux gradually tends to zero. Under such limiting conditions, $R_n$ is given by

$$R_n = \sigma T_{rad}^4 - L_0↓$$

(2)

where $T_{rad}$ is the ultimate temperature at which the surface temperature can decrease because of surface radiative cooling. Using (1) and (2), the potential nocturnal cooling $\Delta T_{max}$ is given by

$$\begin{align*}
\Delta T_{max} &= T_0 - T_{rad} = \frac{R_n}{4\sigma T_0^4} \\
&= T_0 \left[ 1 - \left( \frac{L_0↓}{\sigma T_0^4} \right) \right].
\end{align*}$$

(3)

In this study, $T_0$ is defined as the air temperature at the time of the astronomical sunset when the net radiation changes from positive (heating) to negative (cooling). Because the surface temperature was not measured, the air temperature at a height of 50 cm was used instead. According to Kondo (1982) and Mori and Kondo (1984), air temperature during night conditions near a surface having roughness parameters of $10^{-2}$–1 cm, such as that found at the base of the hollow, is approximately equal to the surface temperature when a stable layer is formed.

b. Actual nocturnal cooling

The atmosphere in a basin is cooled more effectively at night because of the influence of both direct and indirect (i.e., cold-air accumulation from surrounding slopes) radiative cooling at the surface than on a plain or in a well-drained valley (Mori et al. 1983; Maki et al. 1986; Kondo et al. 1989; Kondo and Okusa 1990). On the other hand, the longwave radiation from the sidewall to the basin has an effect of heating and slowing the cooling process in a narrow valley (Whiteman et al. 1989). Thus, the magnitude of air temperature reduction during nighttime at the basin bottom is determined by the net result of the nocturnal cooling and warming processes. The nighttime temperature reduction near the ground surface has been frequently used as an index of the intensity of actual nocturnal cooling (e.g., Kondo and Mori 1983; Maki and Harimaya 1988; Toritani 1990). In the current study, the intensity of actual nocturnal cooling is defined as follows:

$$\Delta T = T_{0} - T_{min},$$

(4)

where $T_{min}$ is the minimum air temperature at the bottom of the hollow between 1800 and 0600 JST (Japanese standard time) and $T_0$ is as defined in (1).

c. Ambient wind conditions

The actual intensity of nocturnal cooling is greatly influenced by ambient winds because of the turbulent mixing of air (Yasuda et al. 1986; Orgill et al. 1992; Petkovsek 1992). Thus, it is necessary to investigate the relationship between nocturnal cooling and ambient wind speed. Haginoya et al. (1984) demonstrated that the correlation between the ambient wind measured at rawinsonde stations and the ambient wind at a mountain top in central Japan is significant provided that the elevation of the mountain exceeds 1500 m. It has been found similarly that the correlation coefficient between observed winds at the summit of Mount Inakadake (2380 m; see Fig. 1) and winds estimated from the rawinsonde data for Tateno and Wajima during 1997 (Iijima and Shinoda 2000) was statistically significant (correlation $R = 0.80$, root-mean-square error $= 1.2$ m s$^{-1}$). Thus, for the observational period in 1999, the nocturnal wind speed at the ridge top of the hollow was estimated by taking the vector average of winds at an elevation of 2380 m observed at three times daily (2100, 0300 and 0900 JST) at the Tateno and Wajima stations.

4. Seasonal changes in downward longwave radiation

To ascertain the seasonal differences most important to surface radiative cooling in the hollow, trends in downward and upward longwave radiations were evaluated (Fig. 2a). Fair and cloudy weather conditions are defined as days when the ratio of observed daily total solar radiation to the estimated value under clear weather conditions (Kondo 1994) is greater than 0.7 (marked as open circles in Fig. 2) and less than 0.4 (black dots), respectively.
The downward longwave radiation under fair weather conditions exhibits a seasonal maximum (approximately 350 W m$^{-2}$) in mid-August and a minimum (210 W m$^{-2}$) in early December, with an abrupt decrease during late September. The upward longwave radiation at local sunset indicates a pattern similar to that of the downward component; the seasonal reduction from summer to autumn is significantly smaller. The time series shows that the net longwave radiation (upward flux expressed as positive) was larger in autumn than in summer, due to a greater reduction in the downward radiation. In other words, the net loss of longwave radiation is enhanced during autumn because of the combination of the substantially reduced downward radiation and slightly reduced upward radiation.

The abrupt reduction in downward radiation during late September coincides with those in air temperature and water vapor pressure (Figs. 2b and 2c). In general, surface air temperature and water vapor pressure are closely correlated and are used for estimating daily downward longwave radiation (Brunt 1932; Robinson 1950; Nakagawa 1977; Sato 1983). In fact, the downward longwave radiation has a high correlation with both the air temperature ($R = 0.88$) and water vapor pressure ($R = 0.94$).

Figure 3 shows a monthly comparison of the downward longwave radiation under fair weather conditions and cloudy days for the period of July–November 1999. The difference in daily averaged downward longwave radiation between fair weather and cloudy days was approximately 50 W m$^{-2}$ in July, August, and September, as compared with 70 W m$^{-2}$ in October and November. The smaller difference during the summer months resulted from the fact that both the fair and cloudy weather days had a relatively high water vapor pressure and high temperature (Figs. 2b and 2c). On the other hand, the larger difference during the autumn months is due to both lower water vapor pressure and lower air temperature on the fair weather days.

Figure 4 represents the diurnal variation in the monthly averaged downward longwave radiation for August, September, and October. The diurnal range of downward
longwave radiation was approximately 30 W m$^{-2}$ throughout the observational period. In August, the downward longwave radiation was relatively high in the evening (2000 JST), but it gradually decreased from August to October. The astronomical sunset time occurred at 1830 JST in August and 1700 JST in October. Because the difference in the radiation continued after sunset, the difference does not result from the warming effect of solar radiation on the surrounding slope (Whiteman et al. 1989).

Cloudiness and water vapor pressure increased in the hollow in the summer evenings under fair weather conditions. Cloudiness and water vapor pressure were lower during autumn (Iijima and Shinoda 1998, 2000). The large water vapor pressure and cloudiness in the summer evening are most likely due to the convergence of water vapor from the plain and the basin to the mountainous area through the thermally induced circulation in central Japan (Kuwagata and Kimura 1995; Kuwagata 1997; Kimura et al. 1997). Thus, the downward longwave radiation, enhanced by the water vapor convergence, has a significant influence on the nocturnal cooling over a mountain. These conditions are met by the study site.

5. Seasonal characteristics of nocturnal cooling during the warm season

The focus of this section is on the relationship between the potential intensity of the calculated surface radiative cooling based on the longwave radiation budget and the actual nocturnal temperature reduction. The seasonal differences in the data and the nocturnal cooling processes are examined.

The seasonal features of the potential nocturnal cooling are shown in Fig. 5a. During late July when the rainy season (known as the Baiu season) ended in central Japan, the potential cooling periodically exceeded 10°C. On these occasions, the radiative cooling was the most
intense and approximated the conditions during the post-rainy seasons of 1995 and 1997 (Iijima and Shinoda 2000). The subsequent summer period exhibited relatively small values of potential cooling because of more frequent cloudy days. After late September, the potential cooling increased and exceeded 20°C during fair weather nights, coinciding with the increased net loss of longwave radiation (Fig. 2a). Figure 5b represents the nocturnal net radiation, which denotes the total component of upward radiation after local sunset until sunrise. The seasonal trend in net radiation coincides generally with the trend in potential cooling. In addition, as mentioned in section 4, the seasonal reduction of downward longwave radiation is much larger than that of upward longwave radiation at sunset (Fig. 2a). This result indicates that downward longwave radiation plays a major role in determining the cooling intensity. That is, nocturnal cooling is suppressed during summer, primarily because of a large downward longwave radiation value, and is enhanced in autumn because of a small radiation value.

The seasonal change in actual nocturnal cooling is presented in Fig. 5c. The ratio of actual to potential cooling for well-cooled days (Iijima and Shinoda 2000), when the actual cooling exceeds 5.0°C, is also displayed in Fig. 5d. During late July, large actual cooling is frequent, reaching values of up to 10°C. Cooling tended to be smaller during August through to the first half of September. After late September, cooling exceeding 5°C was very frequently observed. The strongest cooling, with values of more than 10°C, frequently occurred from late September to mid-October. For the period from 21 July to 15 October, the actual nocturnal cooling was significantly consistent with the potential nocturnal cooling ($R = 0.71$). According to the ratio of actual to potential cooling for the same period (Fig. 5d), nocturnal cooling developed to nearly 70% of the potential intensity. These results suggest that the longwave radiative conditions (expressed as the potential cooling) are of primary importance in the development of nocturnal cooling during this period. Actual cooling conversely has a weaker correlation with potential cooling from 16 October to 3 December ($R = 0.46$) and the ratio was smaller (the average for the period is 0.38). This result suggests that the nocturnal cooling may be weakened by factors other than the longwave conditions, such as the ambient wind. In summary, the effect of surface radiative cooling on nocturnal cooling within the hollow is more significant during summer through to early autumn rather than during late autumn.

Iijima and Shinoda (2000) pointed out that the nocturnal cooling processes are different between summer and early autumn periods, even though the ratios of actual to potential nocturnal cooling are similar. To compare the difference, the diurnal variations of air temperature, water vapor pressure, global solar radiation, and downward longwave radiation were examined (Figs. 6 and 7). These parameters were taken on days where more than 5°C of actual nocturnal cooling occurred during summer and autumn, respectively. The 5°C cutoff was used in Iijima and Shinoda (2000) to detect a typical case of cold-air pool formation in the hollow.

For the summer composite of 8 days (Fig. 6), air temperature decreased gradually through the night in conjunction with a gradual reduction in water vapor pressure (Fig. 6a). With reference to Fig. 6b, local sunset, indicated by an abrupt reduction in global solar radiation around 1600 JST, occurred approximately 3 h prior to astronomical sunset. Although radiative cooling tends to occur after local sunset (Whiteman et al. 1989), there were only small changes in air temperature. Global solar radiation in the afternoon was weaker, having approximately one-half of the August maximum, and downward longwave radiation remained high in the late afternoon (Fig. 6c). These phenomena strongly suggest
cloudy conditions. On the other hand, larger solar radiation was observed in the morning. This change indicates that skies were cloudy in the afternoon and cloudless on the subsequent morning.

Relatively small deviations in air temperature and water vapor pressure in the afternoon indicate that these diurnal variations had very similar patterns among the summer days (Fig. 6a), as compared with the autumn days (Fig. 7a). Moreover, the diurnal conditions are similar to those observed during the summers of 1995 and 1997 (Iijima and Shinoda 1998, 2000). Humid conditions over mountainous areas in the late afternoon are a common characteristic of central Japan and may explain the high value for downward longwave radiation after local sunset. As mentioned in section 4, moist air convergence caused by mesoscale thermally induced circulation is the source of this phenomenon [as proposed by Kuwagata and Kimura (1995) and Kuwagata (1997)]. Those papers also suggested that accumulated moist air over a mountainous area creates conditions favorable to the development of small-scale cumulus clouds. Thus, the formation of cloud and/or fog in the late afternoon, together with humid conditions, is most likely to act to maintain large downward longwave radiation and to weaken and delay nocturnal cooling. Cooling began after midnight from reduced cloud cover and water vapor content, both factors leading to a decline in downward longwave radiation.

A rapid reduction in temperature was conversely observed after local sunset for the autumn composite of 8 days (Fig. 7). When compared with the summer composite, air temperature decreased earlier and more quickly in the evening (Fig. 7a). This occurred in relation to an earlier local sunset and can be seen as an abrupt reduction in the global solar radiation around 1500 JST (Fig. 7b). The global solar radiation in the afternoon and next morning was more than 70% of the maximum radiation in October. This implies that fair weather conditions were persistent during the night in autumn for these selected days. The water vapor pressure is approximately one-half the value for the summer days with little diurnal variation (Fig. 7a). Downward longwave radiation decreased to approximately 300 W m\(^{-2}\) at local sunset and remained at this state throughout the night (Fig. 7c). Thus, typical nocturnal cooling for the autumn days occurred during fair weather and dry atmospheric conditions that persisted throughout the night. The strong cooling after local sunset occurred in accordance with large potential cooling and was not disturbed by water vapor advection, as was found to be the case during the summer periods. In summary, major differences between summer and autumn days are found in relation to conditions of humidity and cloudiness in the late afternoon, which then determines the initial timing and development of nocturnal cooling.

6. The effect of wind on nocturnal cooling during the warm season

Strong nocturnal cooling in a basin or valley tends to be maintained by weak wind conditions. In converse, the cold-air layer is destroyed when the strength of the ambient wind exceeds the threshold wind speed (Yasuda et al. 1986; Orgill and Schreck 1985; Orgill et al. 1992). The intensity of the nocturnal cooling is inversely proportional to the wind speed (Davidson and Rao 1963; Kondo and Mori 1982; Barr and Orgill 1989). In general, the term \textit{ambient wind} refers to the wind around the top end of the atmospheric boundary layer. Over highly mountainous areas such as the study site, the ambient wind is almost equal to the wind on the ridge top, which can be estimated using rawinsonde data.

According to Kondo and Mori (1983), the relationship between the strength of surface nocturnal cooling and the friction wind velocity can be described by the following empirical equation:
The reciprocal of nocturnal ambient wind speed related to nocturnal cooling ratio ($\Delta T/\Delta T_{max}$) on the fair weather days. Nocturnal ambient winds were calculated using averaged wind speed observed at Wajima and Tateno at 21:00, 03:00, and 09:00 JST.

$$\frac{\Delta T}{\Delta T_{max}} = b_0 \tanh \left( \frac{a_0}{u^*} \right),$$

where $\Delta T$ and $\Delta T_{max}$ are the actual and potential nocturnal cooling respectively, $u^*$ is the friction wind velocity at the surface, and $a_0$ and $b_0$ are empirical coefficients. There exists an approximately proportional relationship between the friction wind velocity and ambient wind speed (Kondo and Mori 1983). Therefore, (5) can be rewritten as

$$\frac{\Delta T}{\Delta T_{max}} = b_2 \tanh \left( \frac{a_2}{u_{amb}} \right),$$

where $u_{amb}$ is the ambient wind speed. Kondo and Mori (1982) stated that the coefficient $a_2$ is related to the topographical properties that contribute toward maintaining the cold-air layer, and $b_2$ indicates a cooling ratio under no-wind conditions, which is related to the surface thermal conditions around the site such as a specific heat and thermal conductivity of vegetation and soil.

The relationship between the inverse of the nocturnal wind speed and the cooling ratio ($\Delta T/\Delta T_{max}$) is shown in Fig. 8. Days that fulfill the criteria of fair weather conditions with more than 10°C of potential nocturnal cooling were selected. These days are the most dominant for nocturnal cooling (see Fig. 5a). The solid line on the diagram indicates the optimum curve based on (6) determined by the least squares method. This curve corresponds significantly with the observed value ($R = 0.55$ significant at the 0.1% level). The coefficient $a_2$ of 3.6 m s$^{-1}$ represents an ambient wind speed that when exceeded gradually weakens nocturnal cooling within the hollow. This value is smaller than the estimated value (7–10 m s$^{-1}$) taken from the routine station observation data (Kondo and Mori 1982) and the wind speed threshold (5 m s$^{-1}$) for an above-valley wind that when exceeded erodes the cold-air layer within deep valleys of western Colorado (Orgill et al. 1992). Yasuda et al. (1986), however, showed that the cold-air layer within a shallow valley was formed when the ambient wind speed was less than 4 m s$^{-1}$. Because the base of the hollow is also shallow (less than 200 m in depth), the nocturnal cooling layer within the hollow may be easily destroyed by the relatively weak ambient wind. The dissipation of the nocturnal cooling layer within the hollow at the threshold is likely caused not only by the direct intrusion of the ambient wind into the hollow, but also by intensified turbulent dissipation (Petkovsek 1992). On the other hand, the coefficient $b_2$ represents the expected nocturnal air temperature reduction under fair and calm weather conditions at the hollow. The amplitude ($b_2 = 0.68$) suggests that the actual temperature reduction throughout the night is equal to 68% of the potential nocturnal cooling and is higher than the value (0.40) reported by Kondo and Mori (1982, 1983). The high coefficient indicates that cold-air accumulation effectively strengthens nocturnal temperature reduction at the bottom of the hollow. The surplus in radiative heat loss between the potential and actual cooling (32%) is transformed into ground heat flux throughout the night (Kondo and Mori 1983).

Figure 9 represents the seasonal variation in the nocturnal ambient wind. As shown in this figure, in addition to the before-mentioned threshold of 3.6 m s$^{-1}$, the wind speed of 6 m s$^{-1}$ is used as an upper limit for the formation of a cold-air pool within the hollow, based on the wind observations for the warm season of 1997 (Iijima and Shinoda 2000). Thus, the range between two thresholds represents a transition from strong to weak nocturnal cooling. During summer and early autumn, wind speeds below 3.6 m s$^{-1}$ occurred frequently, and those exceeding 6 m s$^{-1}$ were infrequent. Thus, when compared with the effect of the humidity, the effect of the wind is of secondary importance in determining the actual cooling, because the winds are fairly weak. During late September and October, days of strong actual cooling exceeding 10°C were frequent under weak am-
bient wind conditions. On the other hand, after November, the wind speed increased, exceeding 6 m s$^{-1}$. Moreover, because the prevailing wind direction changed from the west during October to northwesterly, the ambient winds easily intrude into the hollow through the lower ridge. Because of strong winds, only two days experienced actual cooling exceeding 10°C during late autumn, although strong potential cooling occurred almost every day. Thus, it is most likely that the low ratio of actual to potential nocturnal cooling during late autumn (Fig. 5d) was due to a strengthening of the ambient wind.

7. Conclusions

This paper examined the meteorological conditions that affect nocturnal cooling processes in a high mountain hollow during the warm season. Long-term direct measurements of downward longwave radiation (as a radiative factor) and humidity were conducted to calculate the potential intensity of nocturnal cooling. Ambient wind speed was measured as a nonradiative factor of nocturnal cooling. According to these observational results, a schematic diagram showing nocturnal cooling processes specific to each season and related meteorological factors are presented (Fig. 10).

During summer (July and August), the potential intensity of nocturnal cooling decreased because of the increased downward longwave radiation that results from increased cloud cover and water vapor pressure in the evening (Fig. 10a). The convergence of water vapor over a mountainous area may be caused by mesoscale thermally induced circulations under weak ambient wind conditions. The initial nocturnal cooling was delayed until the substantial reduction of cloud cover and water vapor content at midnight. Thus, actual cooling was weakest during summer.

Potential cooling conversely was enhanced during autumn (after September) owing to decreased downward longwave radiation, which is associated with reduced air temperature and water vapor content (Fig. 10b). The dry and fair weather conditions throughout the night cause rapid cooling immediately after local sunset and lead to strong and persistent cooling.

In the case of fair weather days during the warm season, the ratio of actual to potential nocturnal cooling is closely related to the ambient wind speed at the ridge of the hollow, especially during autumn. Nocturnal cooling develops strongly with an ambient wind speed that does not exceed the threshold of 3.6 m s$^{-1}$. Actual cooling was substantially weakened after mid-October because of an increased wind speed, despite strong potential cooling due to decreased downward longwave radiation (Fig. 10c).

These results demonstrated that downward longwave radiation and ambient wind are two major atmospheric factors in determining the nocturnal cooling of high mountainous areas in Japan. These factors are related to Japan’s geographical location. Japan is surrounded by warm oceans that act as a moisture source and is affected by strong midlatitude westerly winds during the cold season. If we consider longwave radiation and ambient winds in relation to a dry continental or maritime climate near the cold ocean, the humidity factor may be less important, and the effect of the wind on suppressing nocturnal cooling may have to be more closely considered. In addition to these atmospheric determinants, future research should consider the seasonal

![Fig. 10. Schematic diagrams illustrating climatic conditions affecting nocturnal cooling at the hollow of Mount Inakodake during (a) summer, (b) early autumn, and (c) late autumn.](image)
variations in surface conditions such as vegetation, soil moisture, and snow cover.

Acknowledgments. The authors thank Dr. T. Kuwagata of the National Institute for Agro-Environmental Sciences for his helpful comments. They are also indebted to many individuals for their observations and other support. This research was supported by the fellowship of the Japan Society for the Promotion of Science for Young Japanese Scientists and a scientific research grant from the Japanese Ministry of Education, Science, Sports and Culture.

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