Diurnal Variations of Snow Precipitation in Wakasa Bay during Winter

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Abstract

The diurnal variation of snow precipitation in the west coastal area near Wakasa Bay has been investigated for two winter seasons of 2001 and 2003 using surface precipitation radar, rain gauge, and satellite infrared data. Radar reflectivity-derived precipitation intensity shows a clear diurnal cycle within Wakasa Bay for both years, although the cycle is clearer in 2001 than in 2003. The precipitation maximum occurs in the early morning, and the minimum occurs in the evening. Using radar data collected during January 2003, the precipitation diurnal cycle within the Bay is compared with three nearby regions: offshore—open water region to the north, inland—land region to the northeast, and inshore—coastal region to the northeast. It appears that all the other three regions have the precipitation maximum during the day and the minimum during night, although the diurnal variation inland (over land) is very small. Additionally, in the offshore region, there exist two precipitation maxima and minima during a 24-hour day. Analysis of precipitation data from AMeDAS (Automatic Meteorological Data Acquisition System) rain gauge stations basically agrees with the radar observations. It shows that a similar diurnal cycle, found in the radar data within Wakasa Bay, can also be found for coastal stations, although the different patterns are shown depending on the location and time period in 2003. When averaging data from both January and February during 2001 and 2003, the diurnal cycle tends to be smoothed out. Cloud top temperature and cloud fraction derived from satellite infrared data do not show a clear diurnal cycle within Wakasa Bay, nearby inshore and inland regions, although the decreases of brightness temperatures are seen around noon in the offshore region for both cases of 2001 and 2003. Analysis using collocated satellite and radar data indicates that cloud top temperature has little skill in reflecting surface precipitation for the winter convective clouds associated with cold air outbreaks. Finally, possible causes of the diurnal variation are discussed, including local land-sea breeze and mountain wind effects, as well as the radiative cooling effect.

1. Introduction

The diurnal variation of local solar insolation influences the atmospheric stability and dynamics, which often leads to a diurnal variation of precipitation. Understanding the change of
the precipitation statistics with the time of day is essential, not only in predicting regional weather, but also in interpreting satellite precipitation retrievals. When data from a non-geosynchronous satellite are used to estimate daily precipitation, the measurements need to be assembled according to the time of day of observation (Bell and Reid 1993). There have been many studies to characterize the diurnal variation of precipitation in various regions (e.g., Oki and Musiake 1994; Chen et al. 1999; Kubota and Nitta 2001). Dai (2001) showed a great deal of analyses of the diurnal variations for various types of precipitation by using surface data in the globe. The characteristics of the diurnal variation of precipitation are highly variable between tropical regions and the middle latitudes over a large scale, and between inland and coastal areas on a smaller scale. There are also season-specific features for some regions (Fujibe 1999).

Although many investigators studied the diurnal variation of rainfall, there have been very few studies on the diurnal cycle of snow precipitation. During the winter months of 2001 and 2003, surface radar observations were conducted in the west coastal area of Japan near Wakasa Bay, as part of the Core Research for Evolutional Science and Technology (CREST) winter Mesoscale Convective Systems (MCSs) project and the US-Japan Aqua validation project. The radar observation provided an excellent opportunity to investigate the diurnal variation of snow precipitation in this area, which is often associated with cold air outbreaks from the Eurasian continent. In this study, the diurnal cycle of snow precipitation is studied within Wakasa Bay and surrounding areas. This variation for cloudiness and cloud top temperature is also investigated by using satellite infrared data.

In studies of the formation of mesoscale circulation in coastal regions, many investigators attributed it to the land-sea breeze (Neumann 1951; Mahrer and Pielke 1977; Nielsen 1989; Yoshikado and Tsuchida 1996). Many features of the diurnal variation of precipitation in the vicinity of Japan have been studied (Fujibe 1988, 1999; Oki and Musiake 1994; Misumi 1999). In these studies, it was reported that a morning maximum prevails in coastal regions. Many previous studies were mainly focused on heavy precipitations in warm seasons. During winter in Japan, precipitation (including snowfall) distribution changes largely by synoptic-scale and mesoscale circulations, although topographical features also play an important role in varying the characteristics (Tachibana 1995). Ishihara et al. (1989) showed the development of mesoscale snow clouds associated with land breeze and winter monsoon flow, and Tsuboki et al. (1989) studied a winter land breeze and its effect on snow clouds. However, the land-sea breeze in the above studies is described as a more seasonal feature than a diurnal one driven by the seasonal land-sea temperature contrast, and can be observed at any time of the day. Here, we will try to explain the diurnal variation of precipitation using a diurnal land-see breeze hypothesis.

In this study, we attempt to answer the following three questions. First, is there a clear diurnal circle of snow precipitation in the Wakasa Bay area? Second, if there is, how different is the diurnal cycle within the Bay, compared to those of the surrounding areas where surface type and surrounding topography are different? Third, is this precipitation diurnal cycle also shown in satellite derived cloudiness or cloud top temperatures? The last question has significant implications to satellite remote sensing. If precipitation signatures for clouds of this type can be extracted from cloud top temperature, precipitation may be estimated from satellite infrared data.

2. Data

Data from two ground-based radars, AMeDAS (Automatic Meteorological Data Acquisition System) rain gauges and GMS satellite are used in this study. Radar observations were conducted from 2 to 17 February 2001 and from 7 to 30 January 2003 in the vicinity of Wakasa Bay, as part of the CREST’s winter MCSs project and the US-Japan Aqua validation project. A 3.2 cm Doppler radar was located at Obama (35.55°N, 135.74°E, refer Fig. 1) in 2001, and a 5 cm Doppler radar was located at Mikuni (36.22°N, 136.14°E, refer Fig. 1). To convert radar reflectivity ($Z$, dBZ) to snowfall rate ($R$, mmh$^{-1}$), The following $Z$-$R$ relationship was used.

$$Z = 18.8963 + 13.1616 \log_{10}(R),$$

(1)
which is derived by comparing radar reflectivity at 1.5 km and surface measured snowfall rate by a weighing type snow gauge (Aonashi et al. 2003). Although this relationship is derived based on the 2001 observations, it was used for radar data of both 2001 and 2003 because our interest is the variation, instead of absolute values of the precipitation.

AMeDAS rain gauge hourly data for four months (January and February during 2001 and 2003) are used. Analysis of precipitation is performed for seven stations near Wakasa Bay and three stations over inland areas. Names of these stations are listed in Table 1, and their locations will be shown in Fig. 4 later.

Satellite infrared (IR) data from GMS for February 2001 and January 2003 are also analyzed. The IR brightness temperature \( (T_B) \) data are obtained from the archives of the Goddard Distributed Active Archive Center. The spatial resolution of pixels is 4 km, and the temporal coverage is every hour.

Figure 1 shows the four domains for which analysis is conducted in this study. Domain 1 (D1: 35.5–36°N, 135.2–136°E) is Wakasa Bay. Domain 2 (D2: 36.7–38°N, 134–136°E) is the offshore area far enough from the land, Domain 3 (D3: 35.7–36.3°N, 132.2–136.8°E) is the inland area, and Domain 4 (D4: 36.4–36.7°N, 136.1–136.5°E) is the inshore coastal area. The Bay is surrounded with mountainous areas.

### 3. Results

#### 3.1 Radar observations of snow precipitation

Using data from the 16-day (02–17 February 2001) radar observations during 2001, the radar-derived precipitation values are averaged over the entire bay area (box D1 in Fig. 1) according to the observation time in a day with 10-minute interval. The diurnal variation of these averaged precipitation anomalies is shown in Fig. 2a by solid dots. The anomaly is defined by removing the domain and whole-period averaged value. The average value is 0.193 mm h\(^{-1}\). We also plot the diurnal variation of radar-derived precipitation for “undisturbed days” in Fig. 2a by triangles. An undisturbed day is defined by the day without large-scale cyclonic cloud systems visible in the domain from satellite images. Undisturbed days are subjectively picked up to avoid changes induced by the passage of synoptic scale low-pressure systems. But they include locally unstable conditions due to the cold air outbreaks. Seven days are found “undisturbed” during the 16-day period, and the average is 0.224 mm h\(^{-1}\). Apparently, a morning high and an evening low can be identified from both plots and the patterns are similar, although the amplitude of the diurnal variation is slightly large for undisturbed days. For all days, the maximum of the anomalies is about 0.1 mm h\(^{-1}\), and the minimum is about –0.09 mm h\(^{-1}\). For undisturbed days only, the maximum is about 0.28 mm h\(^{-1}\), and the minimum is about –0.14 mm h\(^{-1}\).
Similarly, the results for January 2003 are shown in Fig. 2b. Fluctuations of the anomalies are larger than those of February 2001, and the average values are respectively 0.434 mm h\(^{-1}\) and 0.165 mm h\(^{-1}\) for the total period and undisturbed days. On the whole, the positive anomalies show in the morning and the negative anomalies in the evening, except for small positive peaks found around 14 LST and 22 LST in the total period averages. For the 8 undisturbed days, the amplitude of the variation is smaller, and the negative anomalies appear from 16 LST to 23 LST.

In 2003, radar data covering all 4 domains (Fig. 1) are available to us; a similar analysis is done for each domain for comparison purpose using the January 2003 radar data (7–30 January 2003). We excluded data of 23 and 27 January when very large synoptic systems dominated the cloudiness in this total duration. In Fig. 3, the diurnal variations over four domains are represented by hourly averaged precipitation anomalies. For illustration purpose, a sine curve fitting is used to data. Again, it shows a late-night/early-morning high and an evening low in the bay area (Fig. 3a). The phase of the variation in the bay area seems different from those of the other three regions, which show a maximum near the local noon. Offshore over ocean (Fig. 3b), there seem to be two peaks in the morning and afternoon. Over land area (Fig. 3c), while significant, the diurnal cycle seems to be weak. Two minima in the evening and midnight are also shown in Fig. 3c. The local noon maximum is particularly evident for the inshore area (Fig. 3d).
In summary, radar precipitation data from both 2001 and 2003 indicate an early morning maximum and an evening minimum of precipitation within Wakasa Bay. There are also diurnal cycles seen in surrounding areas (over ocean, over land, and along the coast), but the phases of these cycles are different from the one in the Bay. The maxima occur during the day for the surrounding areas. The phase difference implies that the diurnal cycle in the Bay is at least partially caused by mechanisms that do not exist in the other three regions, which possibly is due to its geographical features.

3.2 AMeDAS rain gauge measurements

To examine the variations in surface precipitation measurements, Data from ten AMeDAS rain gauge stations are used. Each diurnal variation of precipitation from the ten stations from 2 to 17 February 2001 is shown in Fig. 4. The variation patterns vary depending on the locations of the rain gauge stations. The values of precipitation in the morning are greater than those in the afternoon and evening for six (KOS, OBA, MAI, MIY, OI, and TAI) of the seven stations close to the shore. Tsuruga (TSU) has several peaks in the morning and afternoon, and the minimum at night. This station is located respectively in the inside area of a small bay within Wakasa Bay. Miyama (MIA), Shiramine (SHI), and Kuzuryu (KUZ) are located in the inland area (D4). MIA and SHI don’t have clear morning maxima, compared to the coastal stations. In KUZ with the high altitude, relatively the smaller amount of precipitation is observed, especially at night. For the whole period of February 2001, the patterns of precipitation are not quite different, except for OI. OI has slightly larger values in the afternoon. For undisturbed days in February 2001, among seven stations close to the shore, six stations have greater precipitation in the morning or early morning, except for Koshino (KOS) and no clear trend can be found for three inland stations. In January 2003, the diurnal precipitation patterns among these stations are very different and complex. Some of them have heavier precipitation in the morning.
than in the afternoon, but some other stations like MAI, OBA, and TSU have the maxima in the afternoon or evening. When the longer-term data (four months of January and February in 2001 and 2003) is used, the diurnal variation is smoothed out by averaging.

In summary, the analysis of AMeDAS rain gauge data for February 2001 basically agrees with the radar observation with respect to the dominance of morning maxima. A similar diurnal cycle to that of the radar data within Wakasa Bay is found for most coastal stations. However, the diurnal variation has a different and more complex pattern, depending on the location and time period during winter in 2003. In addition, the diurnal signal tends to be smoothed out when averaging over a long time period. The diurnal variation of precipitation for inland stations seems to be different from that of the bay area.

3.3 Spectral analysis

Using radar precipitation data areally-averaged over the Bay (D1) during the period of 2 to 17 February 2001 and 7 to 30 January 2003, a spectral analysis is performed. The spectral analysis is based on the fast Fourier transform (FFT). The power spectral density is calculated in order to find any regularity in the data. The Parzen window (Emery and Thomson 2001) is used for smoothing spectral estimates.

Figure 5a is the variance-preserving plot of the power spectral density of precipitation from 2 to 17 February 2001. It is normalized by the variance and frequency-weighted. A clear peak with a period of one day is shown, indicating a strong diurnal cycle. In Fig. 5b, the precipitation power spectrum of January 2003 is shown. While the diurnal signal is not as clear as the previous one, peaks near the period being about 1 and 0.5 day are shown, indicating diurnal and semi-diurnal cycles. The diurnal signals with small magnitudes over the other domains were also found in January 2003 (not shown). Two peaks of the hourly averaged precipitation anomalies are seen over D2 in Fig 3b, but a semi-diurnal signal is not clear in the spectral analysis.

3.4 Satellite infrared brightness temperatures

As the diurnal cycle of snow precipitation was identified in the bay area by surface radar observations, it would be interesting to investigate whether a similar cycle exists for clouds, i.e., cloud top temperature and cloud fraction. Here, GMS IR brightness temperature ($T_B$) is used as cloud top temperature. Cloud fraction is defined by the ratio of the number of pixels colder than a given threshold temperature to the total number of pixels. Several thresholds are used: 260 K, 255 K, 250 K and 245 K.

Figure 6 represents the diurnal variations of the calculated cloud fractions in Wakasa Bay. In Fig. 6a, which shows data from 2 to 17 February 2001, clouds with cloud top temperatures colder than 250 K appear to have small peaks in the morning, evening, and midnight (8–9, 17, and 22–0 LST). The fraction of those cold-top clouds is around 20%. There are a few peaks for the cloud fraction defined by the 260 K threshold: 1, 6–7, and 12–14 LST. The
largest value is found at 1 LST for the 255 K threshold. The morning maxima are not as clear as the radar data. Figure 6b shows the cloud fractions from 7 to 30 January 2003. Several peaks (4–5, 9, 14–15, 20 LST) for the threshold values of 255 K, 250 K, and 245 K are seen. For the 260 K threshold, the trend is similar, but there are no clear maxima.

Using 260 K as the cloud threshold, we derived the averaged cloud top temperature and the cloud fraction for the Bay (D1), offshore (D2) and inland (D3) areas. The results are shown in Fig. 7 and Fig. 8 with the averaged $T_B$ being plotted by lines and cloud fraction by bar graph. In the bay area (Fig. 7a), the average $T_B$ has small variations, and the minima are around 8–9, 18, and 23–0 LST. The 9 o’clock $T_B$ minimum does not correspond with a cloud fraction maximum, which occurs at either 2 hours earlier around 7 LST or several hours later around 12 LST. In all measures, the satellite IR data in 2001 do not show a clear diurnal cycle in the bay area as shown by radar precipitation data. However, as shown in Fig. 7b, in the offshore area of D2 the averaged $T_B$ show a clear diurnal cycle with the colder top temperature showing in the early afternoon and warmer top temperature near midnight. In the inland area (Fig. 7c), a diurnal cycle is not identifiable from either cloud fraction or cloud top temperature, but it is shown that the cloud fraction increases between 12 LST and 20 LST, ...
although this trend is not clear in the brightness temperature variation.

In Fig. 8, several maxima and minima are found in the bay area (Fig. 8a), while TB and cloud fraction do not show any apparent correlation. In the offshore area (Fig. 8b), while the cloud fraction has small variations, the TBs seem to have a diurnal cycle with the coldest value shown near noon. In the inland area (Fig. 8c), a semi-diurnal cycle is shown in TBs with two lows in 10–11 and 19–20 LST. The first low might be related to the precipitation maximum of Fig. 3c, and the second low seems to be related to the increasing trend in the late evening.

To find how different diurnal variations of IR TBs between all days and undisturbed days are, two averaged TBs are also compared (not shown), but it is also hard to identify a distinct pattern for the diurnal variation, especially in 2003.

From the descriptions given above, it appears that the diurnal variations of the precipitation (derived by radar), and the clouds (derived by IR TBs) do not clearly agree with each other. In other words, the clear diurnal cycle in the bay area shown by radar precipitation does not show up in the satellite cloud data, either by cloud fraction or by cloud top temperature. This implies that cloud top temperature and fraction are not good indicators of surface precipitation for this type of clouds (shallow convections). To further elaborate this point, the scatter plot of IR TBs vs. radar precipitation is shown in Fig. 9 using data of January 2003 collocated over the bay area. Since the surface radar grid is a finer than GMS pixel resolution, we averaged all radar observations that fall within the GMS field-of-view for the collocated data. As shown in this figure, the correlation between the two quantities is very weak.

4. Possible causes of the diurnal cycle

During both 2001 and 2003, we found that radar-observed snow precipitation has a maxi-
mum in the early morning and a minimum in the evening. It is also found in AMeDAS rain gauge data in spite of relatively weaker patterns in 2003. The morning maxima of precipitation have been reported in coastal areas of western Japan in warm seasons (Fujibe 1988, 1999; Oki and Musiake 1994). In Wakasa Bay, we discovered the same diurnal variation even during winter. Two possible factors may have contributed to this diurnal cycle in the Bay region: land-sea breeze circulation and radiative cooling at cloud top.

First, the diurnal changes of the wind direction are analyzed using AMeDAS data. As an example, in Fig. 10, the normalized frequency of the wind direction (in degree) in Koshino (KOS), Maizuru (MAI), and Obama (OBA) (ref. Fig. 4) using the AMeDAS wind data of February 2001 is shown in a 2-hour interval. It is defined as the frequency of each wind direction divided by the frequency of the most prevailing wind direction during a given period. We can see the wind direction changes dramatically during the 24-hour period. In Koshino and Maizuru, winds during nighttime are dominated by southeasterly, while they change to, respectively, northerly and northwesterly during daytime. The wind direction also changes greatly in Obama. Winds blow from the southwest during nighttime and morning, and blow northeasterly during daytime. Winds have almost opposite directions between day and night. Therefore, land-sea breeze obviously exists in the Wakasa bay area, and it could have affected the diurnal variation of precipitation. The effect of the land-sea is to enhance convections over water during nighttime. Frequently, the wintertime land-sea breeze effect in coastal areas of Japan is believed so small that it can be ignored, even if the importance of the winter land breeze to the coastal meteorology has been recognized by some studies (Passarelli and Brah 1981; Nielsen 1989). However, wind data collected at these stations clearly showed a wind direction shift between day and night, indicating the land-sea breeze phenomenon in the Wakasa Bay area. As indicated by Mizuma (1995; 1998) for Osaka Bay and western Seto inland sea, although the land-sea temperature contrast appears small for generating a strong circulation, the combined effects of mountain winds caused by the geographic feature of Wakasa Bay surrounded by mountainous areas could have induced land-sea circulation. During nighttime (until early morning), winds blow from the land (mountain side) to the sea. Mahler and Pielke (1977) reported that the combination of land-sea breeze and mountain winds intensified the local circulation through their numerical modeling experiments. From the results in Fig. 10, there are southeasterly winds in the east part of Wakasa Bay and southwesterly winds in the west part during nighttime, i.e., winds are converged within the bay. This convergence enhances the development of clouds during night and early morning hours. Tsuboki et al. (1989) also reported that the land breeze strongly influenced the precipitation distribution and snowfall could be enhanced due to the land breeze. On the contrary, when winds blow from the sea during daytime, the reverse situation occurs, which causes precipitation to decrease in the bay area. The development of this local circulation can also be affected by the synoptic conditions but in this study, the detailed mechanism of the effects of the large-scale system on the local diurnal circulation is not considered.

When clouds present, radiative cooling at cloud top during night makes the cloud layer less stable, therefore, enhances convective activities. Figure 11 shows the profiles of the net radiative cooling rate estimated at 5 different times during the day in the location of 35.75°N and 135.6°E, which are calculated using the SBDART (Santa Barbara DISORT Atmospheric Radiative Transfer) model developed to compute plane-parallel radiative transfer in clear and cloudy conditions within the earth’s atmosphere and at the surface (Ricchiazzi et al. 1998). In this calculation, an ocean surface and a mid-latitude winter atmosphere are assumed. Also, the rural aerosol model option with the visibility, 23 km at 0.55 μm is chosen. The number of internal radiation streams is 4. The cloud assumed in the calculation is a low level cloud with optical thickness of 25 and cloud drop effective radius of 10 μm. It locates between 1.5 km and 2.5 km over ocean surface.

It is seen that the cooling rate near the cloud top is ~6 K/day during nighttime, and ~2 K/day near local noon. Excessive cooling near cloud top during night destabilizes the low level atmosphere and favors convective activities,
which in turn could have increased precipitation. Therefore, the difference in cooling rate near the cloud top between day and night could also be a factor in producing the observed diurnal cycle.

5. Conclusions

The diurnal variation of snow precipitation in the west coast area near Wakasa Bay has been investigated for two winter seasons of
2001 and 2003 using surface precipitation radar, rain gauge, and satellite infrared data. Radar reflectivity-derived snow precipitation intensity showed a clear diurnal cycle within Wakasa Bay for both years, although the magnitude of the cycle is greater in 2001 than in 2003. The precipitation maximum occurs in morning and the minimum occurs in the evening. Using radar data collected during January 2003, we compared the precipitation diurnal cycle within the Bay with three nearby regions: offshore, inland and inshore. It appears that all the other three regions have the precipitation maximum during the day and the minimum during night, but the diurnal variation of the inland (over land) is very small. Analysis of precipitation data from AMeDAS rain gauge stations near the coast represents a similar diurnal variation to that found in the radar precipitation data in Wakasa Bay, although the pattern of the diurnal variation of 2003 is not as clear as 2001. When averaging data from both January and February during 2001 and 2003, the diurnal cycle at any stations tends to be smoothed out.

Morning maximum for convective precipitation has been previously reported over coastal regions in tropics and middle latitudes during warm seasons, and this maximum is more evident near land than open ocean in the coastal areas, particularly on the upwind side of the land (Gray and Jacobson 1977; Fu et al. 1990). The analysis of radar data for Wakasa Bay indicates the morning maximum can be distinctly identified in winter snow precipitations. The causes of this diurnal cycle could be a combination of the effect of land-sea breeze and radiative cooling near the cloud top.

In addition, we found that neither cloud top temperature nor cloud fraction derived from infrared satellite observations can be used to produce the same diurnal cycle as derived from radar data. Furthermore, there does not seem to be any clear correlation between cloud top temperature and surface precipitation intensity for the type of clouds that were studied, which are mostly shallow convections associated with cold air outbreaks. The correlation coefficients between IR $T_{bb}$ and radar precipitation are $-0.18$ in February 2001 and $-0.06$ in January 2003. Also, the values between cloud fractions with the threshold of 260 K and radar precipitation are respectively 0.4 and −0.18. For satellite observation of snow precipitation, high-frequency microwave measurement may be an alternative because it measures the scattering signatures caused by snowflakes. Future studies in this area are recommended.

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