Variation in Radiative Contribution by Clouds to Downward Longwave Flux

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Abstract

Clouds strongly influence downward longwave radiative flux and affect the radiation budget at the surface. We evaluated the cloud radiative effect in both absolute and relative terms on downward longwave radiation at the surface; we considered variations in the cloud radiative effect with changes in cloud amount, precipitable water, and cloud base height, as measured by eight stations of the Baseline Surface Radiation Network. The downward longwave radiation predicted by a radiative transfer model agreed well with observations. The cloud radiative forcing and contribution ranged from -21 to 92 W m^{-2} and from -6% to 38% , respectively. The cloud effect shows a positive correlation to the shortwave diffusivity index (an index of cloud amount) and a negative correlation to precipitable water amount. The absolute effect values are small, depending on site conditions, but the relative effect values are larger under dry conditions than under humid conditions. Under humid conditions, the effect of the shortwave diffusivity index is very small. Under dry and cold conditions, such as those found in polar regions, negative values of cloud radiative contribution appear frequently because clouds absorb the emissions from temperature inversion layers. In comparison with prior research that used the A-Train satellite product, the present study shows a wider distribution and a larger maximum value for cloud forcing from amount of water vapor. Cloud effect has a roughly negative relationship with cloud base height, but a positive correlation with cloud base height occurs under low clouds at Tateno, which is located on the Pacific Ocean side of Japan. This correlation is because of the unusual relationship between cloud base height and cloud effect at Tateno during the summer and winter seasons. These results describe small-scale and near-surface variations in cloud effect, which are difficult to detect by satellite measurements.

Keywords longwave radiation; cloud effect; in situ observation

1. Introduction

Downward longwave radiation (DLR) influences climate change, the water cycle, and the radiation budget at the surface. Clouds are an important factor in DLR because they absorb and emit both downward and upward fluxes. Stephens (2005) reviewed the effect of clouds on radiative transfer and the climate system. Cloud cover, on average across

Earth, produces a radiative net cooling effect at the surface. The vertical location of clouds is important for the cloud effect. High clouds tend to warm the atmosphere, especially at low latitudes, whereas low clouds enhance cooling of the atmosphere, especially at high latitudes.

For longwave radiation, clouds warm the surface and the surface air. Sohn and Bennarts (2008) estimated the influence of cloud water path on upward longwave radiation at the top of the atmosphere (TOA) by using the Advanced Microwave Scanning Radiometer for the Earth Observing System. Their research indicated that the spatial variation in cloud upward longwave forcing was less than 5 W m^{-2} even in tropical regions, where convecting thick clouds

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are often present. Schmidt et al. (2010) estimated the global average contributions of water vapor, carbon dioxide, and clouds to upward longwave flux at the TOA to be 50 %, 19 %, and 25 %, respectively. Stephens et al. (2012) calculated the global average downward longwave flux at the surface by four different methods and estimated the cloud radiative effect with data from A-Train satellites. The global average contribution of cloud for DLR was about 24–34 W m⁻². Arnfield (1979) compared some empirical methods for calculating DLR under clear-sky (Brunt 1932; Swinbank 1963; Idso and Jackson 1969) and all-sky conditions (Sellers 1965) at Simcoe, southern Ontario, during the period from July to September in 1969. Under clear-sky conditions, the primitive methods given in Swinbank (1963) and Idso and Jackson (1969), which use screened air temperature, were preferable over other parameterizations for calculating DLR. Under all-sky conditions, it was important to consider both cloud amount and cloud type.

Satellite observations aid in research on global and continuously changing phenomena. Wild et al. (1998) estimated the cloud effect at 24 W m⁻² as a global average by ECHAM4 from the 10-year data from 1979 to 1988. Zhang et al. (2004) showed that the global average for cloud effect is 31 W m⁻² by analyzing the International Satellite Cloud Climatology Project satellite data for 1983 to 2001. Stephens et al. (2012) estimated the cloud effect as 24–34 W m⁻² from A-Train product for 2006 to 2009. However, estimation of DLR from satellite data is not sufficiently accurate. Henderson et al. (2013) used data from CloudSat and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) with Moderate-Resolution Imaging Spectroradiometer (MODIS) data for the 2006 to 2011 and estimated cloud radiative forcing by the BugsRad program (Ritter and Gleyn 1992). Longwave radiation exhibits large uncertainty because of insufficient precipitation at surface temperatures and lower tropospheric humidity (below 500 hPa level). The error in cloud height is ±240 m for CloudSat, and this error may affect surface radiation calculations by 1.5 W m⁻² for surface DLR.

Clouds are complex and variable, and thus knowledge of the effects of clouds on DLR is limited. Although Yamada et al. (2012) estimated the contributions of water vapor, carbon dioxide, and clouds to DLR, the targeted clouds were limited to those thick enough to constitute an overcast condition. Most studies of the effect of clouds on DLR have used

satellite data, dealt solely with the global average, or used in situ measurements in limited areas. Cloud effect on downward longwave flux varies with local conditions such as atmospheric temperature, humidity, and cloud characteristics. Because cloud and water vapor are dominant factors for longwave radiation, it is necessary to investigate the sensitivity of cloud effect to the condition of the atmosphere and cloud state. The purpose of this paper is to quantify the cloud effect on DLR in various cloud conditions from the subtropical zone to the Antarctic on the basis of in situ observations. We also examine the influences of cloud amount and precipitable water at eight observation stations and estimate the effect of cloud base height at four observation stations.

2. Data and analysis

2.1 Estimation of cloud effect

The absolute and relative effects of cloud on DLR are defined as cloud radiative forcing (CRF) and cloud radiative contribution (CRC) by formulas (1) and (2), respectively:

$$CRF = DLR_{Obs}^{All} - DLR_{Cal}^{Clear} [W m^{-2}], \quad (1)$$

$$CRC = \frac{CRF}{DLR_{Obs}^{All}} [\%], \quad (2)$$

where DLR_{Obs}^{All} is the observed downward longwave flux at the surface under the all-sky condition and DLR_{Cal}^{Clear} is the calculated downward longwave flux at the surface under the clear-sky condition.

2.2 Observation data

The following data were collected at eight stations of the Baseline Surface Radiation Network (BSRN; <http://www.bsrn.awi.de/>; Ohmura et al. 1998): the surface downward radiation, vertical profiles of temperature and relative humidity by radiosonde observations, and corresponding meteorological observation data. For flux observations, BSRN aims to achieve errors of less than 5 % in relative measurements or less than 10 W m⁻² in absolute measurements. Table 1 provides descriptions of the observation stations. Figure 1 gives the locations of the stations. All stations have altitudes below 200 m. ISH (Ishigaki island, Japan), which is the hottest and wettest of the stations, has an annual average surface air temperature (T_s) of about 298 K and an average precipitable water (PW) of about 46 mm. The annual average DLR_{Obs}^{All} is 407 W m⁻². In contrast, the annual averages of T_s , PW, and DLR_{Obs}^{All} at SYO (Showa

Table 1. Description of site conditions. DLR_{Obs}^{All} : Observed DLR at the surface under all-sky conditions [W m^{-2}], T_s : Surface air temperature [K], RH_s : Surface air relative humidity [%], PW: Precipitable water [mm], SDI: Shortwave diffusivity index [0–1].

		Station description					
Station	Abbr.	Country	Lat.	Long.	Alt.	Time Period	Time
Bermuda	BER	Bermuda, UK	32.2670	−64.6670	8	2004–2005	12
Fukuoka	FUA	Japan	33.5817	130.3750	3	2010–2012	00
Ishigakijima	ISH	Japan	24.3367	124.1633	6	2010–2012	00
Lindenberg	LIN	Germany	52.2100	14.12220	125	2001–2006	06, 12, 18
Ny-Ålesund	NYA	Ny-Ålesund, Spitsbergen	78.9250	11.9300	11	2001–2011	12, 18
Sapporo	SAP	Japan	43.0600	141.3283	17	2010–2012	00
Syowa	SYO	Cosmonaut Sea, Antarctica	−69.0050	39.5890	18	2001–2011	00, 12
Tateno	TAT	Japan	36.0500	140.1333	25	1996–2012	00
			[°N]	[°E]	[m]	[UTC]	
		Annual mean					
Station	Abbr.	DLR_{Obs}^{All}	T _s	RH _s	PW	SDI	Pyrgeometer
Bermuda	BER	376.5	294.4	76.5	29.7	0.65	PIR
Fukuoka	FUA	357.6	291.3	67.7	28.5	0.67	CG4
Ishigakijima	ISH	407.2	297.8	76.9	46.0	0.72	CG4
Lindenberg	LIN	307.1	280	87.3	15.6	0.72	PIR/CG4
Ny-Ålesund	NYA	255.1	269.1	73.0	6.9	0.78	PIR
Sapporo	SAP	320.9	284.2	69.7	19.1	0.68	CG4
Syowa	SYO	218.8	262.7	68.1	3.6	0.70	CG4
Tateno	TAT	336.6	287.5	73.7	24.0	0.61	CG4
		[W m ^{−2}]	[K]	[%]	[mm]	[0–1]

Fig. 1. Observation sites. At these sites, the calculated and observed values show good agreement.

Table 2. Characteristics of pyrgeometers.

	PIR (Eppely)	CG4 (Klpp & Zonen)	
Wavelength	4~50	4.5~42	μm
Sensitivity	~5	5~10	$\mu\text{V (W m}^{-2}\text{)}^{-1}$
Response Speed	2 (1/e)	6 (63 %)	sec
Temperature Dependability	<1	<1	%
Environmental Temperature	-20~40	-40~80	$^{\circ}\text{C}$

Station (SYO), Antarctica), which is the coldest and driest station, are 263 K, 4 mm, and 219 W m^{-2} , respectively. We categorized the area around stations into three types of regions by annual average PW. Dry regions have annual average PW less than 10 mm; this includes the areas of Ny-Ålesund (NYA) and SYO. Moderate regions have PW values of at least 10 mm but less than 25 mm, and humid regions have values greater than 25 mm. Lindenberg (LIN), Sapporo (SAP), and Tateno (TAT) are in moderate regions. Fukuoka (FUA), Bermuda (BER), and Ishigakijima (ISH) are in humid regions.

BSRN observation data are separated into subsets. We used 0100 basic observation data (DLR_{Obs}^{All} , direct shortwave radiation (DIR), and diffuse shortwave radiation (DIF)), 1000 meteorological data (cloud fraction by human observation and wind direction), and 1100 radiosonde data (vertical distributions of air temperature and relative humidity). To evaluate cloud amount, we used the shortwave diffusivity index (SDI; Long and Ackerman 2000) because of limits in the number and accuracy of cloud amount observations by human observation. SDI is defined by the formula

$$SDI = \frac{DIF}{DIR \times \mu + DIF}, \quad (3)$$

where μ is the cosine of the solar zenith angle. We used only daytime data in estimating SDI. Cloud fraction from human observation is used for comparing calculations and observations only.

DLR_{Obs}^{All} was observed using CG4 (Kipp & Zonen) or PIR (Eppley) pyrgeometers. Table 2 shows the characteristics of the pyrgeometers. BSRN station radiation observations have a sensitivity of 1 Hz and are taken on average once per minute. FUA, ISH, SAP, SYO, and TAT use CG4 pyrgeometers; BER and NYA use PIR pyrgeometers; LIN uses both types. The performance differs between PIR and CG4 pyrgeometers in terms of some properties such as the wavelength range. Payne (2004) compared these two types of pyrgeometers in 2002 for nine months in

Massachusetts and found that the difference in DLR measurements was less than 5 W m^{-2} in a 1-minute average over the observation period. In this study, the DLR observations made by CG4 pyrgeometers are not distinguished from those made by PIR pyrgeometers. To limit the effects of small perturbations and instrumental error, hourly averages are used for radiation data.

Ceilometer and lidar are useful instruments for detecting clouds. Among the eight BSRN stations, LIN and NYA both have an LD-40 Ceilometer (Vaisala) by which they observe cloud base height (CBH); the measurement wavelength is 855 nm and the temporal resolution is 15 s. SAP and TAT each have a Mie lidar from the National Institute for Environmental Studies (NIES). The Mie lidar observes the attenuated backscattering coefficient (ABC) using 532-nm- and 1064-nm-wavelength lasers every 15 min. CBH is estimated from the 1064-nm-wavelength signal (Shimizu et al. 2010). CBH is estimated as the height at which the increment of the ABC for 1064 nm is larger than $4 \times 10^8 \text{ sr}^{-1} \text{ m}^{-1}$. The height resolution of the ABC is 6 m; however, the height resolution of CBH is 30 m because five data segments are averaged to increase the signal-to-noise ratio. The accuracy of cloud detection in the lower atmosphere is unreliable, and thus Mie lidar does not provide CBH below an altitude of 120 m.

Table 3 describes the two instruments. The wavelength and vertical resolution differ between ceilometer and lidar measurements. Here, low clouds are the main target because high clouds are not easily detected by in situ observations, and the difference in accuracy between lidar and ceilometer detection of low clouds is small. CBH is limited to the range between 150 and 10,000 m to account for the lidar measurement range and the poor detection accuracy of thin high-altitude clouds.

2.3 Calculation scheme

To calculate DLR_{Cal}^{Clear} , we used mstrnX (Sekiguchi and Nakajima 2008), which is a two-stream radiative

Table 3. Characteristics of instruments used for observing CBH.

	LD-40 (Vaisala)	Mie lidar (NIES)	
Wavelength	855	1064	[nm]
Mesurement range	8–13000	150–18000	[m]
Vertical resolution	8	30	[m]
Temporal resolution	15 sec.	15 min.	

transfer model. In this study, mstrnX had 370 layers with thicknesses of 1 km from the TOA to 30 km and of 100 m from 30 km to the surface. MstrnX calculates DLR_{Cal}^{Clear} from the vertical distribution of air temperature and concentration of greenhouse gases (GHGs). The concentration of GHGs other than CO_2 was taken from the 1976 U.S. Standard Atmosphere, whereas the concentration of CO_2 in the troposphere was taken from observations (Keeling et al. 2008). The CO_2 concentration varies by season and location. Although this horizontal distribution reaches near 10 ppmv, the effect on DLR_{Cal}^{Clear} calculated by mstrnX is less than 0.2 W m^{-2} , which is negligible. The vertical distributions of air temperature and water vapor were obtained from radiosonde observations.

For the eight stations, DLR_{Cal}^{Clear} shows good agreement with DLR_{Obs}^{All} under the clear-sky condition. Figure 2 shows the comparison between calculated and observed values under clear-sky conditions. The clear-sky condition has values of SDI less than 0.3 and a cloud fraction of 0 oktas by human observation. The data give the following: calculation – observation = $-2.84 \pm 6.16 \text{ W m}^{-2}$ with correlation coefficient $r = 0.996$. Therefore, it is reasonable to estimate cloud effect. We calculated CRF and CRC from data given by mstrnX calculations under the clear-sky condition and by observations under the all-sky condition.

To estimate the cloud effect on DLR at the surface under various conditions, we took SDI, PW, and CBH as parameters. SDI was estimated as the ratio of the sum of direct and diffuse shortwave downward radiation to the diffuse shortwave downward radiation; a value of 0 indicates a completely clear sky and 1 indicates a completely cloud-covered sky (SDI is typically larger than 0 because of aerosol and gases diffusion). PW was calculated from radiosonde observations of the vertical profile of water vapor, and CBH was estimated using the ceilometer and the NIES Mie lidar.

Fig. 2. Comparison between DLR_{Cal}^{Clear} and DLR_{Obs}^{All} under the clear-sky condition. In the clear-sky condition, $SDI < 0.3$ and the cloud amount recorded by human observation is 0 oktas.

3. Results and Discussion

3.1 Impacts of SDI and PW

Figure 3 illustrates the relationships at the eight observation stations between SDI and other measures: CRF, average DLR_{Obs}^{All} (average of ± 0.05 of SDI every 0.1), and CRC. CRF and CRC exhibit large variability, from -21 to 92 W m^{-2} and from -6% to 38% , respectively. Most scatterplots in Fig. 3 are from -10 to 80 W m^{-2} and from -5% to 35% . The average cloud effect increases both absolutely and relatively with increasing SDI.

PW is an important factor in evaluating cloud effect because water vapor is a strong absorber and emitter of longwave radiation. We classified each observation site as having one of three humidity conditions according to its PW: dry (annual mean $PW < 10 \text{ mm}$), moderate ($10 \leq PW < 25 \text{ mm}$), or humid ($PW \geq 25 \text{ mm}$). Figure 4 shows the relationship for each

Fig. 3. Relationships between SDI and CRF (top), average (center), and CRC (bottom). Red curves show the mean with error bars (1 standard deviation).

humidity condition between SDI and other measures: CRF, average DLR_{Obs}^{All} , and CRC. When SDI is less than 0.6, CRF is unaffected by the humidity condition. A comparison of the average CRF for dry and humid conditions when SDI was between 0.9 and 1.0 showed a difference of about 20 W m^{-2} . A comparison of DLR_{Obs}^{All} under the three humidity conditions shows a difference of $40\text{--}60 \text{ W m}^{-2}$, which means that the relative contributions of cloud differ significantly by humidity. The standard deviation of CRC is larger in the dry condition than in the humid condition. Under the dry condition, the average CRC ranges from about 0 % to more than 20 %. In general, the average CRC decreases with increasing PW. The effect of cloud is large under the dry condition because the effect of water vapor is small. Accordingly, differences in

Fig. 4. Relationships between SDI and CRF (top), average (center), and CRC (bottom). Red, green, and blue curves show means for humid (annual mean PW at least 25 mm), moderate (annual mean PW less than 25 mm and more than 10 mm), and dry conditions (annual mean PW less than 10 mm), respectively.

cloud physical characteristics (e.g., cloud thickness, altitude, and amount) are more important under the dry condition than under the humid condition, which causes CRC to exhibit large variability.

Figure 5 shows the relationship that PW has to CRF, DLR_{Obs}^{All} , and CRC. Stephens et al. (2012; hereafter S12) showed the relationship between CRF and column water vapor by using A-Train satellite data. The temporal and spatial resolutions of data were 3 h and $1^\circ \times 1^\circ$, respectively; these resolutions are much coarser than those of the data used in the present study (1 h and 1 pixel). Our results show good correspondence with those of S12 overall. The three following

Fig. 5. Relationships between PW and CRF (top), average (center), and CRC (bottom). Black curves show the mean with error bars (1 standard deviation).

relationships between CRF and column water vapor were found in S12. In areas where the column water vapor amount is about 0–10 mm, CRF increases with increasing water vapor amount. In areas where the water vapor amount is about 10–40 mm, CRF is inversely proportional to the amount of water vapor. Under very humid conditions, in which the water vapor amount is larger than 40 mm, CRF is insensitive to changes in water vapor amount. However, S12 did not show details of the variation in cloud effect. There are two clear differences between the present study and that of S12. One difference is that the present study shows a high incidence of a large negative cloud effect in very cold and dry regions. The other difference is the wider range for CRF in the present study than in S12.

Cloud effect sometimes takes a negative value. In an Arctic study, Shupe and Intrieri (2003) suggested that negative values for cloud effect can be attributed to model or observation errors under the clear-sky condition. However, negative cloud effect values show a tendency to concentrate in the low-PW region in the scatterplots, suggesting that there may be another reason for the negative values. Figure 6 shows the vertical profiles of air temperature and relative humidity on a day when CRC was strongly negative (less than -3%) and on a following day when CRC was very large (more than 20%) at SYO, which belongs to a dry region in this study. A common feature of Figs. 6a, 6b, and 6d is the presence of a strong temperature inversion near the surface when CRC was less than -3% ; panel c also shows a weak inversion. Figure 7 shows the relationship between PW and CRF, partitioned on the presence of a temperature inversion. In the figure, the orange plot and curve show data for which a temperature inversion exists below 3000 m. Data for which no temperature inversion exists are plotted in black. The two conditions show similar tendencies. However, under the dry condition (PW is smaller than 10 mm), the average CRF with temperature inversion is smaller than that with no inversion, although there is large variation in the values.

According to the Stefan–Boltzmann law, black-body radiative emission is proportional to the fourth power of absolute temperature. Downward longwave flux, empirically calculated under the clear-sky condition, exhibits more dependence on air temperature than is accounted for by this law; for example, Swinbank (1963) suggests that the flux is proportional to the sixth power of surface air temperature. The reason for the strong temperature effect is that an increase in the air temperature causes an increase in water vapor pressure at the surface and thus an exponential increase in the column water vapor amount (Deacon 1970). King (1996) indicated that empirical parameterization at the Arctic yielded downward longwave flux at the surface proportional to the fourth power of air temperature or to the sixth power of cloud-base temperature. In polar regions, strong temperature inversions appear frequently because of the ice-covered surface. The inversion layer is strongest in winter, when it is sometimes warmer than the surface air by 30 K, and affects the radiation budget (Phillipot and Zillman 1970; Stone 1993; Curry et al. 1996). The frequent occurrence of negative CRC values in the present study is attributed to the absorption of radiation from an upper-air temperature inversion due

Fig. 6. Vertical profiles of air temperature and relative humidity on days when CRC was smaller than -3% (dashed) and when CRC was larger than 20% (continuous) at SYO (located in a dry region). Continuous (resp. dashed) measurements are from a) 10 Nov 2011, 00 UTC (11 Nov 2001, 00 UTC); b) 22 Jul 2006, 12 UTC (7 Jun 2006, 00 UTC); c) 27 Jul 2006, 12 UTC (23 Jul 2006, 12 UTC); and d) 22 May 2009, 00 UTC (23 May 2009, 12 UTC).

to clouds.

Figure 8 shows the relationship that PW has to CRF and CRC under nearly clear ($\text{SDI} < 0.5$), cloudy ($0.5 \leq \text{SDI} < 0.9$), and overcast ($\text{SDI} \geq 0.9$) condi-

tions. Under nearly clear conditions, the average CRC (average of ± 5 mm of PW every 10 mm) is near 0% , and the scatterplot shows low dispersion, except in dry regions ($\text{PW} < 10$ mm). In contrast, average CRC

Fig. 7. Relationship between PW and CRF. Each dot represents a data point and black line shows the mean with error bars (1 standard deviation). (a) Days with temperature inversion below 3000 m. (b) Days with no temperature inversion.

under overcast conditions is large and roughly inverse to PW. Under cloudy conditions, CRC lies between these two extremes; the scatterplot shows a large dispersion from dry to wet climates.

The present study and S12 differ in the distribution of data. While S12 shows CRF scattered from about -10 W m^{-2} to about 70 W m^{-2} , this study shows CRF scattered from about -20 to about 90 W m^{-2} . In addition, S12 found that CRF was concentrated in a narrow curved band. In contrast, this study finds CRF scattered widely, rather than concentrated. However, in Fig. 8, CRF under the overcast condition is distributed in a way similar to that found in S12, although the maximum value is larger here than in S12. This indicates that the A-Train product cannot detect clouds under nearly clear conditions and therefore misses the effect of small clouds.

DLR is susceptible to local conditions. The small CRC when PW is 10 mm, reported by S12, is ascribed to the difficulty in observing the near-surface atmosphere using a satellite when cloud cover is present; this difficulty is because of the coarse spatial and

temporal resolutions. Yamamoto and Sasamori (1954) showed that 80 % of DLR is emitted from water vapor at pressures below 900 hPa; this finding was based on radiation chart data for Sendai, Japan. Philipona et al. (2004) calculated downward longwave flux using the MODTRAN software package (Version 4.2; Berk et al. 1987) for data from Payerne, Switzerland, and found that more than 90 % of DLR is emitted from atmosphere below 1000 m; 39 % is emitted from atmosphere below 10 m. It is difficult for satellite observations to retrieve near-surface water vapor distribution, which has a disproportionately large influence on DLR.

3.2 Impact of cloud base height

The LIN and NYA stations measure CBH using an LD-40 ceilometer. The SAP and TAT stations use Mie lidar instruments to measure CBH; these sites are near each other but have different climates because of monsoon effects. In this section, we focus on data from these four stations to estimate the effect that CBH has on the cloud radiative effect. Because

Fig. 8. Relationships between PW and CRF (upper panels) and CRC (lower panels) under (a, d) nearly clear ($\text{SDI} < 0.5$), (b, e) cloudy ($0.9 > \text{SDI} \geq 0.5$), and (c, f) overcast ($\text{SDI} \geq 0.9$) conditions.

estimated errors in CBH are large for high-altitude clouds, data showing CBH more than 10 km are discarded.

Figure 9 shows the relationship that CBH has to CRF, DLR_{Obs}^{All} , and CRC at the four sites. CRF and CRC (averaged per 1000 m of height) show generally inverse correlations with CBH, as expected. Cloud effect varies widely due to other forcing parameters such as SDI or PW. For low clouds ($\text{CBH} < 2000$ m), the cloud effect is very different among the four sites. At TAT only, the cloud effect increases with increasing CBH below 2000 m. The larger the annual average of PW, the smaller the average of cloud effect.

Figure 10 shows CRF dependence on CBH under clear ($\text{SDI} < 0.5$; a), cloudy ($0.5 \leq \text{SDI} < 0.9$; b), and overcast ($\text{SDI} \geq 0.9$; c) conditions. Under nearly clear conditions, the effect of CBH is very small, and the average CRF is about $2\text{--}20 \text{ W m}^{-2}$. Increasing cloud amounts leads to wide dispersion in CRF. Under cloudy and overcast conditions, CRF decreases with increasing CBH from the surface to 10 km. Very high clouds are difficult to detect with lidar because of attenuation of the laser, and therefore the standard deviation of CRC at CBH above 5 km is very large. Under nearly clear conditions, CRF shows little change from changes in CBH, and the difference among sites is also small. At TAT, under cloudy and overcast conditions, average CRF increases with

increasing CBH when CBH is less than 2 km; this is the same as under the all-sky condition, which is contrary to the expectation that CRF should decrease with decreasing CBH.

Figure 11 shows the joint frequencies of annual mean and seasonal variation of the relationship between CBH and CRF at LIN, NYA, SAP, and TAT. When CBH is low, low clouds are dominant at all stations for annual mean (Fig. 11-1). Large CRF occurs frequently at LIN and NYA. When high CRF and low CBH have a high joint frequency, an inverse relationship between CBH and CRF is reasonable because of the gradual decrease in temperature with increasing altitude. In contrast, the CRF annual mean is often small at SAP and TAT stations. The low joint frequency of high CRF and low CBH weakens the inverse relationship between CBH and CRF, resulting in a directly proportional relationship.

Figures 11-2 (spring), 11-3 (summer), 11-4 (autumn), and 11-5 (winter) show seasonal variations in the relationship between CBH and CRF; the data for NYA in winter is excluded because radiosonde observations are not performed there during the daytime in winter. Seasonal variations at LIN and NYA are small. CBH below 2000 m occurs frequently at LIN throughout the year. When low clouds are present, CRF exhibits bimodal characteristics; CRF occurs frequently at values greater than 60 W m^{-2} and at small values around 0 W m^{-2} . The tendencies at

Fig. 9. Relationships between CBH and CRF (top), average (center), and CRC (bottom) at four sites. Observations are grouped by color: orange (TAT), green (SAP), blue (LIN), and purple (NYA). TAT, SAP, and LIN sites are in moderate humidity regions. NYA is in a dry region.

NYA are generally similar to those at LIN. The incidence of low clouds is high and CRF exhibits bimodal characteristics. NYA differs from LIN with respect to high clouds with CBH around 5000 m, which occur slightly more often at LIN.

At SAP and TAT, the joint distribution of CBH and CRF varies dramatically by season. High clouds are more frequent than low clouds during spring and summer at SAP and during autumn at TAT. Although these high and cold clouds are commonly thin, with small CRF, the incidence of CRF larger than 30 W m^{-2} is high for spring and summer at SAP. During autumn and winter at SAP, low clouds occur frequently, but the distribution of CRF differs from

Fig. 10. Relationships between CBH and CRF under (a) nearly clear ($\text{SDI} < 0.5$), (b) cloudy ($0.5 \leq \text{SDI} < 0.9$), and (c) overcast ($\text{SDI} \geq 0.9$) conditions at four sites. Orange, green, and blue colors represent TAT, SAP, and LIN data, respectively.

that observed during spring and summer. The incidences of low clouds with small CRF and low clouds with high CRF are approximately the same.

Fig. 11. Joint frequencies between CBH and CRF at (a) LIN, (b) NYA, (c) SAP, and (d) TAT. Periods shown are (1) annual, (2) spring (March to May), (3) summer (June to August), (4) autumn (September to November), and (5) winter (December to February). Frequency is indicated by brightness, with black indicating the highest frequency.

The relationship between CBH and CRF at TAT for summer and winter is quite different from the other stations and seasons. During summer, near-surface clouds ($\text{CBH} < 1000 \text{ m}$) dominate, and it is hard to detect clouds at other altitudes. The incidence of low-altitude clouds is high for CRF from near 0 to 60 W m^{-2} . The incidence of low- or middle-altitude clouds is high during winter. In Fig. 9, the annual average of CRF at TAT increases with increasing CBH when CBH is less than 2 km. This is attributed to the unusual relationship between CBH and CRF during summer and winter at TAT.

A positive correlation exists between CBH and CRF at TAT for very low clouds. Because low clouds can be treated as an optically thick black body, the positive correlation is not attributed to the effect of cloud properties (e.g., cloud optical thickness or liquid water path), but instead to atmospheric moisture conditions. Figure 12 shows the joint frequencies between CBH and PW at the four stations for annual averages (Fig. 12-1) and by season (Figs. 12-2, 12-3, 12-4, and 12-5 correspond to spring, summer, autumn, and winter, respectively).

At LIN, SAP, and TAT, clouds with CBH less than 2000 m occur more frequently than any other type. Only NYA has high incidences of both low and high clouds. Among low clouds, 0–1000 m clouds are more frequent than 1000–2000 m clouds at LIN and NYA; 1000–2000 m cloud is more frequent than 0–1000 m cloud at SAP and TAT. This difference may account for the difference in cloud height seasonal characteristics among stations.

Although a high incidence of only low clouds is obtained for all seasons at LIN, the frequency distributions can be classified into one of two types. During autumn and winter, the CBH below 1000 m dominates. Compared with near-surface clouds, CBH in the 1000–2000 m range is more frequent during spring and summer.

At NYA, the relationship between PW and CBH is consistent throughout the year: quite low clouds occur frequently; during spring and autumn, high clouds with small PW occur slightly more often, as expected from the relationship between CRF and CBH.

At SAP, the joint frequency between PW and CBH indicates large seasonal variations. Middle- and high-altitude clouds are found frequently in spring and summer, similar to that observed at NYA during spring and autumn. In summer, high clouds appear more frequently than low clouds. During autumn, low-altitude clouds appear during dry conditions; this is similar to the pattern at LIN. In winter, most clouds

detected occur below 2,000 m and under quite dry conditions.

At TAT, the joint frequency between CBH and PW shows large seasonal variations. During spring, the incidence of low clouds is high, similar to those observed in all seasons at LIN and in autumn at SAP. In summer, the incidence of near-surface clouds is high for PW from 20 mm to 60 mm. An incidence peak is attained under wet conditions. Both high and low clouds occur frequently at TAT during autumn, similar to that at NYA for spring and autumn and SAP for spring and summer. When high clouds are present, PW is lower than when low clouds are present. One aspect of the joint frequency between CBH and PW at TAT is unusual for winter: although low clouds and low PW frequently co-occur at other stations and seasons, clouds of all heights frequently occur under dry conditions at TAT.

Because TAT is located on the Pacific Ocean side of Japan and is affected by the East Asian monsoon, the wind direction changes with the season and determines atmospheric conditions and cloud characteristics (Jhun and Lee 2004; Hirano and Matsumoto 2010; Wang and Ho 2002). Figure 13 shows the relationship between CBH and surface wind direction at TAT for summer and winter from radiosonde observation data. During summer, the monsoon forms the Baiu Front, which causes heavy clouds and frequent rainfall on the Pacific Ocean side of the island, where TAT is situated. Figure 13a shows frequent north-eastern and southern winds; however, near-surface clouds occur frequently regardless of wind direction.

In contrast to summer, the winter monsoon season causes heavy snowfall on the Japan Sea side of Japan, while the Pacific Ocean side of Japan is dry with low precipitation; this occurs because the mountains that extend from the north to the south at the center of the Honshu island of Japan block humid winds. Figure 13b shows that two wind directions frequently co-occur with clouds. When the wind direction is northern or northeastern, only low clouds occur frequently; when the wind direction is western or northwestern, clouds at all heights occur frequently. These results indicate that wind direction corresponds to atmospheric conditions and affects the incidence of clouds. This relationship is particularly strong during the winter season.

4. Summary

We estimated the effect of clouds on DLR at the surface by using observed DLR with absolute (CRF) and relative (CRC) values under the all-sky condi-

Fig. 12. Joint frequencies between CBH and PW at four stations.

Fig. 13. Joint frequencies between CBH and surface wind direction at TAT for (a) summer and (b) winter seasons.

tion and the calculated DLR at eight BSRN stations. CRF and CRC varied widely, from -21 to 92 W m^{-2} and from -6% to 38% , respectively. The average of cloud radiative effect increased with increasing SDI and with decreasing PW. The distribution of the cloud effect varied greatly when the cloud cover was high and PW was small. Two results of this study are different from those of Stephens et al. (2012): large negative values of cloud radiative effect were observed in the polar region and were attributed to temperature inversions; and the joint distribution of cloud radiative effect and column water vapor amount was widely scattered owing to the use of in situ observations for near-surface atmosphere and cloud cover.

At four sites (NYA, LIN, SAP, and TAT), we compared CBH to the cloud radiative effect. Average CRC mostly showed an inverse relationship with CBH, although the average CRC of low clouds increased with increasing CBH when the CBH was less than 2 km at TAT. This phenomenon occurred because the unusual relationship between CBH and CRF for summer and winter seasons affects atmospheric humidity and CBH. The incidence of different cloud base heights was affected by wind direction. The present study used ground-based observation

targets, mainly low clouds and surface water vapor, which are difficult to estimate by satellite observation, and showed the variability in the low cloud effect, a pattern caused by regional and seasonal characteristics.

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