NOTES AND CORRESPONDENCE

Estimation of Precipitable Water and Relative Humidity over the Tibetan Plateau from GMS-5 Water Vapor Channel Data

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Abstract

The brightness temperature of the Geostationary Meteorological Satellite (GMS) -5 water vapor channel ($Tb_{6.7}$) and relative humidity, precipitable water, specific humidity and temperature observed by radiosondes over the Tibetan Plateau are compared for the summer of 1995. A wavelength of $6.7\ \mu\text{m}$ was employed, so most of the radiation sensed by the satellite comes from the atmospheric layer between 300 and 600 hPa. The plateau altitude is 3,000 to 4,000 m, and GMS-5 $Tb_{6.7}$ over the Plateau represents the moisture information above that. The regression equations of upper tropospheric humidity (UTH) and precipitable water with $Tb_{6.7}$ are presented in this paper. The relationships between $Tb_{6.7}$ and observed temperature and specific humidity at the standard pressure levels are then compared to each other. Results show that, over the Plateau, $Tb_{6.7}$ is sensitive to the moisture information around 200 to 300 hPa height. The results demonstrate that $Tb_{6.7}$ is useful for not only qualitatively analyzing the moisture transporting field over and around the Tibetan Plateau, but also quantitatively inferring UTH and precipitable water above 400 hPa (W_{400}). Estimates of W_{400} for July and August 1995 and its diurnal variability over and around the Tibetan Plateau are presented.

1. Introduction

Meteosat was the first geostationary satellite to make images of mid-to-upper-troposphere water vapor at $6.7\ \mu\text{m}$ in addition to visible and 10–12 μm infrared wavelengths. The Japanese Geostationary Meteorological Satellite (GMS) -5, which has a $6.7\ \mu\text{m}$ channel, has been operating since June 1995. At a wavelength of $6.7\ \mu\text{m}$, most of the radiation sensed by the satellite comes from the atmospheric layer between 300 and 600 hPa (Morel et al. 1978); thus it measures middle lev-

els of the troposphere. The water vapor channel sensor of Meteosat and the Television and Infrared Observation Satellite (TIROS) Operational Vertical Sounder (TOVS) have been used for deducing clear-sky, upper-tropospheric moisture or upper tropospheric humidity (UTH) (e.g., Poc et al. 1980; Soden and Bretherton 1993). This information is derived for cloud-free conditions, since if there are high clouds, radiation from the clouds is observed. Recently, Jedlovac and Atkinson (1996) derived a water vapor transport index (WVTI) by transforming UTH of GOES into specific humidity.

As in the above, water vapor channels of Meteosat or TOVS are quantitatively used as an indicator of relative humidity or water vapor content in the middle and upper troposphere. Here, the brightness temperature in the $6.7\ \mu\text{m}$ channel (denoted as $Tb_{6.7}$) of GMS-5, and moisture infor-

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mation over the Tibetan Plateau, are compared. Since the water vapor channel measures radiation of the atmospheric layer between 300 and 600 hPa and since the altitude of the plateau is about 3,000 to 4,000 m, $Tb_{6.7}$ over the plateau is expected to be related to the total moisture information there. The atmosphere over the Tibetan Plateau is very unique because warmer and wetter atmosphere exists in the middle troposphere compared with that in other upper tropospheres in the same latitude in summer. The Tibetan Plateau is very important and interesting, from the view point of meteorology and global water circulation because it is an elevated heat source. In spite of the importance, routine radiosonde observation is very poor over the plateau. Therefore, moisture information derived from GMS-5 should be useful for understanding hydrological circulation over and around there.

Here, we intend to clarify what $Tb_{6.7}$ represents over the plateau. $Tb_{6.7}$ is compared with observed temperature, specific humidity, relative humidity at the standard pressure levels, and precipitable water. Then we make regressions.

2. Data and Method

2.1 Data used in this study

This study uses the radiances measured in the $6.7\ \mu\text{m}$ ($6.5\sim 7.0\ \mu\text{m}$) and $11\ \mu\text{m}$ ($11.5\sim 12.5\ \mu\text{m}$) spectral bands. It is comprised of measurements taken daily at 00, 06, 12 and 18 UTC for July and August 1995 by the GMS-5, which covers 80°E to 160°W and 60°N to 60°S . The dataset was already geometrically corrected, and it was spatially averaged in 0.25° (latitude) $\times 0.25^\circ$ (longitude) resolution.

The 00 and 12 UTC radiosonde data at Lhasa

Fig. 1. Locations of Lhasa and Yushu. Contours indicate altitude of 1500, 3000, 4500 and 5500 m. Elevation higher than 3000 m are shaded.

(29.67°N , 91.13°E , 3,650 m asl) and Yushu (33.192°N , 97.02°E , 3,682 m asl) from the Global Telecommunications System (GTS) routine radiosonde network are used for July and August 1995. Figure 1 shows the location of these stations. Surface pressure (Ps), surface temperature (Ts), dew-point depression at surface ($TTDs$), the geopotential height (z), temperature (T), and dew-point depression (TTD) at seven standard pressure levels (500, 400, 300, 250, 200, 150, and 100 hPa) are used. To compute the specific humidity (q) and the relative humidity (R), the saturation vapor pressure is calculated by Goff-Gratch's equation (Goff and Gratch 1946). The precipitable water (W) is calculated by integrating q from Ps to 200 hPa height, because the data above 200 hPa is sometimes missing. Precipitable water above 500 hPa (W_{500}) and 400 hPa (W_{400}) is also computed.

Fig. 2. Weighting function of GMS-5 water vapor channel computed by "rstar" with standard summer atmosphere. Mean standard pressure levels (straight line), and the height of Lhasa (3,650 m, dashed line) are indicated. The mean summer pressure at Lhasa is around 650 hPa.

Table 1. Response function of GMS-5 IR3 (water vapor channel).

Wave Length [μm]	6.50	6.60	6.70	6.80	6.90	7.00	7.10	7.20	7.30	7.40
Response	.004	.057	.522	.677	.966	.858	.993	.327	.019	.000

2.2 Characteristics of the GMS-5 water vapor channel

Table 1 shows the response function of the GMS-5 6.7 μm channel. The weighting function is computed with an accurate and efficient radiative transfer scheme called "rstar" (Nakajima and Tanaka 1986, 1988) with LOWTRAN 7's atmospheric models (Kneizys et al. 1988). The standard summer temperature and the response function are used, and the radiance change with a change of 20% specific humidity every 1,000 m is computed. Figure 2 shows the plot of the weighting function with altitude. For comparison, the mean standard pressure level and the height of Lhasa are indicated. A strong response is observed with the change of moisture at 400 hPa. Due to the elevation of the plateau, the brightness temperature may be influenced by the surface temperature when the atmosphere is very dry. However, if it is wet, a strong relationship is expected between $Tb_{6.7}$ and vertically mean humidity or precipitable water.

The weighting function over the Tibetan Plateau may differ from Fig. 2 because of characteristics of

the vertical profile of temperature and humidity and the existence of ground surface in the middle troposphere over the plateau. However, a sufficient dataset over there has not been assembled yet, therefore in this study, observed q , t and R of each standard level, precipitable water, and mean R are compared with $Tb_{6.7}$ in order to clarify what $Tb_{6.7}$ of GMS-5 indicates.

3. Results

3.1 Comparison of $Tb_{6.7}$ with precipitable water

Figure 3 shows the time series of the $Tb_{6.7}$ and precipitable water computed at Lhasa. For $Tb_{6.7}$, a pixel including Lhasa is used. The time series of these two values match very well except for a very low $Tb_{6.7}$, which are due to clouds. They appear mostly between the evening (12 UTC) and midnight (18 UTC), which is consistent with the well-known phenomenon that strong convection is observed over Tibet from evening (e.g., Luo and Yanai 1983; Murakami 1983). A relatively dry condition is observed on 9 July, 18 July, 12 August, and 15 to 21 August in both panels. On hu-

Fig. 3. Time series of $Tb_{6.7}$ (upper panel) and precipitable water at Lhasa (lower panel) for July and August, 1995. Circles denote the observation at 00 UTC (Coordinated Universal Time) and dots are at 12 UTC. White and black triangles of the upper panel are at 06 and 18 UTC, respectively.

mid days, $Tb_{6.7}$ of 00 and 06 UTC are generally lower than those of 12 and 18 UTC, which shows that more humid conditions or cloud are observed through evening and midnight.

Figure 4 shows the scatter diagram between $Tb_{6.7}$ and Tb_{11} at Lhasa and Yushu. According to Ueno (1997), the threshold value of detecting convective clouds over the Tibetan Plateau using Tb_{11} is 240 K. This value is indicated in this figure as a straight line. It corresponds to 228 K for $Tb_{6.7}$. Figure 5a shows scatter diagrams between the two data sets shown in Fig. 3. These two thresholds match well. The 228 K of $Tb_{6.7}$ seems to be a threshold value for detecting convective cloud areas. The precipitable water when $Tb_{6.7}$ is under 228 K does not vary very much. The inclinations of the plots differ above and below 228 K, and $Tb_{6.7}$ of warmer than 228 K and precipitable water seem to have a linear relationship. The pixels warmer than 228 K seems to show moisture information, although lower level cloud and cloud smaller than 1 pixel size may exist. Though we cannot say it is completely cloud-free radiance, when we only eliminate pixels colder than 228 K, however, empirically we may consider it shows moisture information. Assuming $Tb_{6.7}$ under 228 K indicates convective clouds and eliminating these data, the correlation of these two

Fig. 4. Comparison of $Tb_{6.7}$ and Tb_{11} at Lhasa and Yushu. The threshold value of cloudiness, 240 K by Ueno (1997) is indicated in this figure as a solid line, while the estimated threshold of $Tb_{6.7}$ (228 K) is shown with a dashed line.

Fig. 5. a) Scatter diagram of $Tb_{6.7}$ and precipitable water at Lhasa. Regression is indicated with a solid line. Threshold of cloud region is shown with a dashed line. b) The same as a) but for Yushu.

at Lhasa is $r = -0.483$ (79 samples), and that at Yushu is -0.555 (99 samples). The regression of precipitable water is shown in the figure. When an exponential of W is used, correlation coefficients slightly increase for both stations. In these scatter diagrams, data at 12 UTC are more in cloud region. Data at 00 UTC are more highly correlated than at 12 UTC. One reason for the low correlation at 12 UTC may be the effect of the surface.

Over the Tibetan Plateau, the surface is strongly heated and the daily variation of temperature is very large. $Tb_{6.7}$ is thus contaminated by the surface temperature. Therefore, total moisture above 500 hPa or 400 hPa, namely not including the sur-

face, should be examined. It might correspond to $Tb_{6.7}$ more accurately than the total precipitable water from the surface. Since it is difficult to separate the information of temperature and specific humidity from $Tb_{6.7}$, diurnal variation should be analyzed carefully. Next, the correlation between $Tb_{6.7}$ and humidity of each level (q_{500} , q_{400} , q_{300} , q_{250} , q_{200} , R_{500} , R_{400} , R_{300} , R_{250} and R_{200}), temperature of each level (t_{500} , t_{400} , t_{300} , t_{250} and t_{200}), the upper tropospheric precipitable water (W , W_{500} , W_{400}) and the upper tropospheric humidity ($R_{300-500}$, $R_{250-500}$, $R_{250-400}$ and $R_{200-300}$), respectively, are shown.

Table 2. Correlation coefficients between $Tb_{6.7}$ and specific humidity (q), temperature (t), relative humidity (R), precipitable water (W) at Lhasa and Yushu, respectively. Numbers with parameters indicate pressure levels. Shade indicates correlation coefficients beyond -0.7 . Numbers in parentheses indicate the respective number of samples.

Parameters	$Tb_{6.7}$ (00,12Z) vs. Sonde (Lhasa)	Ln(para- meter) vs. Sonde (Lhasa)	$Tb_{6.7}$ (00,12Z) vs. Sonde (Yushu)	Ln(para- meter) vs. Sonde (Yushu)
q_{500}	-0.419 (80)	-0.416	-0.478 (100)	-0.481
q_{400}	-0.649 (79)	-0.604	-0.611 (100)	-0.634
q_{300}	-0.658 (80)	-0.898	-0.747 (95)	-0.914
q_{250}	-0.693 (80)		-0.744 (93)	
q_{200}	-0.719 (78)		-0.766 (92)	
t_{500}	0.160 (80)	0.160	0.026 (100)	0.025
t_{400}	-0.363 (80)	-0.364	-0.437 (100)	-0.437
t_{300}	-0.274 (80)	-0.274	-0.226 (95)	-0.226
t_{250}	-0.253 (80)	-0.253	-0.212 (93)	-0.212
t_{200}	-0.071 (78)	-0.093	-0.011 (92)	-0.011
t surface (T_s)	0.110 (79)	0.108	0.044 (99)	0.038
Precipitable Water (W)	-0.483 (79)	-0.467	-0.555 (99)	-0.569
P. W. above 500 hPa (W_{500})	-0.682 (79)	-0.646	-0.676 (100)	-0.654
P. W. above 400 hPa (W_{400})	-0.675 (80)	-0.916	-0.764 (95)	-0.910
R_{500}	-0.396 (80)	-0.408	-0.486 (100)	-0.481
R_{400}	-0.556 (79)	-0.526	-0.507 (100)	-0.456
R_{300}	-0.619 (80)	-0.602	-0.824 (95)	-0.810
R_{250}	-0.684 (80)	-0.692	-0.848 (93)	-0.839
R_{200}	-0.768 (78)	-0.762	-0.847 (92)	-0.824
R mean 300-500 hPa ($R_{300-500}$)	-0.586 (79)	-0.573	-0.723 (95)	-0.706
R mean 250-500 hPa ($R_{250-500}$)	-0.631 (79)	-0.618	-0.786 (93)	-0.771
R mean 250-400 hPa ($R_{250-400}$)	-0.679 (79)	-0.665	-0.813 (93)	-0.791
R mean 200-300 hPa ($R_{200-300}$)	-0.725 (78)	-0.731	-0.868 (92)	-0.844

3.2 $Tb_{6.7}$ compared to radiosonde data of each pressure level

Table 2 shows the correlation coefficients between cloud-free $Tb_{6.7}$ and specific humidity, temperature of each level, surface temperature (T_s), precipitable water and relative humidity. $Tb_{6.7}$ may approximate the relative humidity. Less radiance means either that the atmosphere is colder at the same level, or that there is more water vapor. In either case the high relative humidity is likely to be in the colder temperature case. Here, several mean relative humidity values are also compared with $Tb_{6.7}$.

Among these variables, the natural logarithm of precipitable water above 400 hPa height ($\ln(W_{400})$) shows the highest negative correlation with $Tb_{6.7}$. Furthermore, the relative humidity and specific humidity around 200, 250 and 300 hPa are highly correlated. According to the weighting function (Fig. 2), a strong response is observed with the change of moisture at 400 hPa for standard summer temperature. One reason that R or q of 400

hPa is less related with those at levels higher than 400 hPa is the existence of a wet atmosphere over the plateau. Another reason may be contamination of the surface temperature up to 500 and 400 hPa height. Figure 6 shows scatter diagrams of the surface temperature, relative humidity and specific humidity of 400 hPa height at Lhasa and Yushu. The scatter patterns of $Tb_{6.7}$ and T_s are completely different between 00 UTC and 12 UTC (Figs. 6a and b). The T_s at 12 UTC is sometimes more than 10 degrees above that of 00 UTC, and it affects temperature at 400 hPa (T_{400}). R and q at 400 hPa also exhibit somewhat different distributing patterns between 00 UTC and 12 UTC, especially in R_{400} at Lhasa (Fig. 6c). At 500 hPa, the difference is clearer, while at 300 hPa it is unclear (not shown).

3.3 Regression with upper tropospheric humidity and precipitable water above 400 hPa

According to Soden and Bretherton (1993), $Tb_{6.7}$ is proportional to the natural logarithm of the ap-

Fig. 6. a) Scatter diagram of $Tb_{6.7}$ and surface temperature at Lhasa. Threshold of cloud region is shown with a dashed line. b) The same as a) but for Yushu. c) The same as a) but for relative humidity at 400 hPa. d) The same as c) but for Yushu. e) The same as a) but for specific humidity at 400 hPa. f) The same as e) but for Yushu.

appropriate vertical average of the relative humidity divided by the cosine of the viewing zenith angle. Since the Tibetan Plateau is located in the edge of GMS images, the pathway should be corrected using satellite zenith angle θ . Here, UTH is obtained by simply averaging the values of R_{300} , R_{250} and R_{200} , which corresponds to the bottom line of Table 2. Figure 7a shows the scatter diagram between $Tb_{6.7}$ and $\ln(R_{200-300}/\cos\theta)$. This shows $Tb_{6.7}$ is strongly correlated with $\ln(R_{200-300}/\cos\theta)$ under cloud free conditions. The value of 12 UTC is less than that of 00 UTC at the same $Tb_{6.7}$. Therefore, we separate the regression for 00 UTC and 12 UTC. These are

$$\ln(R_{200-300}/\cos\theta) = -4.93 \times 10^{-2} Tb_{6.7} + 16.3 \quad (00 \text{ UTC}), \quad (1)$$

and

$$\ln(R_{200-300}/\cos\theta) = -4.99 \times 10^{-2} Tb_{6.7} + 16.4 \quad (12 \text{ UTC}), \quad (2)$$

The correlation coefficients are -0.758 and -0.734 , respectively.

The regression with precipitable water above 400 hPa height is also examined. Figure 7b is the scatter diagram between $Tb_{6.7}$ and $\ln(W_{400}/\cos\theta)$. In this case the difference between 00 UTC and 12 UTC is not clear in comparison to Fig. 7a. The regression of cloud free conditions from both times is

$$\ln(W_{400}/\cos\theta) = -6.52 \times 10^{-2} Tb_{6.7} + 15.7. \quad (3)$$

The correlation coefficient is -0.710 .

3.4 Estimates of upper tropospheric water vapor for the summer 1995

Using Eqs.(1), (2) and (3), the upper tropospheric mean relative humidity and precipitable water above 400 hPa are estimated by GMS $Tb_{6.7}$ and the satellite zenith angle for July and August (summer) 1995. Figure 8 shows (a) the mean cloud free W_{400} for the summer and (b) the difference of the summer mean W_{400} at 12 UTC and 00 UTC. Wet atmosphere prevails from the Indo-china peninsula to the eastern part of the Tibetan Plateau. In fig. 10b, the difference from 12 UTC (evening) to 00 UTC (morning) is quite interesting. This figure implies that W_{400} is higher in the evening over the central, the northeast and along the southern periphery of Tibet, which corresponds to the

Fig. 7. a) Scatter diagram between $Tb_{6.7}$ and $\ln(R_{200-300}/\cos\theta)$. Regressions are indicated with a solid line. Threshold of cloud region is shown with a dashed line. b) Scatter diagram between $Tb_{6.7}$ and $\ln(W_{400}/\cos\theta)$. Regression is indicated with a solid line. Threshold of cloud region is shown with a dashed line.

(a) Estimated Precipitable Water

(b) P.W. 12utc-00utc

Fig. 8. a) Monthly mean precipitable water upper level of 400 hPa (W_{400}) in July and August 1995 (unit: mm). b) Difference between 12 UTC and 00 UTC of W_{400} in July and August 1995 (unit: mm). White contours indicate altitude of 3,000 and 4,000 m, and pink contour indicates altitude of 5,000 m.

higher altitude region over the plateau (see Fig. 2). The pattern of diurnal variation of W_{400} over the plateau aligns east-west direction, which corresponds to major mountains and valleys on the Tibetan Plateau aligned along east-west direction. Around the plateau within about 500 km, namely, over the southern periphery of the Himalayas, eastern Tibet, and around the Taklimakan desert, W_{400} is higher in the morning. Since W_{400} is an index of up-welling motion, this figure shows a difference in the diurnal variation of up-welling motion and related moisture content between central Tibet and the surrounding area. The similar characteristics are seen with UTH patterns (figures not shown) derived from Eqs.(1) and (2).

4. Concluding Remarks

The relationship between $Tb_{6.7}$ and moisture information over the Tibetan Plateau in summer was investigated using radiosonde data. $Tb_{6.7}$ was proportional to the natural logarithm of around 200 to 300 hPa average of the relative humidity divided by the cosine of the viewing zenith angle. A strong relationship is also found between $Tb_{6.7}$ and precipitable water above 400 hPa height (W_{400}) over the Tibetan Plateau. The weighting function of GMS-5 $Tb_{6.7}$ showed that it is sensitive to around 400 hPa atmosphere. However, over the plateau, the relationship was a little bit different. A stronger relationship was seen around 200 to 300 hPa. The surface temperature contaminates the moisture information of $Tb_{6.7}$, especially at 12 UTC around the 500 to 400 hPa height levels.

Using the obtained regression equation, the monthly mean horizontal distribution of W_{400} over and around the plateau in July and August 1995 was displayed. As an example, the diurnal variation of up-welling motion over and around the Tibetan Plateau was indicated. Since radiosonde observation of the upper troposphere is very limited, objective analysis data highly depend on model characteristics in such an area. Therefore, information from independent (not assimilated) stationary satellite is very useful for investigating diurnal variation, and moisture advection. The detailed structure of moisture (UTH) intrusion can be seen if we look at some snap shots (not shown).

During the GEWEX Asian Monsoon Experiment (GAME) intensive observation period (IOP), radiosonde data were available four times each day at several stations (GAME International Science Panel, 1998). Combining this data and GMS-5

data may produce a spatial pattern of the daily variations of precipitable water or UTH. Comparing T and R of the objective analysis data produced by the Japan Meteorological Agency four dimensional data assimilation (4DDA) system (GANAL) with those of radiosondes, GANAL underestimates W and Ts , and the diurnal circulation derived by GANAL is smaller than that derived by radiosonde (Yatagai 1999). GANAL has not included moisture information of GMS-5 or TOVS yet. UTH of GMS-5 may affect important information to assimilated data because TOVS does not have enough information of diurnal variation.

Since GMS-5 covers about 80°E to the east, the Tibetan Plateau exists at the western edge. The slant path effect should be investigated. Foreshortening also occurs in this region. If the FY-II satellite can be used, the water vapor channel of the satellite will yield more precise, wide-coverage information.

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