Estimation of Land Surface Temperature over the Tibetan Plateau Using GMS Data

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ABSTRACT

Geostationary Meteorological Satellite Visible/Infrared Spin-Scan Radiometer (GMS VISSR) images have been used to estimate diurnal variations of land surface temperature distributions over the Tibetan Plateau. The infrared split-window algorithm developed for NOAA Advanced Very High Resolution Radiometer (AVHRR) has been adapted for this purpose. Radiative transfer simulations are carried out to obtain the atmospheric transmittances and the difference temperatures that are involved in the internal coefficients of the split-window algorithm. Precipitable water distribution that is required by this algorithm is estimated from 6.7-μm brightness temperature utilizing spectral characteristics of the GMS water vapor channel. Cloud removal plays an important role in the surface temperature retrieval process. To identify convective cloud activity, many researchers use satellite infrared measurements with a fixed threshold technique. In this study, it is necessary to remove not only convective clouds but also warm clouds. For this purpose, a variable threshold technique is proposed. The threshold varies both seasonally and diurnally, and its value is determined on the basis of surface observations. With a variable threshold, it becomes possible to remove relatively warmer clouds in summer and detect colder ground surfaces at nighttime in the winter. The results of comparing estimated surface temperature from GMS data using this algorithm with in situ surface measurements show correlations around 0.8.

1. Introduction

The Tibetan Plateau is thought to play an important role in the progress of the Asian summer monsoon as an elevated heat source/sink protruding into the middle of the troposphere. To understand the interactions between the land surface and the atmosphere over the Tibetan Plateau in the context of Asian monsoon research, intensive meteorological observations were conducted in the Global Energy and Water Cycle Experiment (GEWEX) Asian Monsoon Experiment (GAME) Tibet project (Koike et al. 1999). During GAME/Tibet, surface sensible and latent heat fluxes were measured at a set of sites using eddy covariance techniques together with monitoring of relevant surface parameters such as land surface temperature, soil moisture, net radiation, and surface albedo (Tsukamoto et al. 1999). From GAME/Tibet surface measurements, the plateau boundary layer was found to be characterized by a strong diurnal cycle (Tanaka et al. 2001). Although the land surface–atmosphere interaction was observed experimentally at these sites, it only represents fluxes over a patch scale around the measurement sites. Patch-scale knowledge needs to be integrated with a regional-scale understanding of the plateau. Remote sensing from satellites offers the possibility of determining regional distributions of various surface meteorological properties in combination with a sparsely distributed set of field experiment stations. Wang et al. (1995) and Ma et al. (2002) estimated the distributions of land surface variables over the Heihe basin field experiment (HEIFE) area by combining Landsat Thematic Mapper (TM) data with field observations. Similarly estimations were made over the GAME/Tibet enhanced observation area by Ma et al. (2003) using National Oceanic and Atmospheric Administration Advanced Very High Resolution Radiometer (NOAA AVHRR) data. However, strong diurnal variations are one of the outstanding features of the plateau surface that has been revealed by field observations, which cannot be detected in the data of polar-orbiting satellites.

To measure the diurnal cycle, the continuous data stream of a geostationary satellite is therefore required. For the Tibetan Plateau, Geostationary Meteorological Satellite-5 (GMS-5) provides continuous information over three infrared channels, two split-window (IR1, 11 μm; IR2, 12 μm) and one water vapor channel (WV, 6.7 μm).

In this paper, a method for the retrieval of one of the important surface parameters, land surface temperature, from GMS data is presented. The retrieval algorithm used for NOAA AVHRR data is applied to GMS-5 split-window measurements with some modifications. Figure 1 describes the method with a process flowchart.
2. Land surface temperature calculation

Algorithms for determining surface temperatures from the radiation data collected by earth observation satellites have been proposed by many authors, including Wang et al. (1995), Sobrino et al. (1994), Ma et al. (2003) calculated the surface temperature distribution of a 100 km × 200 km region over the Tibetan Plateau using the split-window technique applied to the infrared brightness temperatures of NOAA AVHRR images. The same retrieval algorithm is applied to GMS Visible/Infrared Spin-Scan Radiometer (VISSR) data. As shown by the comparison in Fig. 2, the filter functions of GMS-5 VISSR and NOAA-14 AVHRR2 split windows are considerably different. The radiative transfer code, MODTRAN (Berk et al. 1989), is useful to simulate the radiance that would be measured by a satellite sensor and can be run for a given path geometry in a particular atmosphere defined from meteorological data provided by actual measurements. Therefore, a number of filter function parameters for GMS-5 VISSR radiation data were calculated using MODTRAN.

a. Data

GMS-5 VISSR measures four spectral bands—the visible band, split infrared bands (11 and 12 μm), and water vapor band (6.7 μm)—with images transmitted hourly to the Disaster Prevention Research Institute of Kyoto University. Data for the 11-, 12-, and 6.7-μm brightness temperature in grid format over longitude and latitude having a resolution of 0.1° were interpolated from original image data that are archived in line-pixel format. The instantaneous field of view of GMS-5 VISSR is about 5 km for these three channels at the subsatellite point. Since GMS-5 was being operated in a geostationary orbit at 140°E, the satellite zenith angle from the Tibetan Plateau is more than 60° so that the actual field of view of the plateau is between 7 and 10 km.

NOAA AVHRR images, as collected from the NOAA Satellite Active Archive Web site, are also used to estimate surface emissivity because it is not possible to derive emissivity from GMS data (for details, see section 2c). Four narrow bands of NOAA data were used (channel 1: 0.58–0.68 μm, channel 2: 0.73–1.10 μm, channel 4: 10.3–11.3 μm, and channel 5: 11.5–12.5 μm). Atmospheric profiles of the plateau were taken from radiosonde observations made at Amdo during the GAME/Tibet intensive observation period (IOP), from June to August 1998. Operational radiosonde data are also used. Surface temperature data measured by IR thermometer in Automatic Weather Stations (AWSs) in 1998 is used for calibration and verification. The locations of observation points are shown in Fig. 3, with more detailed information given in Table 1.

b. Algorithm

The algorithm proposed by Sobrino et al. (1996) for NOAA AVHRR data is followed in the derivation of an algorithm for GMS data. For a cloud-free atmosphere under local thermodynamic equilibrium, the radiative transfer equation gives the spectral radiance $I_{\nu \theta}$ measured from space by a satellite at a zenith observation angle $\theta$ and is expressed as

$$I_{\nu \theta} = B_{\nu}(T_{\nu \theta}) = \varepsilon_{\nu \theta} B_{\nu}(T_{\text{sfc}}) \tau_{\nu \theta} + B_{\nu}(T_{\text{atm}}) \tau_{\nu \theta} + R_{\nu \theta}(T_{\nu \theta})$$

where $B_{\nu}$ is Planck’s function, $T_{\nu \theta}$ is brightness temperature, $\varepsilon_{\nu \theta}$ is ground surface emissivity, $B_{\nu}(T_{\text{sfc}})$ is the radiance that would be measured if the surface were
blackbody with a surface temperature of \(T_{sfc}\), \(\tau_{atm}\) is total atmospheric path transmittance, \(R_{atm}^{ref}\) is upwelling radiance from the atmosphere, and \(R_{ref}^{atm}\) is atmospheric radiance reflected from the surface. For convenience, radiance from the atmosphere \(R_{atm}^{ref}\) is assumed to be

\[
R_{atm}^{ref} = (1 - \tau_{atm})R_{sfc}(T_{sfc}).
\]

where \(T_{sfc}\) is the mean temperature of the atmosphere between the surface and top of the atmosphere. For \(R_{ref}^{atm}\), isotropic sky radiance is assumed, and downwelling atmospheric radiance is taken to be the radiance emitted by the atmosphere in the 53° direction, that is,

\[
R_{ref} = (1 - \epsilon_{atm})(1 - \tau_{sfc})R_{sfc}(T_{sfc}).
\]

Equation (1) can be applied to the 11- and 12-μm channels and can be rewritten as

\[
B(T_{IR1}) = \epsilon_1 B(T_{sfc}) \tau_1 + [(1 - \tau_1) + (1 - \epsilon_1)(1 - \tau_{sfc})] B(T_{air})
\]

and

\[
B(T_{IR2}) = \epsilon_2 B(T_{sfc}) \tau_2 + [(1 - \tau_2) + (1 - \epsilon_2)(1 - \tau_{sfc})] B(T_{air}),
\]

where subscripts 1 and 2 represent the 11- and 12-μm channels, respectively, and \(\tau_{sfc}\) is the transmittance at a zenith angle of 53°. By applying the Taylor expansion to Eqs. (4)–(5) and solving for \(T_{sfc}\) gives

\[
T_{sfc} = T_{IR1} + A(T_{IR1} - T_{IR2}) - B - C(1 - \epsilon_1) - D\Delta \epsilon,
\]

where \(\epsilon = (\epsilon_1 + \epsilon_2)/2\) is the average emissivity over both channels and \(\Delta \epsilon = (\epsilon_1 - \epsilon_2)\) is the spectral variation in emissivity (Li and Becker 1993). The coefficients \(A, B, C,\) and \(D\) are given by

\[
A = \frac{1 - \tau_1}{\tau_1 - \tau_2},
\]

\[
B = A(1 - \tau_1)(T_{air} - T_{air}),
\]

\[
C = \frac{1 - \tau_2 \tau_{sfc}}{\tau_1 - \tau_2} (T_{IR1} - T_{IR2}) + \tau_{sfc} T_{IR1} \frac{\Delta \epsilon}{4.667}
\]

and

\[
D = \tau_2 AC.
\]
Fig. 4. Atmospheric transmittance for GMS VISSR 11-μm band \( \tau_1 \) (circle), and 12-μm \( \tau_2 \) (dot) as a function of precipitable water \( W \) with regression curve.

In Eq. (6), \( T_{sfc} \) is expressed in terms of the brightness temperature of the split windows, \( T_{IR1} \) and \( T_{IR2} \). To estimate \( T_{sfc} \), it is necessary to know \( \epsilon, \Delta \epsilon, \tau_1, \tau_2, \tau_{53} \), and \( T_{1air} - T_{2air} \). These parameters are derived from the results of radiative transfer simulations using MODTRAN as described later.

c. Surface emissivity

The procedure for estimating \( \epsilon \) and \( \Delta \epsilon \) is same as Sobrino and Raissouni (2000), in which they are derived from the normalized difference vegetation index (NDVI) calculated from NOAA visible channels 1 and 2. GMS VISSR also has a visible sensor, but it is not split window like that of NOAA, so that NDVI cannot be derived from GMS data. Surface emissivity depends on surface characteristics such as vegetation, so that diurnal variations are expected to be relatively small. Therefore, surface emissivities can be estimated from NOAA measurements. NDVI is defined as

\[
\text{NDVI} = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1},
\]

where \( \rho_1 \) and \( \rho_2 \) represent the surface reflectance measured in NOAA AVHRR channels 1 and 2, respectively. In this study, the distribution of NDVI over the plateau is estimated every 10 days at a 0.1° resolution.

d. Atmospheric transmittance

Path transmittances depend on the vertical profile of trace gases in the atmosphere such as water vapor and carbon dioxide. Among these gases, while spatial and temporal variations in water vapor are significant, other atmospheric components can be assumed to be constant. Many authors assume that fluctuations in atmospheric attenuation depend only on precipitable water, which is a vertical integration of water vapor. This assumption is followed, with asymptotic formulas for atmospheric transmittance derived as a function of precipitable water. During the GAME/Tibet IOP in 1998, intensive radiosonde observations were conducted from June to August. More than 300 vertical profiles of pressure, temperature, and relative humidity were obtained at Amdo. Atmospheric transmittances \( \tau_1, \tau_2, \) and \( \tau_{53} \) are computed by a radiative transfer code, MODTRAN, for each of these profiles. For profiles of atmospheric gaseous components, the typical midlatitude profiles are selected from the MODTRAN options. In addition to these atmospheric data, geographical and surface conditions at Amdo—such as the zenith angle from GMS and surface emissivity—are input to MODTRAN. Considering the filter function of the GMS-5 split-window channels, transmittance is computed for discrete wavenumbers across the range of the filter function, from which total transmittance for each channel is computed. Results for \( \tau_1 \) and \( \tau_2 \) are plotted as a function of precipitable water in Fig. 4. Smaller atmospheric transmittance can be seen in areas of higher precipitable water, as expected. Originating in the characteristics of the sensor, \( \tau_2 \) tends to be lower than \( \tau_1 \), with differences between \( \tau_1 \) and \( \tau_2 \) becoming remarkably large as precipitable water increases.

e. Atmospheric temperature

The term for the difference in atmospheric temperature measured by a split window at nadir, \( T_{1air} - T_{2air} \), is vanishingly small when precipitable water is less than 2.0 g cm\(^{-2}\) (Sobrino et al. 1996), with the coefficient of this term, \( B \), becoming significant only when precipitable water is 2.0 g cm\(^{-2}\) or more. Ma et al. (2003) omitted the term completely when using NOAA data at relatively small satellite zenith angles because precipitable water is usually less than this threshold value at
small angles. In this study, however, this term is included because the zenith angle of GMS is more than 50° from the plateau, increasing absorption pathlength and resulting in greater atmospheric attenuation. Using a method similar to the estimation of atmospheric transmittance, $R_{\text{atm}}(\theta_u)$ is simulated by MODTRAN. From $R_{\text{atm}}$ and $\tau_{\text{atm}}$, $B_n(T_{\text{air}})$ is derived with the aid of Eq. (2). For each channel, $11$ and $12$ $\mu$m, $T_{\text{air}}$ and $T_{\text{air}}$ are calculated using Planck’s function and the filter function of the sensor. The term $T_{\text{air}} - T_{\text{air}}$ and the coefficient $B$ are shown as a function of precipitable water in Fig. 5. Because of the difference in water vapor absorption between the split-window sensors, brightness temperature sensed by an $11$-$\mu$m sensor is higher than that by a $12$ $\mu$m. Thus, the coefficient $B$ should be taken into consideration in GMS measurements when precipitable water is greater than $0.5$ g cm$^{-2}$.

f. Zenith angle dependency

The target area of this study, the Tibetan Plateau, extends from $80^\circ$ to $100^\circ$E, with the satellite zenith angle ranging from $52.5^\circ$ to $73.5^\circ$ (see Fig. 1). At the larger satellite zenith angles, atmospheric attenuation is expected to increase with the increased absorption pathlength. Therefore, atmospheric transmittance over the plateau needs to be determined as a function of both precipitable water and satellite zenith angle. Radiative transfer was calculated for satellite zenith angles ranging from $52.5^\circ$ to $73.5^\circ$ at $3^\circ$ intervals. For each result, regression curves were estimated and coefficients $A$ to $D$ of Eq. (6) are derived as functions of both precipitable water and satellite zenith angle, as shown in Fig. 6.

g. Precipitable water

In the previous section, formulas for the computation of surface temperatures from the split-window channels of GMS-5 were derived. To use these algorithms, however, the distribution of precipitable water is needed. Yatagai (2001) showed that there is a strong relationship between $6.7$-$\mu$m brightness temperature $T_{\text{WV}}$ and precipitable water $W$ over the Tibetan Plateau. However, $6.7$-$\mu$m brightness temperature is usually a poor indicator of precipitable water because the weighting function of the $6.7$-$\mu$m sensor has a maximum near $400$ hPa and does not reflect humidity at lower levels. Over the Tibetan Plateau, fortunately, the ground elevation is over $4000$ m MSL and so this is not a problem. Figure 7 shows the scatter diagram between $6.7$-$\mu$m brightness temperature $T_{\text{WV}}$ and precipitable water $W$ calculated from radiosonde data observed at Amdo. Dots corre-

![Fig. 6. Distribution of the coefficients (a) $A$, (b) $B$, (c) $C$, and (d) $D$ of Eqs. (7)-(10) as a function of both precipitable water $W$ (horizontal axis) and satellite zenith angle $\theta$ from GMS (vertical axis).]
3. Cloud removal

Although surface temperature determination is straightforward over cloud-free areas, land surface temperature over cloudy areas cannot be estimated because Eq. (1) is only applicable to radiation reaching the satellite from the surface, not from the cloud top. To derive land surface temperature accurately, cloud-covered areas need to be correctly identified. Although cloud detection using visible channel data (hereinafter referred to as CD0) is relatively easy, visible data are not available at night.

Infrared data are useful for the detection of clouds, even at night. For example, to detect and estimate the activity of convective cloud over the equator, Nitta and Sekine (1994) utilized 11-μm brightness temperature $T_{IR1}$ and defined convective index $I_c$ as

$$I_c = T^* - T_{IR1} \quad \text{for} \quad T_{IR1} < T^*$$

$$I_c = 0 \quad \text{for} \quad T_{IR1} = T^*,$$

where $T^*$ is constant threshold. Since 400 hPa generally corresponds to an air temperature of 250 K, a value of $T^* = 250$ K was used. Thus $I_c$ represents an index for deep convective cloud with top height in excess of about 400 hPa. Areas in which $T_{IR1} < T^*$ are then taken to be cloudy, with this technique referred to as CD1. Ueno (1997) used CD1 with $T^* = 240$ K over the Tibetan Plateau during the monsoon period to investigate convective activity in clouds. CD1 is very effective for detecting deep convection, but sometimes misses shallow convection and layered clouds.

According to Tanaka et al. (2001), the surface temperature observed on the plateau at midnight in winter sometimes drops below 240 K, with a diurnal range
Fig. 10. Seasonal (horizontal axis) and diurnal (vertical axis) variation of land surface temperature. Data were obtained from hourly regression curves, such as Fig. 9.

exceeding 30 K. Figure 9 shows annual variations in surface temperature at specific hours as observed by AWSs locating roughly along a north–south line near 92°E. Local standard time (LST) is 6 h ahead of coordinated universal time (UTC). The solid curve in the figure shows a Fourier transform regression. Regression curves are obtained every hour. Placing these 24 curves in the direction of the ordinate, a yearday–UTC map of surface temperature is constructed as shown in Fig. 10. The figure shows annual (horizontal axis) and diurnal (vertical axis) variations of typical surface temperature $T_{sfc}$ on the plateau. The parameter $T_{sfc}$ reaches highs of 300 K at 1200 LST in the beginning of June, and lows of 240 K at 0700 LST in the end of January. Annual variations range from 28 K at 1200 LST to 41 K at 0700 LST, and diurnal variations range from 18 K during the postmonsoon period to 32 K during the premonsoon period. As a result, CD1 sometimes misidentifies cold ground surface as cloud or relatively low cloud as ground surface. A new variable threshold cloud detection technique (CD2) is therefore necessary in which threshold temperature is adjusted for large seasonal and diurnal variations in surface temperature.

Threshold temperatures are determined from the typical surface temperature $T_{sfc}$ shown in Fig. 10. If

$$ T_{IR1} < T_{sfc} - \Delta T $$

an area is assumed to be cloudy. Thus, instead of the

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Fig. 9. Seasonal variation of land surface temperatures from measurements at four AWSs (D66, Tuotuoh, D110, MS3608) locations at (a) 0000, (b) 0600, (c) 1200, and (d) 1800 UTC in 1998 observed by infrared radiation thermometer. The solid line is regression curve estimated using Fourier transforms.
constant threshold $T^*$ of CD1, a time-varying threshold $T^*_{st} - \Delta T^*$ is applied to CD2, with the performance of the method depending on the value of $\Delta T^*$.

To determine the optimum value of $\Delta T^*$, cloudy and cloud-free areas are first defined using NOAA visible channel 1. The distinction between cloudy and cloud-free areas is then based on reflectance,

\[
\rho_1 > \rho^*_1: \text{cloudy,} \\
\rho_1 \leq \rho^*_1: \text{cloud free,}
\]

where $\rho^*_1$ is a threshold value. To estimate the accuracy of this identification procedure, the correlation between the 11-μm brightness temperature of NOAA channel-4 $T_4$ and surface temperatures $T_{sfc}$ observed by AWSs was computed. A high correlation indicates effective cloud identification by the CD0 method. This procedure was repeated on all data available for 1998 for various threshold values of $\rho^*_1$, with the results shown in Fig. 11 in which the correlation coefficient and relative frequency of cloud identifications are plotted as a function of $\rho^*_1$. As $\rho^*_1$ increases, cloud frequency can be seen to decrease. The correlation coefficient reaches a maximum at $\rho^*_1 = 0.12$, meaning that CD0 is good for identifying cloud-free areas. Cloud detection is therefore assumed to be accurate using $\rho^*_1 = 0.12$ as a threshold for CD0.

An optimal value of $\Delta T^*$ is then investigated assuming CD0 is accurate when $\rho^*_1 = 0.12$. In AWS sites, the difference between $T^*_{st}$ and $T_4$ and differences between identification by CD0 and CD2 are expected to be minimized by using an optimal $\Delta T^*$. An example of such a comparison between $T^*_{st}$ and $T_4$ is shown in Fig. 12 for D110-AWS. Stars indicate areas identified as cloud free by both CD0 and CD2, squares indicate areas identified as cloudy by both techniques, and dots indicate areas of identification mismatches. Black dots indicate areas that were identified as cloudy by CD0 but as cloud free by CD2, and vice versa for gray dots. To define optimal values of $\Delta T^*$, various $\Delta T^*$ were tested at 1-K intervals for each AWS site. Figure 13 shows the

**Fig. 11.** Correlation coefficient $R$ of surface temperature observed by D66-AWS and 11-μm brightness temperature of NOAA channel 4 for the pixel that includes D66-AWS (solid curve, left axis) and relative frequency of cloud identifications (white bar, right axis) as a function of threshold value $\rho^*_1$ reflectance of NOAA channel 1.

**Fig. 12.** Scatter diagram of various $\Delta T^*$ over 11-μm brightness temperature of NOAA channel-4 $T_4$ for the pixel that includes D110-AWS (horizontal axis) and surface temperature observed by D110-AWS $T_{sfc}$ (vertical axis). Stars indicate identification as cloud free by both CD0 and CD2, squares indicate identification as cloudy by CD2 technique, black dots indicate CD0 identifying cloud when CD2 identified cloud free, and vice versa for gray dots. Here $R$ is the correlation coefficient of temperature in cases that both techniques identify as cloud free.

**Fig. 13.** (left axis) The correlation coefficient (solid line) of $T_4$ and $T_{sfc}$ for cases identified as cloud free by CD2 with threshold $\Delta T^*$ at D110-AWS. (right axis) The relative frequency of identifications where the white portion corresponds to identification as cloudy by both CD0 and CD2, the gray portion represents identification as cloudy by CD2 only and the black portion as identification by CD0 only. The total percentage of gray plus black corresponds to mismatches between the techniques. The minimum amount of misidentification is indicated by a thick arrow.
relative frequency of cloud identification of CD0 and CD2, with the white portion corresponding to identification by both CD0 and CD2, the gray portion to identification by CD2 alone, and the black portion to identification by CD0 alone. The correlation coefficient of $T_{slc}$ and $T_s$ was calculated for cases identified as cloud free by CD2 and is also shown in Fig. 13. Misidentifications (black plus gray) become minimum at around 10 K, at which point the correlation between $T_{slc}$ and $T_s$ is high. The same procedure was also applied to other AWS sites, giving similar but less marked results. Thus $\Delta T^*$ is set to 10 K.

4. Validation

a. Data acquisition rate

Using the method presented, the distribution of land surface temperature over the Tibetan Plateau was calculated using data from 1998. As mentioned earlier, the calculation of surface temperature is only possible in cloud-free areas. Figure 14 shows the percentage of the target that is cloud free. For each month, the ratio of cloud-free to total grids was measured and plotted as a horizontal map. In this study, the area above 4000 m MSL is defined as the Tibetan Plateau. Cloud-free ratio is generally found to be high during the autumn and winter months. In the summer months, especially July and August, the cloud-free ratio of the southern plateau drops to less than 50%. Diurnal variation averaged across the plateau is shown in Fig. 15. In each GMS observation, the ratio of cloud-free to total grids was counted and plotted as a function of yearday and UTC. From the sunset to next morning, from 1200 UTC to 0000 UTC, surface temperatures are retrieved for more than 80% of the total plateau except for the summer season. In the local afternoon, from 0600 to 1200 UTC,

<table>
<thead>
<tr>
<th>AWS Site</th>
<th>$T_{ms}$</th>
<th>$T_{ms}$ with CD1</th>
<th>$T_{ms}$ with CD2</th>
</tr>
</thead>
<tbody>
<tr>
<td>D66</td>
<td>0.5122 (23.54)</td>
<td>0.6501 (13.82)</td>
<td>0.9532 (4.43)</td>
</tr>
<tr>
<td>Tuotuohe</td>
<td>0.4731 (24.73)</td>
<td>0.4595 (20.77)</td>
<td>0.9456 (5.53)</td>
</tr>
<tr>
<td>D110</td>
<td>0.4694 (21.17)</td>
<td>0.5945 (12.56)</td>
<td>0.9028 (7.16)</td>
</tr>
<tr>
<td>MS3608</td>
<td>0.3699 (27.20)</td>
<td>0.5593 (14.92)</td>
<td>0.9054 (6.50)</td>
</tr>
<tr>
<td>Amdo</td>
<td>0.5162 (17.02)</td>
<td>0.5838 (14.67)</td>
<td>0.8304 (8.60)</td>
</tr>
<tr>
<td>Shiquanhe</td>
<td>0.5297 (23.13)</td>
<td>0.4865 (22.85)</td>
<td>0.8181 (11.99)</td>
</tr>
<tr>
<td>Gaize</td>
<td>0.5058 (24.57)</td>
<td>0.5118 (21.98)</td>
<td>0.8969 (9.52)</td>
</tr>
</tbody>
</table>
Fig. 15. Seasonal (horizontal axis) and diurnal (vertical axis) variation of cloud-free ratio over the Tibetan Plateau in 1998. The ratio of cloud-free grids to the total grids over the plateau is counted for each available GMS image.

Surface temperatures are calculated over less than one-half of the total plateau because of cloud cover, especially in March, April, July, and August, which is less than 10%.

b. Comparison with AWS data

The surface temperatures measured by satellite were compared with corresponding observations measured by IR thermometers at AWS sites. At all AWS sites, correlation coefficient and root-mean-square error (rmse) of satellite surface temperature measurements are computed and listed in Table 2. Very high correlations of over 0.9 and rmse with several degrees were achieved at four sites—D66, Tuotuohe, D110, and MS3608. This is to be expected, because data from these sites were used to establish the cloud removal threshold of CD2. At other sites, correlations are close to 0.8, with rmse of about 10°. To demonstrate the effectiveness of the new cloud removal technique, raw 11-μm brightness temperature and surface temperatures calculated using CD1 and CD2 cloud removal are plotted against observed values. A 240-K threshold value was used in CD1. Figure 16 shows data from the region of maximum correlation, Tuotuohe, whereas Fig. 17 shows data for the region of minimum correlation, Shiquanhe.
Using CD1, there are many data points at which calculated surface temperatures are exceedingly underestimated. This seems to have been caused by misidentification of cloudy areas as cloud free by CD1. These misidentified areas, however, are more effectively identified by CD2, resulting in higher correlation and lower rmse. Figure 18 shows a time series over 3–7 September of surface temperatures calculated using CD2 from $T_{IR1}$ (triangle), cloud identification from $T_{IR1}$ (square), surface temperatures measured by AWS (solid line), and solar radiation. The diurnal variation can be seen to be reproduced correctly by the CD2 calculations.

Calculated surface temperatures, however, exhibit considerable rmse with respect to observed value. This seems to be attributed to the difference in the spatial resolution of GMS images and AWS observations. The pixel size of GMS images is 51 km$^2$ at Tuotuohe and 74 km$^2$ at Shiquanhe, since GMS views the plateau from a geostationary orbit at 140°E. Thus, surface temperatures calculated from the satellite data are averages over large areas, whereas AWS observations are point measurements. Thus, AWS measurements are not always representative of the typical value over a GMS pixel, particularly in that subpixel-scaled cloud formations that affect the AWS would be missed by GMS.

Errors can also be attributed to precipitable water estimation errors, as shown in Fig. 8. Since fluctuations in atmospheric attenuation are assumed to depend only on precipitable water, the value of precipitable water has a great impact on surface temperature calculations. This is particularly true in the western plateau, where satellite zenith angles are so large that even precipitable water estimates will contain considerably large errors.

c. Surface temperature over the Tibetan Plateau

Hourly surface temperatures over the Tibetan Plateau in 1998 were calculated. As an example, diurnal variation on 25 April is shown in Fig. 19. After sunrise, surface temperatures begin to rise in the eastern plateau.
In some areas, surface temperatures rise by more than 30 K from early morning to noon. White blanks in each figure correspond to cloudy areas over the plateau. In the morning, surface temperatures can be calculated for most of the plateau. By afternoon, however, surface temperatures for only about one-half of the plateau can be calculated because of cloud cover, the development of which is probably caused by rapid heating of the ground.

To compare variations in surface temperature over the different parts of the Tibetan Plateau, monthly mean daily minimum and maximum surface temperatures are calculated from available data for each grid. Spatial distribution of monthly mean daily maximum surface temperature is depicted in Fig. 20. From May to July, daily maximum surface temperature in the western part of the plateau is higher than that in the east. During this period, daily minimum surface temperature is lower in the western part of the plateau than in the east, but the west-east contrast is less obvious. Consequently, diurnal range of surface temperature in the western part of the plateau becomes greater than that of the eastern as shown in Fig. 21. This seems to be due to a difference in surface conditions between the western and eastern parts of the plateau. The western part of the Tibetan Plateau is relatively dry, so that radiative cooling in the nighttime is expected to be much stronger than in the east. In addition, the ground surface of the western plateau is easy to heat and cool. Thus, the diurnal range 

![Fig. 19. Horizontal distribution of available retrieved surface temperatures from 0000 to 1100 UTC, 25 Apr 1998. Blank white region over the Tibetan Plateau is a cloudy area.](image1)

![Fig. 20. Horizontal distribution of monthly mean daily maximum surface temperature in 1998.](image2)
of surface temperature is about 10 K greater in the western plateau than in the east. In winter and spring, the diurnal range of surface temperature is more than 30 K in both the east and west of the plateau, reaching a maximum in March and April. In 1998, the onset of the monsoons took place in early June (Matsumoto et al. 1999). After the onset of monsoons, surface soil moisture increases gradually because of increased precipitation. Reflecting this, the diurnal range of surface temperature drops, with a minimum value in the midmonsoon period.

5. Conclusions and remarks

The algorithm for estimating surface temperatures from NOAA data was modified for application to GMS. Specifically, the determination of split-window coefficients in Eq. (6) was performed by MODTRAN radiative transfer code using the appropriate 11- and 12-μm channel filter functions of GMS VISSR. GMS VISSR measurement of radiation from the ground surface over the Tibetan Plateau is simulated to predict spectral radiances and transmittances under various atmospheric profiles. These coefficients were defined to be functions of both precipitable water and satellite zenith angle at each grid point.

In addition, the fluctuation of atmospheric attenuation was assumed to depend only on precipitable water. Precipitable water was estimated from 6.7-μm brightness temperature using simple linear regression by comparison with sonde observations. Furthermore, a new cloud detection technique was proposed in which the threshold value is a function of both local time of day and the day of year. This technique identifies cloudy areas much more effectively than the fixed threshold method, resulting in a much higher correlation factor.

Results of the proposed algorithm were compared with AWS observation data, resulting in high correlation factors, indicating that the estimated surface temperatures are a good match to observed values, even though there is a considerable rmse of nearly 10 K. This can be partially explained by the difference in the areas measured by the satellite and ground measurement techniques, since a single GMS VISSR pixel represents an area of several tens of square kilometers, whereas an AWS IR thermometer covers an area of less than 1 m².

The greatest advantage of GMS is that observations of the earth are made from a stationary position above the equator from which meteorological phenomena can be monitored continuously, making the analysis of intradurnal variations over large areas possible. Polar-orbiting satellites such as NOAA enable observation of the entire globe by moving rapidly relative to the surface of the earth. While this means that polar areas can be observed frequently, long time frames are needed to make repeated observations of mid- and low-latitude
locations when compared with geostationary satellites such as GMS.

As the successor to GMS-5, the Multifunctional Transport Satellite (MTSAT) will provide precise infrared data to a precision of 10 bits, as compared with the 8 bits of GMS-5. Furthermore, the Chinese meteorological satellite FY2c will operate in geostationary orbit at 105°E. FY2c will also contain a split-window infrared channel with the same precision as MTSAT, but will view the Tibetan Plateau from a smaller satellite zenith angle than GMS-5. Images observed by these satellites will allow the accuracy of surface temperature measurements to be improved. In the future, the estimation of surface energy fluxes across the plateau from nearly continuous, long time-scale monitoring is expected to be possible.

REFERENCES


