Estimation of Land Surface Heat Fluxes over the Tibetan Plateau Using GMS Data

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Abstract

A Surface Energy Balance System (SEBS) originally developed for the NOAA Advanced Very High Resolution Radiometer was applied to Geostationary Meteorological Satellite (GMS)-5 Visible/Infrared Spin-Scan Radiometer data that were supplemented with other meteorological data. GMS-5, which is a geostationary satellite, recorded continuous hourly information. Surface temperatures obtained from the GMS-5 data were entered into SEBS to estimate the hourly regional distribution of the surface heat fluxes over the Tibetan Plateau. The estimated fluxes are verified by using corresponding field observations. The diurnal cycle of estimated fluxes agreed well with the field measurements. For example, the diurnal range of the estimated sensible heat flux decreases from June to August. This reflects the change of dry to wet surface characteristics resulting from frequent precipitation during the summer monsoon. Over the Tibetan Plateau, the diurnal range of the surface temperature is as large as the annual range, so that the resultant sensible heat flux has a large diurnal variation. Thus, the hourly estimation based on the GMS data may contribute to a better understanding of the land surface–atmosphere interaction in this critical area.

1. Introduction

The Tibetan Plateau is assumed to play a crucial role in the progress of the Asian summer monsoon through thermal effects on the surrounding midtroposphere (e.g., Yanai et al. 1992; Yanai and Li 1994). To understand the role of the Tibetan Plateau as an elevated heat source in this region, intensive meteorological observations were conducted during the Global Energy and Water Cycle Experiment (GEWEX) Asian Monsoon Experiment (GAME) Tibet project. During GAME/Tibet, the surface sensible and latent heat fluxes, as well as the relevant surface parameters, were measured at different sites. Using the data obtained in these experiments, the land surface interaction was observed experimentally at these sites. However, these results only provided the fluxes for the immediate area surrounding the measurement sites. This local knowledge of the fluxes needed to be combined with an overall understanding of the plateau. Satellite remote sensing permits the derivation of a regional distribution of the surface energy fluxes in addition to the data provided from local field experiment stations. Ma (2003) estimated the distribution of land surface variables over the GAME/Tibet enhanced observation area by combining the National Oceanic and Atmospheric Administration (NOAA) Advanced Very High Resolution Radiometer (AVHRR) data with field observations. According to Tanaka et al. (2001), however, strong diurnal variations are one of the outstanding features of the plateau surface, 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. The use of a geostationary satellite for surface energy flux retrievals has already been extensively studied (e.g., Diak et al. 2004). In this study, the Surface Energy Balance System (SEBS) retrieval algorithm that is used

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for the NOAA AVHRR data (Su 2002) is applied to the Geostationary Meteorological Satellite (GMS)-5 data.

2. Surface Energy Balance System

The net radiation flux \( R_n \) is estimated as

\[
R_n = (1 - \alpha)R_S^↓ + R_L^↓ - \varepsilon_{sfc}\sigma T_{sfc}^4,
\]

(1)

where \( R_S^↓ \) is the downward shortwave radiation as a function of the solar constant, the atmospheric transmittance, and local time, \( \alpha \) is the surface albedo, \( \varepsilon_{sfc} \) is the emissivity of the surface, \( R_L^↓ \) is the downward longwave radiation as a function of atmospheric temperature, \( \sigma \) is the Stefan–Boltzmann constant, and \( T_{sfc} \) is the surface temperature. In this equation, the diurnal range of \( \alpha \) and \( \varepsilon_{sfc} \) is expected to be vanishingly small relative to \( R_S^↓ \), \( R_L^↓ \), and \( T_{sfc} \). It is sufficient to allow the net radiation retrieval to be given by \( \alpha \) (Valiente et al. 1995) and \( \varepsilon_{sfc} \) (Sobrino and Raissouni 2000) using the monthly mean value given in the NOAA data. Here \( R_S^↓ \) is computed from the solar constant \( J_{sc} \) multiplied by solar zenith angle \( \theta \) and the atmospheric transmittance \( \tau \) of the shortwave spectrum (e.g., Iqbal 1983), that is,

\[
R_S^↓ = e_0 J_{sc} \tau \cos \theta,
\]

(2)

where \( e_0 \) is the eccentricity-orbiting correction factor. Many authors assume that the fluctuations in atmospheric attenuation depend only on precipitable water, which is determined from the vertical integration of water vapor. If this assumption is accepted, then asymptotic formulas for atmospheric transmittance can be derived as a function of the precipitable water. Atmospheric transmittance is computed using the Moderate-Resolution Atmospheric Radiance and Transmittance Model (MODTRAN) radiative transfer code (Berk et al. 1989) for different profiles. For profiles of the atmospheric gaseous components, the typical midlatitude profiles were selected from the MODTRAN options. Results are plotted as a function of the precipitable water in Fig. 1. Precipitable water is estimated directly from the brightness temperature of the water vapor channel. Because the weighting function of the 6.7-μm sensor on GMS-5 has a maximum near 400 hPa and does not reflect humidity at lower levels, the 6.7-μm brightness temperature is usually a poor indicator of precipitable water. Over the Tibetan Plateau, however, the ground surface is in the 500–600-hPa level, so that 6.7-μm brightness temperature is expected to be related to the precipitable water information there (Oku and Ishikawa 2004). The \( R_L^↓ \) is computed as

\[
R_L^↓ = \varepsilon_{air}\sigma T_{air}^4,
\]

(3)

where \( \varepsilon_{air} \) is obtained from an empirical formula given in Swinbank (1963).

The soil heat flux \( G_0 \) is estimated using net radiation as follows:

\[
G_0 = R_n[\Gamma_c + (1 - f_c)(\Gamma_s - \Gamma_c)],
\]

(4)

where \( \Gamma_s \) is the ratio between \( G_s \) and \( R_s \) for bare soil and \( \Gamma_c \) is the ratio between \( G_c \) and \( R_c \) for a surface fully covered with vegetation. The value of \( \Gamma_s \) is taken as 0.315 (Kustas and Daughtry 1990), and \( \Gamma_c \) has a value of 0.05 (Monteith 1973). The fractional vegetation cover \( f_c \) is determined using the normalized difference vegetation index (NDVI; Sobrino and Raissouni 2000).

To derive the sensible heat flux \( H \), the similarity theory is used. In a complex landscape there is a height, called the blending height, where the impact of the underlying surface diminishes (e.g., Mason 1988; Grant 1991). Although there are various atmospheric profiles over a single pixel of GMS-5 measurement, the blending-height concept allows one to assume a representative profile over the complex landscape in the Tibetan Plateau. At the blending height \( z \), wind speed \( u \) and air temperature \( T_{air} \) satisfy the general conditions given by Monin–Obukhov similarity theory (Monin and Obukhov 1954). They are described as

![Fig. 1. Atmospheric transmittance for the shortwave spectrum as a function of the precipitable water (including the regression curve).](Image)
so that the actual field of view of VISSR is about 5 km for instantaneous field of view of VISSR is about 5 km for the plateau is between 7 and 10 km. The diurnal flux retrieval. According to Oku and Ishikawa (2004), hourly surface temperatures are estimated using satellite data. Surface temperature is one of the most important parameters in the diurnal flux retrieval. According to Oku and Ishikawa (2004), hourly surface temperatures are estimated using an infrared split-window technique based on the GMS-5 VISSR data. The detailed model evaluation has been done and presented in Oku and Ishikawa (2004). The results of comparing estimated surface temperature from GMS-5 data using this algorithm with in situ surface measurements by an automatic weather station

\[
u = \frac{u_w}{k} \left[ \ln \left( \frac{z - d_0}{z_{om}} \right) - \Psi_m \left( \frac{z - d_0}{L} \right) \right] - \Psi_h \left( \frac{z_{om}}{L} \right) \]

(5)

\[
T_{sfc} - T_{air} = \frac{H}{k \mu_p C_p} \left[ \ln \left( \frac{z - d_0}{z_{om}} \right) - \Psi_h \left( \frac{z - d_0}{L} \right) \right] - \Psi_h \left( \frac{z_{om}}{L} \right) \]

(6)

where \( k \) is von Kármán’s constant, \( \rho \) is the air density, \( C_p \) is the specific heat constant, \( u_w \) is the friction velocity, \( d_0 \) is the zero-plane displacement height (Stanhill 1969), \( z_{om} \) is the roughness height for momentum transfer (Gupta et al. 2002), \( z_{oh} \) is the roughness height for heat transfer (Su 2002), \( \Psi_m \) is the stability correction function for momentum transfer, and \( \Psi_h \) is the stability correction function for sensible heat transfer (Brutsaert 1982). The Monin–Obukhov stability length \( L \) is defined as

\[
L = - \frac{\rho C_p T_{air} u_w^2}{g h}.
\]

where \( g \) is the acceleration resulting from gravity. Derivation of the sensible heat flux \( H \) using Eqs. (5)–(7) requires only the wind speed \( u \) and temperature \( T_{air} \) at the blending height \( z \), as well as the surface temperature \( T_{sfc} \).

The latent heat flux \( \lambda E \) is the residual resulting from an application of the energy budget theorem to the land surface:

\[
\lambda E = R_n - G_0 - H.
\]

Cloud removal has an important part in these fluxes’ retrieval process. To identify convective cloud activity, many researchers use satellite infrared measurements with a fixed threshold technique. In this study, however, it is necessary to remove not only convective clouds but also all kinds of clouds. For this purpose, a variable threshold technique proposed in Oku and Ishikawa (2004) is used:

\[
T_{IR1} < T_{IR1}^{*} (DOY, UTC) \text{ for cloudy}
\]

\[
T_{IR1} \geq T_{IR1}^{*} (DOY, UTC) \text{ for cloud free}
\]

where \( T_{IR1} \) is GMS-5 brightness temperatures at 11 \( \mu \)m. The threshold \( T_{IR1}^{*} \) varies both seasonally [day of year (DOY)] and diurnally (UTC), and the values are determined on the basis of surface observations. As a result of adopting this technique, it becomes possible to remove relatively warmer clouds in summer and to detect colder ground surfaces in winter nighttime.

3. Data

GMS-5 was launched on 18 March 1995 and provides coverage of the Asia–Pacific region from Hawaii to India. Continuous hourly information was recorded for about 8 yr to 22 May 2003 when operation was terminated. GMS-5 had an onboard Visible/Infrared Spin-Scan Radiometer (VISSR), which senses three infrared channels—two split-window channels (IR1, 11 \( \mu \)m; IR2, 12 \( \mu \)m) and a water vapor channel (6.7 \( \mu \)m). The instantaneous field of view of VISSR is about 5 km for these channels at the subsatellite point. Because 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.

Figure 2 shows the process flowchart of the method used in this study. Surface parameters, including surface temperature \( T_{sfc} \), albedo \( \alpha \), and emissivity \( \epsilon \), as well as atmospheric parameters, including wind speed \( u \) and air temperature \( T_{air} \), at the blending height \( z \) are required to obtain surface fluxes using SEBS. The surface parameters are calculated using satellite data. Surface temperature is one of the most important parameters in the diurnal flux retrieval. According to Oku and Ishikawa (2004), hourly surface temperatures are estimated using an infrared split-window technique based on the GMS-5 VISSR data. The detailed model evaluation has been done and presented in Oku and Ishikawa (2004). The results of comparing estimated surface temperature from GMS-5 data using this algorithm with in situ surface measurements by an automatic weather station.
(AWS) shows a high correlation coefficient (nearly 0.9), with a root-mean-square error (RMSE) of about 10 K. Other surface parameters, such as albedo and emissivity, do not appreciably vary during the interdiurnal time period. Instead, these values vary seasonally. These values are estimated from the visible reflectance measured in the NOAA-14 AVHRR channels 1 and 2, based on a monthly mean. All of the available GMS and NOAA images from May 1998 to April 1999 were used to estimate the flux distribution over the Tibetan Plateau. The study area was defined as the area higher than 4000 m MSL. The satellite data in grid format with a longitude and latitude resolution of 0.1° were interpolated from the original image data that were archived in a line-pixel format. However, neither the GMS nor the NOAA data can be used to determine the wind speed and air temperature at the blending height. Instead, they are determined using radiosonde (“sonde”) measurements or are obtained from objective analysis data. In this study, the first step for the calculation of the surface fluxes used the sonde observation data at Amdo during the GAME/Tibet intensive observation period (IOP) from April to September 1998 to derive the required values. Then, using the 40 yr of data from the European Centre for Medium-Range Weather Forecasts (ERA-40; Simmons et al. 2000), we determined their distribution over the plateau so that the surface fluxes over the whole Tibetan Plateau could be estimated.

Surface fluxes measured at Amdo were used as the basis for comparing the estimated fluxes determined from the presented algorithm. The location of Amdo is shown in Fig. 3. The net radiation flux was obtained using data from the four-component radiation system. The soil heat flux is computed using the soil temperatures, soil moisture profiles, and the thermal conductivity equation given below:

\[
G_0 = T_c \left( \frac{\partial T_{sfc}}{\partial t} \delta_{sfc} + \frac{\partial T_{5cm}}{\partial t} \delta_{5cm} + \frac{\partial T_{10cm}}{\partial t} \delta_{10cm} \right) + G_{10cm}.
\]

According to Tanaka et al. (2003), the \( \delta \) terms are the effective thicknesses of each measured temperature: \( T_{sfc}, T_{5cm}, \text{ and } T_{10cm} \). These terms are determined empirically. Here \( G_{10cm} \) is the observed soil heat flux at a depth of 10 cm after correction for soil moisture; \( T_c \) is the average volumetric heat capacity. The sensible heat flux is calculated from turbulence data sampled in 10 Hz by a sonic anemometer using the eddy correlation method every 30 min. A problem was identified in the latent heat flux data calculated using the eddy correlation method that could be attributed to the performance of the infrared hygrometer. As the summer monsoon progressed, precipitation became frequent and the performance of the infrared hygrometer degraded as a result of weakening of the incident infrared radiation. This can be expected to result in an underestimation of the latent heat flux. Thus, the latent heat flux is considered to be the residual energy budget.
for land as described in Eq. (8). For an annual cycle verification, the sensible heat flux was computed using the boundary layer profile data observed at Amdo and the bulk transfer method (Tanaka et al. 2004), because the turbulence measurements were only conducted during the IOP phase of GAME/Tibet. All data, except for the turbulence measurements, were sampled every 1 s and recorded in a data acquisition logger. Averages were calculated for every 30-min time period.

4. Validation

Before an estimation of the flux over the plateau could be attempted, the sonde observational data at Amdo were entered into the algorithm to determine the diurnal flux cycle. As compared with sonde measurements, ERA-40 may be a kind of quasi-observational data. However, there is not a sufficient number of sonde observation stations across the Tibetan Plateau, and thus the ERA-40 data were used to determine the flux distributions over the plateau. Because of the sonde measurement-period limitation, fluxes retrieved with sonde data are validated with surface observation data obtained from April to September 1998, whereas those retrieved with ERA-40 reanalysis data are validated from May 1998 to April 1999.

a. Land surface energy flux retrieval with sonde data

As shown in Fig. 4 and Table 1, the surface fluxes obtained using the presented algorithm were compared with the corresponding surface fluxes observed at
Amdo every hour. A high correlation coefficient suggests that the diurnal cycles for the computed surface fluxes reproduce well the observed surface flux diurnal cycle. However, the computed net radiation considerably overestimates the actual net radiation. This can possibly be attributed to a difference in both the temporal scale and the spatial scale of measurement used by the satellite sensor and the ground-based instrument.

The temporal scale difference can be attributed to measurement differences. Although the surface observational net radiation was determined using 30-min-averaged data that were sampled every second by the radiation sensors, the net radiation using SEBS was calculated to be the radiation under a completely clear sky at the instantaneous time of GMS observation. For example, as shown in Fig. 5 at 0830 UTC 11 July 1998 (1430 LST at Amdo), the observed net radiation was 378.7 W m$^{-2}$ while the calculated net radiation was 499.36 W m$^{-2}$ using SEBS retrieval. As seen in the right panel of Fig. 5, the downward shortwave radiation measured at 0800 UTC (1400 LST) was smaller than that expected with a completely clear sky. In fact, Fig. 6 shows the cloud images around the Amdo station, which suggest that it was cloudy at Amdo at this time. Thus, during the 30-min time period used to determine the average value it could have been partly cloudy, and the value of the observed surface radiation was smaller than the calculated value using SEBS. On the other hand, no difference between the observed and calculated net radiation values is expected if the sky was clear for the duration of the 30-min period.

The spatial scale difference can be attributed to a problem with the resolution of the images. A single GMS infrared pixel represents the area of several tens of square kilometers over the Tibetan Plateau, whereas the ground-based instrumentation observes an area of less than a square meter. In particular, sub-pixel-scale cloud formations that affect ground-based measurement could easily be missed by the GMS. In Fig. 6, it is difficult for the GMS to detect a cloud band found at Amdo in the NOAA image. In the NOAA image, the brightness temperature of the pixel that includes Amdo is much lower than expected. This pixel seems to represent radiation from the top of a cloud rather than that from the surface. The current algorithm does not permit the retrieval of surface fluxes from pixels with cloud. A comparison of the GMS image with the NOAA image shows that the pixel including Amdo in the GMS consists of both a clear area and a cloudy area.

### Table 1. The statistics for the measured vs calculated surface energy fluxes at Amdo.

<table>
<thead>
<tr>
<th></th>
<th>$R_n$</th>
<th>$G_0$</th>
<th>$H$</th>
<th>$\lambda E$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R$</td>
<td>0.9575</td>
<td>0.9388</td>
<td>0.7899</td>
<td>0.6638</td>
</tr>
<tr>
<td>RMSE</td>
<td>82.506</td>
<td>44.161</td>
<td>79.585</td>
<td>112.538</td>
</tr>
<tr>
<td>$N$</td>
<td>784</td>
<td>411</td>
<td>721</td>
<td>721</td>
</tr>
</tbody>
</table>

Fig. 5. The time series of (left) SEBS-calculated and (right) the measured surface energy fluxes for $R_n$ (open circles), $G_0$ (open diamonds), $H$ (filled circles), $\lambda E$ (filled upside-down triangles), and the half-value of the measured downward shortwave radiation (shaded area with gray line; right panel) at Amdo on 11 Jul 1998. The horizontal axis is the local time at 90°E, and the vertical axis is the energy flux density (W m$^{-2}$).
The brightness temperature of the GMS pixel that includes Amdo is higher than that of the NOAA image when the surface radiation is assumed based on clear-sky conditions. Thus, this algorithm miscalculates the surface flux at this pixel, which leads to an overestimation in the net radiation using SEBS.

As shown in Fig. 5, the diurnal soil heat flux has a phase difference between the observed values and the values calculated using SEBS. The daily maximum in observed soil heat flux occurs earlier than in the calculated soil heat fluxes. This situation is caused by a difference in the estimation method used for the field measurement and SEBS. According to Eq. (4), the soil heat flux calculated using SEBS is determined from the net radiation and the ratio between the soil heat flux and the net radiation \[ \frac{\Gamma_s}{H} \], which have constant daily values. Because diurnal variation of soil heat flux depends on that of the net radiation, the soil heat flux takes its daily maximum at local noon. On the other hand, the observed soil heat flux is obtained using Eq. (10). In this equation, the diurnal variation of the soil heat flux depends on the time derivative with respect to soil temperature. As shown in the right panel of Fig. 7, \( \delta T/\delta t \) takes its daily maximum value before local noon. Hence, the observed soil heat fluxes reach their daily maximum earlier than the calculated values. This problem was previously mentioned by Kustas et al. (2000). Even after allowing for this phase difference, the estimated soil heat fluxes still exhibit considerable RMSE with respect to the observed values. This can partially be explained by noting that the ratio between the soil heat flux and net radiation in Eq. (4) does not consider the effect of soil moisture conditions. The ratio in Eq. (4) can only vary between 0.05 and 0.315 and

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**Fig. 6.** (left) GMS-5 VISSR IR1 brightness temperature image at 0823 UTC and (right) the NOAA-14 AVHRR channel-4 brightness temperature image at 0828 UTC 11 Jul 1998. (top) The white square corresponds to the area from which the image in the lower panels is taken. (bottom) The white arrow indicates the pixel that includes the Amdo station, and the value shows the brightness temperature in this pixel.
depends on the fractional vegetation cover $f_v$. However, this ratio should also be affected by soil moisture. If soil moisture increases, this ratio is also expected to increase, for example, to 0.5 (Santanello and Friedl 2003).

Although negative sensible heat fluxes are often observed during the night, especially around dawn, some of these are not reproduced in the SEBS method (Fig. 4). In general, a stable layer is well developed during the night because of strong radiative cooling in the Tibetan Plateau. Because the Monin–Obukhov similarity theory applies only to the boundary layer under near-neutral conditions, the boundary layer height cannot be
defined under very stable conditions. Thus, some negative sensible heat fluxes are estimated by SEBS as positive values.

b. Land surface energy flux retrieval with reanalysis data

In the previous section, sonde observational data were used as the input wind speed and air temperature for SEBS. To compute surface fluxes over the Tibetan Plateau, the spatial distribution of wind speed and air temperature is required. However, it cannot be obtained because of the sparse distribution of sonde data over the plateau. In this section, the ERA-40 data are used, so that they could act as an alternative to the sonde data in providing the necessary wind speed and air temperature for SEBS. To assess the applicability of ERA-40, the ERA-40 data for the nearest grid point to Amdo (the filled circle in Fig. 3) were compared with the sonde observational data as shown in Fig. 8. The correlation coefficient and RMSE between ERA-40 and the sonde data were 0.857 and 1.29 K for air temperature and 0.458 and 3.44 m s$^{-1}$ for wind speed. These RMSEs suggest that an error of 5 W m$^{-2}$ in net radiation, an error of 2 W m$^{-2}$ in soil heat flux, and an error of 20 W m$^{-2}$ in sensible heat flux can be expected for the computed SEBS values.

The ERA-40 data were input into SEBS to calculate the surface flux across the Tibetan Plateau. The surface fluxes were computed from the ERA-40 data using the presented algorithm and were compared with the corresponding observational surface values at Amdo (Fig. 9 and Table 2). Thus, the fluxes calculated using the ERA-40 data closely agree with the values obtained using sonde observations. Figure 10 displays the sea-
sonal variation of the calculated fluxes in comparison with measured fluxes. In general, more solar energy is absorbed in the summer than in the winter. Thus, the annual daily maximum net radiation results in a sine wave pattern. In rough terms, the net radiation calculated using SEBS appears to reproduce the same pattern. In 1998, the onset of the monsoons took place in early June (Matsumoto et al. 1999). After the onset of the monsoons, surface soil moisture increased gradually because of increased precipitation. The sensible heat flux accordingly decreases and the latent heat flux exceeds the sensible heat flux. This dramatic change between sensible and latent heat flux can be seen in both the SEBS-calculated values and the surface measurements. Overall, the seasonal variation in the calculated fluxes agrees with the observational curve.

5. Conclusions and remarks

In this study, the algorithm originally developed for determining the land surface energy fluxes from

| Table 2. Same as Table 1, but ERA-40 data are used to calculate the fluxes. |
|-----------------|-----------------|-----------------|-----------------|-----------------|
| $R_n$           | $G_0$           | $H$             | $\lambda E$     |
| $R$             | 0.9555          | 0.9086          | 0.8687          | 0.7134          |
| RMSE            | 86.026          | 44.588          | 47.691          | 134.439         |
| $N$             | 1235            | 679             | 935             | 935             |

Fig. 10. Thick line is the time series of daily averaged surface energy fluxes observed by surface measurements and open circles are that estimated using SEBS (from top to bottom: $R_n$, $G_0$, $H$, and $\lambda E$) at Amo from May 1998 to April 1999 (W m$^{-2}$).
Fig. 11. The horizontal distribution of the available calculated surface energy fluxes (W m$^{-2}$; from top to bottom: $R_n$, $G_0$, $H$, and $\lambda E$) from 0000 (0600 LST at 90°E) to 0600 UTC 25 Apr 1998. The white region over the Tibetan Plateau is a cloudy area.
NOAA AVHRR data is used to derive a similar algorithm for GMS-5 VISSR data. Because the land surface temperatures, the shortwave radiation, and the longwave radiation predominantly vary diurnally, these values are primarily determined from the GMS data. Because surface albedo, emissivity, and NDVI do not vary appreciably during the day but instead vary seasonally, these parameters can be obtained from NOAA data.

First, the diurnal cycle in surface fluxes was computed using the GMS data based on sounding observations. The results of this algorithm were compared with the observed surface values. High correlation factors were observed, suggesting that the calculated surface fluxes were in close agreement with the observed values, despite the high RMSE. The high RMSE can partially be explained by differences in temporal and spatial scales used by the satellite and ground measurement instrumentation. A single GMS-5 VISSR pixel represents an area of several tens of square kilometers, whereas a surface measurement covers an area of less than 1 m². In addition, the method used to compute the soil heat flux is different for the surface measurements and the satellite measurements. The difference in computation produces a phase difference between the calculated and observed soil heat fluxes. Microwave radiation emitted from the earth’s surface can be used to detect soil moisture remotely. Satellite-based passive microwave sensors offer an effective way to observe soil moisture data over vast areas. There are currently several satellite systems that detect soil moisture (e.g., Wen et al. 2003). The ratio between the soil heat flux and the net radiation should be improved by reflecting soil moisture data obtained from satellite measurements, in the next stage. On the other hand, the surface energy balance is assumed to obtain latent heat flux.

Second, the surface fluxes were computed using the GMS data with the ERA-40 data. Because there are few sounding observation stations across the Tibetan Plateau, the ERA-40 data are required to estimate the surface flux distribution over the plateau. The surface fluxes calculated using SEBS with the aid of the ERA-40 data can be calculated to the same accuracy as the sounding data. Furthermore, the seasonal variation in the calculated fluxes agrees with the experimentally observed variation. Figure 11 shows the spatial distributions of the calculated fluxes. The greatest advantage in using GMS is that the earth is observed from a stationary position, which allows the meteorological phenomenon to be continuously monitored.

This study is the first attempt to apply a surface energy flux computation algorithm to geostationary satellite data for the Tibetan Plateau. Utilizing the surface flux data calculated using this technique, a better quantitative understanding of the interactions between the surface and the atmosphere over the Tibetan Plateau can be obtained. For example, the heat source $Q_1$ and the moisture sink $Q_2$ (Yanai et al. 1992) can be obtained from the ERA-40 data as an index of the heat budget over the troposphere and across the plateau. In comparing $Q_1$ and $Q_2$ with the surface fluxes, the atmospheric heat budget from the ground surface to the atmosphere can be quantified for a diurnal-seasonal time period. Thus, the surface flux data can contribute useful information not only for meteorology or climatology, but also for environmental studies, civil engineering, and agricultural needs. The data can also be used to monitor the surface water and energy budgets.

As a successor to the GMS-5, the Multifunctional Transport Satellite (MTSAT) provides infrared data to a precision of 10 bits, as compared with only an 8-bit precision for GMS-5. Furthermore, the Chinese meteorological satellite FY2c operates at a geostationary orbit at an altitude of $105^\circ$E. FY2c also contains a split-window infrared channel with the same precision as MTSAT, but which views the Tibetan Plateau from a smaller satellite zenith angle than does GMS-5. In the future, the images obtained from these satellites will allow the accuracy of calculated surface fluxes to be improved.

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