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Estimation of daily actual evapotranspiration from remotely sensed data under complex terrain over the upper Chao river basin in North China

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Daily actual evapotranspiration over the upper Chao river basin in North China on 23 June 2005 was estimated based on the Surface Energy Balance Algorithm for Land (SEBAL), in which the parameterization schemes for calculating the instantaneous solar radiation and daily integrated radiation were improved by accounting for the variations in slope and azimuth of land surface and terrain shadow in mountainous areas. The evapotranspiration (ET) estimated from satellite data in this study for the whole watershed ranges from 0 mm to 7.3 mm day\(^{-1}\) with a mean of 3.4 mm day\(^{-1}\), which was validated by Penman–Monteith approaches for water body and paddy land. The comparison of ET estimates for a wide range of land cover types reflected distinct mechanisms of energy partition and water removal of various land cover types, showing differences in the spatial distribution pattern of ET, which could be not only the reflection but also the driving force of advection and local circulation that may violate the surface energy balance equation in the vertical direction. The spatial variation in daily solar radiation and ET estimates under the complex terrain of forest land were elaborated and evaluated by exploring the relationship between ET estimates and elevations for woodland and grassland. In addition, the utility and limitations of SEBAL’s applicability to watersheds with various land cover types and complex terrain were analysed.

1. Introduction

Apart from rainfall and runoff, evapotranspiration (ET) including water evaporation from soil surfaces and vegetation transpiration represents a fundamental process of hydrological cycle and a key element of water resources management, particularly in semi-arid and arid regions, because it controls the partition of energy and water fluxes at the earth’s surface. Numerous measurements available have demonstrated that the temporal and spatial pattern of ET depend on a large variety of influencing factors, specifically vegetation and soil types, topography, and the meteorological conditions. The primary methods used conventionally to measure ET are subject to field or landscape scales (Bowen ratio, eddy correlation system, soil water balance), but do not allow the estimation of fluxes when dealing with larger spatial scales because of the heterogeneity inherent in land surfaces and the dynamic nature of water–heat transport processes. Nowadays, remote sensing data

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with increasing imagery resolution is probably the only technique for providing variables at various temporal and spatial scales required to estimate ET with different models being developed. Surface Energy Balance Algorithm for Land (SEBAL) (Bastiaanssen et al. 1998a, 1998b) uses spectral radiances recorded by satellite-based sensors plus traditional meteorological data to solve the energy balance at the land surface. Many successful applications of this model have been demonstrated in recent years by Bastiaanssen et al. (1998a, 1998b), Chemin et al. (2004) and Kimura et al. (2007).

Most studies with respect to ET estimates using remote sensing data specifically focus on the homogeneous land cover type under simple terrain conditions, such as grassland (Bastiaanssen 2000; Kimura et al. 2007), woodland (Bastiaanssen 2000; Goodrich et al. 2000; Wu et al. 2006), irrigation area (Bastiaanssen 2000; Jaime and Christopher 2005), etc. However, for heterogeneous land cover types under complex terrains at the regional scale, it becomes considerably difficult because:

1. The parameterization schemes for computing the instantaneous and daily net radiation available for sloped lands are complicated due to their distinction in various slopes and azimuths. Available radiation is a critical input variable in the energy balance equation and the most sensitive variable in ET estimates (Zhang et al. 2005). Therefore, the accuracy of retrieved instantaneous and daily net radiation determines the accuracy of estimates of ET to some extent; especially for the middle and high latitude regions the difference in available radiation for various slopes and azimuths of terrain is always pronounced (Fu 1983). In many applications, the daily net radiation is obtained from meteorological stations located in flat areas or calculated based on simplification of the terrain via the assumption of uniform extensive slope and azimuth, which may lead to some gross errors in cases of rough terrain (Allen et al. 2006). For instance, if the solar declination angle is very small, specifically in winter, the north facing surfaces in middle or high latitude regions receive relatively smaller net radiation compared to the south facing surfaces. If the terrain is very complex, some areas may not receive any direct radiation despite being south facing because of the high obstacles surrounding them. Although these situations are not always dominant, the net radiation available for rough terrain is distinct in general (Fu 1983; Allen et al. 2006). Thus, it is essential to accurately retrieve the net instantaneous and daily net radiation for applications involving estimation of the ET and other components of surface energy fluxes.

2. The advection could be formed under complex terrain with heterogeneous land cover because a watershed involves mixed land cover and water vapour losses may exhibit large spatial variation (Biftu and Gan 2000), which may violate the energy balance principle in the vertical direction. For instance, the water body, woodland, grassland, wetland, and irrigated area surrounded by a very hot sandy land or in other dry environment can induce an advection due to the oasis effect (Spronken-Smith et al. 2000). In a dry environment, the very hot surface and surface layer may be subject to the confounding influences from soil moisture and water vapour advection, namely inverse the humidity phenomenon. Furthermore, the processes and mechanisms of energy and mass transfers will become more complicated under the complex terrain due to the substantial differences in radiation availability caused by various slopes and azimuths of surfaces, and the non-negligible advection
Effect. It is uncommon to estimate the ET account for advection and local circulation in practice. However, as a first step, research focused on the net radiation retrieval considering the characteristics of the complex terrain and the exploration of ET spatial variation in diverse types of land cover are very important for further in-depth research into the boundaries and transition zones between different land covers where advection and local circulation may be active.

The emphasis of our study will be placed on issues of radiation availability for slope land and spatial distribution of ET, additionally attempting to find the effect of advection upon ET derived by SEBAL. The primary intention of this study is to compute the instantaneous and daily net radiation for various slopes and azimuth terrains in order to make each component of the energy balance equation more accurate, and consequently to improve the accuracy of ET estimates under complex terrains at the watershed scale. The specific objectives are (1) to examine the relationship between ET and land cover types; (2) to examine the spatial distribution of ET under complex terrain over semi-humid to semi-arid watersheds, and (3) to examine the uncertainties of SEBAL in ET estimates.

2. Study area

The study was conducted in the watershed between 41°02′ to 41°37′ N and 116°08′ to 116°45′ E, located around Dage hydrological station in the upper Chao River basin, which lies in the transition zone between Inner Mongolia Plateau and North China Plain (figure 1). Administratively it belongs to Fengning Manchu Autonomous County, Hebei Province. The northwest part of the watershed lies along the southern border of the Inner Mongolia plateau, the southeast part along

![Figure 1. Location and land use map of the study area.](image-url)
the northern foothills of Yanshan Mountains, and the outlet is in the southeast. Elevation decreases from northwest to southeast and ranges from 2213 m to 650 m, with a mean of 1164 m. The climate of the area is the transition zone from semi-humid to semi-arid continental monsoon climate in a temperate zone. The mean annual air temperature of this area is approximately 6.8°C (the coldest -11.7°C in January and warmest 22.4°C in July), its mean annual precipitation and mean annual runoff are 457.1 mm and 54.9 mm, respectively. The area of the watershed is 1850 km², in which there is 26.8% cultivated land, 54.7% forestland, 16.5% grassland, 0.6% water body, 0.7% built-up land, 0.5% sandy land, and 0.2% others (figure 1). Rough terrain covers 91.6% of the watershed, with 12.3% slopes ranging from 0 to 5°, 43.8% from 5° to 15°, 29.2% from 15° to 25° and 6.3% slopes larger than 25°. Located in a farming-pastoral zone of North China, the soil erosion and desertification of land in the northwest part of the watershed are severe because of long time overgrazing and land reclamation. In addition, since the 1950s, the inflow of Minyun reservoir has shown a tendency to decrease (Gao et al. 2002), which has threatened the drinking water security of Beijing. The research on ET over this region is obviously significant and necessary for the issues of inflow decrease and drinking water supply for Beijing.

3. General introduction of the SEBAL model

SEBAL is an intermediate approach to estimate ET for a horizontal surface using both empirical relationships and physical parameterizations based on the energy balance in the vertical direction, assuming that advection and the light energy required for photosynthesis are negligible. This model has been designed to compute energy partitioning at the regional scale from satellite data in conjunction with minimum ground data. The latent heat flux is computed as the residual of energy balance that can be expressed by the following equation:

$$\lambda E = R_n - G - H$$

where $\lambda$ is the latent heat of vaporization (J kg$^{-1}$); $E$ is the evapotranspiration rate (kg m$^{-2}$ s$^{-1}$); $R_n$ is the net radiation flux density (w m$^{-2}$); $G$ is the soil heat flux density (w m$^{-2}$) and $H$ is the sensible heat flux density (w m$^{-2}$).

In the SEBAL model, surface albedo, NDVI and surface temperature could be derived from remotely sensed data. Semi-empirical relationships are used to estimate emissivity from NDVI. The atmospheric transmittance and atmospheric emissivity could be estimated from meteorological data, such as the air temperature, humidity, etc. The elevation, slope and azimuth of land surface could be extracted and computed from the digital elevation model (DEM). Parameters mentioned above in the combination of auxiliary data, such as satellite overpass time and Julian day number, can be used to compute the instantaneous net radiation.

Soil heat flux is smaller compared to the other three energy balance components, but for the low fractional vegetation cover area, the soil heat flux cannot be neglected. Empirical relationships are generally used to estimate soil heat flux from net radiation, albedo, surface temperature and NDVI.

The treatment of sensible heat flux is the critical issue for retrieving latent heat flux based on the residual method of the energy balance equation. In order to estimate the sensible heat flux, an iterative feedback procedure which combines the sensible heat flux with the wind speed and the underlying surface characteristics
(such as vegetation height, zero plane displacement and roughness length, etc.) is employed. The determination of wet and dry surfaces in the study area is required to extract the threshold values which are used to compute the sensible heat flux. SEBAL assumes that temperature difference between heat source and reference height is linearly proportional to the surface temperature $T_s$ and can be developed using two extreme pixels, termed the coldest pixel and the warmest pixel. The coldest pixel is usually selected from land surface where the fractional vegetation cover is high and water supply is sufficient based on the assumption that at the limiting case the sensible heat flux for the coldest pixel is zero. For the warmest pixel, the latent heat flux is assumed to be zero. For unstable conditions, Monin–Obukhov’s similarity hypothesis is used by the model to correct the friction velocity and aerodynamic resistance.

Assuming the evaporative fraction to be nearly constant during the daytime, the daily actual ET can be estimated from daily net radiation and instantaneous evaporative fraction.

4. Data source and flow diagram of the model

The Landsat5 Thematic Mapper (TM) data were acquired at 10:41 a.m. local time on June 23, 2005. The spatial resolution is 30 m × 30 m for the reflectance bands of TM, 120 m × 120 m for the thermal band, and 100 m × 100 m for DEM. The air temperature, the wind speed and the actual vapour pressure were obtained from eight meteorological stations shown in yellow in figure 1 and were interpolated for the whole watershed region taking into account the influences of terrain and elevation. The vegetation height was estimated from leaf area index (LAI) derived from red and infrared bands of TM. The flow diagram of the inputs, data processing and outputs of this study are presented in figure 2.

5. Algorithm for ET estimation

5.1 The instantaneous net radiation

The instantaneous net radiation can be written as

$$R_n = (1 - r)S_{in} + (L_{in} - L_{out}) - (1 - \varepsilon)L_{in},$$  \hspace{1cm} (2)

where $r$ is the surface albedo which could be estimated from the reflectance of each band (Bands 1, 2, 3, 4, 5 and 7 of Landsat5 TM) (Bastiaanssen et al. 1998a, Bastiaanssen 2000, Wang et al. 2000, Chen and Ohring 1984, Koepke et al. 1985); $S_{in}$ is the incoming short wave radiation (w m$^{-2}$); $L_{in}$ is the incoming long wave radiation (w m$^{-2}$); $L_{out}$ is the outgoing long wave radiation (w m$^{-2}$); and $\varepsilon$ is the surface emissivity, which could be estimated by NDVI retrieved from the red and infrared bands (Bastiaanssen 1995, Bastiaanssen et al. 1998a). The incoming short wave radiation is a function of solar elevation, solar radiation intensity, atmospheric transmittance, topographic correction factor, etc. In this study, the incoming short wave radiation of mountainous terrain was computed by equation (3) as follows (Fu 1983, Tasumi et al. 2000) in which the diffuse radiation was neglected:

$$S_{in} = \frac{I_0}{d^2} \cos(i) \times \tau$$  \hspace{1cm} (3)

where $I_0$ is the solar constant (1367 w m$^{-2}$); $d$ is the earth–sun distance in
astronomical units; and $\tau$ is the one way transmittance ($0.75 + 2 \times 10^{-5} \times$ elevation in metres) (Tasumi et al. 2000).

\[
\cos(i) = \sin \delta (\sin \varphi \cos z - \cos \varphi \sin z \cos \beta) + \cos \delta \cos \omega (\cos \varphi \cos z + \sin \varphi \sin z \cos \beta) + \cos \delta \sin \beta \sin z \sin \omega
\]

in which $\varphi$ is the geographic latitude (rad); $\delta$ is the solar declination (rad); $\omega$ is the solar angle ($= \pi (t-12)/12$, $t$ is the local time) (rad); $z$ is the slope (rad); and $\beta$ is the azimuth (rad) (from due south, clockwise positive value, counterclockwise negative value).

\[
L_{in} = \varepsilon_a \sigma T_a^4
\]

\[
\varepsilon_a = 1.24 \left( \frac{\varepsilon_a}{T_a} \right)^{1/7}
\]

where $\varepsilon_a$ is the atmospheric emissivity (Brutsaert 1975); $\sigma$ is the Stefan–Boltzman constant ($5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$); $\varepsilon_a$ is the actual vapour pressure (hPa); and $T_a$ is the atmospheric mean temperature (K).

\[
L_{out} = \varepsilon \sigma T_s^4
\]

where $T_s$ is the surface temperature (K), which could be retrieved by the thermal infrared band (Band 6) (Bastiaanssen 2000, Chander and Markham 2003).
5.2 The instantaneous soil heat flux density

The soil heat flux density can be estimated by equation (8) for vegetated surfaces (Bastiaanssen 1998),

\[
G = \frac{R_n T_s (0.0032r + 0.0062r^2) (1 - 0.978NDVI^4)}{r}
\]  

(8)

but for bare soil in our study area it is taken as 0.3 times the net radiation:

\[
G = 0.3R_n
\]  

(9)

5.3 The instantaneous sensible heat flux density

The sensible heat flux density can be expressed by the following equation:

\[
H = \rho c_p (T_{aero} - T_2) / r_{ah}
\]  

(10)

where \( \rho \) is the air density (kg m\(^{-3}\)); \( c_p \) is the air specific heat at constant pressure (J kg\(^{-1}\) K\(^{-1}\)), \( T_{aero} \) is the aerodynamic temperature (K); \( T_2 \) the air temperature at reference height (commonly 2 m above canopy top); and \( r_{ah} \) is the aerodynamic resistance for heat transport (s m\(^{-1}\)) given by:

\[
r_{ah} = \left[ \ln\left( \frac{z - d}{z_m} \right) - \psi_m \right] \left[ \ln\left( \frac{z - d}{z_h} \right) - \psi_h \right] / k^2 u
\]  

(11)

in which \( z \) is the reference height (m); \( d \) is the zero plane displacement (m); \( z_m \) and \( z_h \) are the roughness length for momentum transport (m) and heat transport (m), respectively; \( u \) is the velocity at the reference height (m s\(^{-1}\)), \( k \) is the von Karman constant; and \( \psi_m \) and \( \psi_h \) are the stability correction factor for momentum and sensible heat, respectively.

The value of wind velocity at the meteorological stations were converted into the value at 200 m height where the atmospheric pressure is stable, assuming that the friction velocity \( u^* \) is constant with height over the meteorological stations and the wind speed at 200 m above the land surface is uniform over the satellite image. \( d, z_m \) and \( z_h \) were estimated by the formulas below (Allen et al. 1998):

\[
d = 0.667h_c
\]  

(12)

\[
z_m = 0.123h_c
\]  

(13)

\[
z_h = 0.1z_m
\]  

(14)

where \( h_c \) is the vegetation height (m), which can be estimated by the vegetation index (Zhang 1996, Kimura et al. 2007) and our field investigations.

Stability correction factors for momentum and sensible heat for an unstable atmosphere are expressed as (Paulson 1970):

\[
\psi_m = 2 \ln\left( \frac{1 + \sqrt{x}}{2} \right) + \ln\left( \frac{1 + \sqrt{x^2}}{2} \right) - 2 \tan^{-1}(\sqrt{x}) + \frac{\pi}{2}
\]  

(15)

\[
\psi_h = 2 \ln\left( \frac{1 + \sqrt{y}}{2} \right)
\]  

(16)
\[ x = (1 - 16\Gamma)^{1/4} \]  
\[ y = (1 - 16\Gamma)^{1/2} \]

where \( \Gamma \) is the atmospheric stability and is a function of Monin–Obukhov length, expressed by equations (19) and (20):

\[ \Gamma = \frac{z}{L} \]  

in which \( L \) is the Monin–Obukhov length defined by

\[ L = -\frac{\rho c_p u^3 T_2}{kgH} \]

where \( u^* \) is the local scale friction velocity (m s\(^{-1}\)) and \( g \) is the acceleration of gravity (9.8 m s\(^{-2}\)).

As the atmospheric stability factors are also the function of the sensible heat flux density \( H \), in order to estimate \( H \), an iterative feedback procedure is used. First, the atmosphere is assumed to be in a stable condition, in this case, \( \psi_m \) and \( \psi_h \) are both zero. The difference between the aerodynamic temperature and the reference height temperature can be obtained by the warmest and coldest pixels (Bastiaanssen et al. 1998a, 1998b). Therefore the first estimation of \( H \) can be computed and is then used to obtain the stability correction factors for the second estimation of \( H \). The loop of equations (10), (11), (15)–(20) is repeated until the convergence in aerodynamic resistance of the target pixels is reached.

### 5.4 The improved daily net radiation parameterization scheme and the daily actual evapotranspiration

After having computed the net radiation, the soil heat flux density and the sensible heat flux density by the previous steps, the latent heat flux density can be obtained as the residual of the energy balance equation. Assuming the evaporative fraction to be nearly constant during the daytime, the daily actual ET can be estimated from daily net radiation and evaporative fraction as follows:

\[ \Lambda = \frac{\lambda E}{\lambda E + H} = \frac{R_n - G - H}{R_n - G} \]  
\[ \text{ET} = \frac{86400 \times \Lambda \times R_{day}}{\lambda} \]

where \( \Lambda \) is the evaporation fraction; ET is the daily actual evapotranspiration (mm day\(^{-1}\)) and \( R_{day} \) is the daily surface net radiation (w m\(^{-2}\)).

The estimated instantaneous components of the energy balance are shown in table 1 for the southeast, southwest, northeast and northwest facing surfaces at the satellite overpass time. The results indicated that the southeast facing surfaces received the largest radiation as they were closest to the solar azimuth and the solar elevation was near the solar angle at noon. As both southwest and northeast facing surfaces distributed on both sides of the southeast are perpendicular to the solar azimuth, their net radiation is smaller than that of the southeast facing surfaces. Moreover, because the northeast facing surfaces are closed to the north direction, its
net radiation is slightly smaller than that of the southwest facing surfaces. Opposite to the southeast, the northwest facing surfaces receive the least net radiation. The spatial distributions of soil heat flux, sensible heat flux, and latent heat flux are similar to that of the net radiation, indicating that the larger the radiation the land surface receives, the larger amount of heat and water fluxes it has. The results shown above proved not only that the net radiation is the main driving source of ET processes but also that the instantaneous fluxes retrieved from SEBAL are in agreement with reality.

The daily net radiation can be expressed as:

\[ R_{\text{day}} = (1 - r)K_{24} + L_{24} \]  

(23)

where \( K_{24} \) is the daily solar radiation (\( \text{W m}^{-2} \)) and \( L_{24} \) is the daily net long wave radiation (\( \text{W m}^{-2} \)).

As the terrain of our study area is complex with undulating topography, the impact of slope and azimuth of surface on available radiation should be considered pixel by pixel in the calculation of instantaneous and daily net radiation. When computing the instantaneous net radiation, the shaded areas (pixels) were excluded from imageries with the aid of the HILLSHADE function in ARCGIS through importing the parameters of solar azimuth and solar elevation at the satellite overpass time. The daily net radiation is estimated by an integral of equation (3) from sunrise angle to sunset angle and by substituting daily atmospheric transmittance for one-way transmittance \( \tau \) with \( (a + bn/N) \):

\[
K_{24} = \left( a + \frac{bn}{N} \right) \cdot \frac{I_0}{2\pi d^2} \int_{\omega_1}^{\omega_2} \left[ \sin \delta \cdot u + \cos \delta \cos \omega \cdot v + \cos \delta \sin \beta \sin x \sin \omega \right] d\omega \\
= \left( a + \frac{bn}{N} \right) \cdot \frac{I_0}{2\pi d^2} \left[ \sin \delta \cdot u + \cos \delta \cos \omega \cdot v + \cos \delta \sin \beta \sin x \sin \omega \right] \\
= \left( a + \frac{bn}{N} \right) \cdot \frac{I_0}{2\pi d^2} \left[ u \sin \delta (\omega_2 - \omega_1) + v \cos \delta (\sin \omega_2 - \sin \omega_1) \\
- \sin \beta \sin x \cos \delta (\cos \omega_2 - \cos \omega_1) \right] \\
\]

\[ u = \sin \varphi \cos x - \cos \varphi \sin x \cos \beta \]  

(25)

\[ v = \cos \varphi \cos x + \sin \varphi \sin x \cos \beta \]  

(26)

\[ N = \frac{12(\omega_2 - \omega_1)}{\pi} \]  

(27)

<table>
<thead>
<tr>
<th>Net radiation (( \text{W m}^{-2} ))</th>
<th>Southeast azimuth</th>
<th>Southwest azimuth</th>
<th>Northeast azimuth</th>
<th>Northwest azimuth</th>
</tr>
</thead>
<tbody>
<tr>
<td>669.3</td>
<td>633.1</td>
<td>629.5</td>
<td>589.5</td>
<td></td>
</tr>
<tr>
<td>Net radiation with slope&gt;30°</td>
<td>721.3</td>
<td>567.1</td>
<td>577.5</td>
<td>410.9</td>
</tr>
<tr>
<td>Net radiation with slope&lt;30°</td>
<td>668.7</td>
<td>633.8</td>
<td>630.3</td>
<td>590.9</td>
</tr>
<tr>
<td>Soil heat flux (( \text{W m}^{-2} ))</td>
<td>82.3</td>
<td>74.2</td>
<td>74.7</td>
<td>65.4</td>
</tr>
<tr>
<td>Sensible heat flux (( \text{W m}^{-2} ))</td>
<td>235.7</td>
<td>207.9</td>
<td>220.2</td>
<td>186.6</td>
</tr>
<tr>
<td>Latent heat flux (( \text{W m}^{-2} ))</td>
<td>351.3</td>
<td>351.0</td>
<td>334.6</td>
<td>337.5</td>
</tr>
</tbody>
</table>

Table 1. The instantaneous fluxes of four azimuthal directions.
where $a$ and $b$ are the coefficients of the solar radiation depending on the latitude, climate and other factors of study area, respectively; $a+b$ is the fraction of extraterrestrial radiation reaching the earth on clear sky days. According to the research from Chen et al. (1995), we take $a=0.17$, $b=0.54$; $n$ is the actual sunshine duration, $N$ the potential sunshine duration, $\omega_1$ and $\omega_2$ are the sunrise and sunset angle, respectively. The difficulty in retrieval of the daily solar radiation focuses on calculation of the sunrise and sunset angle for the tilted surfaces. The sunrise and sunset angles for horizontal surfaces are given by Fu (1983) and Tasumi et al. (2000)

$$\omega_H = \cos^{-1}(\tan \varphi \tan \delta)$$

(28)

The sunrise and sunset angle for tilted surfaces can be obtained by a simple mathematical manipulation from equation (4) by setting $\cos(i)=0$, leading to

$$\omega = \cos^{-1}\left(-uv \tan \delta \pm \sin \beta \sin \varepsilon \sqrt{1-u^2(1+\tan^2 \delta)} \right)$$

(29)

The positive or negative sign of equation (29) in the numerator are determined by equation (30):

$$\omega = \sin^{-1}\left(-u \sin \beta \sin \varepsilon \tan \delta \pm v \sqrt{1-u^2(1+\tan^2 \delta)} \right)$$

(30)

Let $\omega_{s1}$ and $\omega_{s2}$ be the roots of $\omega$, respectively, and $\omega_{s2} > \omega_{s1}$. Note that the surface receives the solar radiation only if $\cos(i)$ in equation (4) is greater than 0. Several relationships are given below to determine the sunrise and sunset angles ($\omega_1$, $\omega_2$):

- If $\omega_{s1} \leq \omega \leq \omega_{s2}$ and $\cos(i) \geq 0$, then $\omega_1 \leq \omega_{s1}$, $\omega_2 \leq \omega_{s2}$;
- If $\omega < \omega_{s1}$, $\omega > \omega_{s2}$ and $\cos(i) > 0$, then $\omega_{s1} < \omega_{s1}$, $\omega_{s2} > \omega_{s2}$.
- Meanwhile, the sunrise and sunset angles for tilted surfaces must also satisfy the condition that sunrise is no earlier and sunset is no later than those for horizontal surfaces. Namely $\omega_1 \geq |\omega_H|$,$\omega_2 \leq |\omega_H|$.

The results of $\omega_1$ and $\omega_2$ for our study are shown in figure 3.

Daily net long wave radiation:

$$L_{24} = \overline{v_a} \sigma T_a^4 - \varepsilon \sigma T_s^4$$

(31)

$$\overline{v_a} = 9.2 \times 10^{-6} T_a^2$$

(32)

where $\overline{v_a}$ is the daily average atmospheric emissivity (Campbell and Norman 1998) and $T_a$ is the daily mean atmospheric temperature (K). Due to the surface temperature obtained at 10h30, it could represent the daily average surface temperature for estimation of the surface daily long wave radiation (Granger 2000).

6. Results and validation

The actual daily ET on June 26, 2005 for our study area is derived using the method proposed above. The distribution and frequency of the estimated ET are displayed in figure 4. The mean of the actual daily ET on 26 June 2005 for the whole watershed is 3.4 mm and its standard deviation is 1.2 mm.
6.1 Accuracy assessment and error analysis for instantaneous latent heat flux

There is a small reservoir located in the northern part of the watershed (see figure 5). The actual instantaneous latent heat fluxes for this reservoir are estimated from the Penman equation (Penman 1948) which was used for many years to estimate the open water evaporation and the potential ET from saturated surfaces. The correlation analysis is performed between the actual instantaneous latent heat fluxes

Figure 3. Sunrise and sunset angles (rad) of the study watershed on DOY 174, 2005.

Figure 4. ET (mm) and its frequency (%) distribution of the study watershed on 23 June 2005.
derived from the Penman equation (denoted as $\lambda E_{\text{penman}}$) and retrieved from the SEBAL model (denoted as $\lambda E_{\text{retrieved}}$). In figure 6, the retrieved $\lambda E_{\text{retrieved}}$ is plotted against the $\lambda E_{\text{penman}}$. The 1:1 line is also plotted in the graphs. The mean absolute percent differences (MAPD) between $\lambda E_{\text{retrieved}}$ and $\lambda E_{\text{penman}}$ are quantitatively calculated using:

$$\text{MAPD} = \frac{1}{n} \sum_{i=1}^{n} \left( \left| \frac{\lambda E_{\text{retrieved}(i)} - \lambda E_{\text{penman}(i)}}{\lambda E_{\text{penman}(i)}} \right| \times 100\% \right)$$  (33)

where $i$ is the water surface pixel number and $n$ is total number of water pixels.

Figure 5. Location of the reservoir and the irrigated spring maize.

Figure 6. Validation of the retrieved latent heat flux against calculation results by Penman-Monteith together with a 1:1 line.
The results showed that the retrieved instantaneous latent flux ($\lambda E_{\text{retrieved}}$) is in good correlation with the latent heat fluxes derived by the Penman equation ($\lambda E_{\text{penman}}$) ($R^2=0.852$, standard deviation = 24.6 w m$^{-2}$) and the retrieved accuracy is relatively high with MAPD = 4.6%. $\lambda E_{\text{retrieved}}$ were slightly higher than $\lambda E_{\text{penman}}$, and the larger the $\lambda E$, the larger the difference between $\lambda E_{\text{retrieved}}$ and $\lambda E_{\text{penman}}$. This difference could result from substituting observed daily mean wind speed for wind speed at a reference height at the time of satellite overpass. As the reservoir is located in the outlet of the valley, turbulent exchanges in the daytime are probably stronger than that at night and the wind speed at water surface reference height in the daytime is generally higher than the daily mean value. Therefore, the results from the Penman–Monteith equation are slightly lower than the actual latent heat flux, implying that the difference between $\lambda E_{\text{retrieved}}$ and actual latent heat fluxes would be even smaller.

### 6.2 Accuracy assessment and error analysis for daily actual evapotranspiration

The actual ET of crops can be estimated directly from empirical relationships. Many theoretical or empirical approaches commonly use the crop potential evapotranspiration ($ET_0$) and the soil crop coefficient $K_{sc}$ to estimate crop actual evapotranspiration ($ET_{act}$), namely (Lei et al. 1988, Abdelhadi et al. 2000):

$$ET_{act} = K_{sc} ET_0$$  \hspace{1cm} (34)

Here soil crop coefficient $K_{sc}$ characterizes the integral effects of the processes of soil water supply and crop evapotranspiration. It can be described by:

$$K_{sc} = K_{s0} K_{c0}$$  \hspace{1cm} (35)

in which $K_{s0}$ is the soil water supply coefficient and $K_{c0}$ is the crop coefficient. $K_{s0}$ characterizes the effect of soil moisture on crop evapotranspiration. When the soil water is sufficient, $K_{s0} = 1$; under normal circumstance, $K_{s0} < 1$. The influences of soil moisture on crop ET are very complicated. In order to eliminate the constraint of soil water supply, the irrigated spring maize distributed on both sides of the river terrace with sufficient soil water supply was selected as the study target (figure 5). Consequently, the soil water supply coefficient $K_{s0} \approx 1$ and the crop actual ET is determined by both crop coefficient $K_{c0}$ and crop potential ET $ET_0$. The FAO56 Penman–Monteith equation, adopted in this study, has been widely applied to numerous countries because of its feasibility to estimate crop reference ET ($ET_0$) (Allen et al. 1989, Allen 2000, Kite and Droogers 2000, Jensen et al. 1990).

Combined with the meteorological data, the reference ET ($ET_0$) on DOY 174 (26 June 2005) is estimated to be 5.4 mm. The seedtime of the spring maize in the paddy field is April 15, 2005 and its growth length is 125 days. DOY 174 is at the mid-term growth stage, the 59th in 125 days. On the basis of the similar climatic, geographical and farming system conditions, the experimental data (figure 7) from Jianping county, located in the western part of Liaoning province (40°17′–42°21′ N latitude, 119°10′–120°2′ E longitude), is used as the crop coefficient in our study area. The crop coefficient of spring maize is obtained as 0.74 on DOY 174. Considering the prevailing drought which occurred over the watershed in 2005, it is plausible or even realistic to set the crop coefficient as 0.7 for our study area. Consequently, the daily actual ET of irrigated spring maize is: $ET_{act} = K_{c0}$.
Using the average daily actual ET (3.5 mm) derived by SEBAL for the irrigated spring maize shows that the error between the retrieved and the actual ET is about 7.9%.

7. Analysis and discussion

7.1 Relationship between ET and land use

From the results shown in table 2 we can draw a conclusion that the retrieved ET obviously differs in different types of land cover. The higher values of ET correspond to the reservoir, forest and paddy land, especially for the underlying surface with a specific terrain which receives more energy. These surfaces have generally lower surface temperature and smaller sensible heat flux, and consequently have higher ET values. For woodland, ET values of forest (4.3 mm) are higher than those of shrub (3.8 mm) and woods (3.3 mm) depending on the vegetation fraction.

![Crop coefficient curve of spring maize in Jianping County (Chen et al. 1995).](image)

Table 2. Daily actual ET of different land use types on DOY=174.

<table>
<thead>
<tr>
<th>Land use classification (1st level classes)</th>
<th>Land use classification (2nd level classes)</th>
<th>Pixel number</th>
<th>Area percent (%)</th>
<th>Retrieved daily ET (mm)</th>
<th>Retrieved daily ET (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cropland</td>
<td>Paddy land</td>
<td>3805</td>
<td>26.8</td>
<td>3.2</td>
<td>2.7</td>
</tr>
<tr>
<td></td>
<td>Flat dry land</td>
<td>170459</td>
<td>2.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Slope dry land</td>
<td>626467</td>
<td>2.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Woodland</td>
<td>Forest</td>
<td>613816</td>
<td>54.7</td>
<td>4.3</td>
<td>3.9</td>
</tr>
<tr>
<td></td>
<td>Shrub</td>
<td>690010</td>
<td>3.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Woods</td>
<td>327709</td>
<td>3.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Others</td>
<td>1737</td>
<td>3.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Grassland</td>
<td>Dense grass</td>
<td>298837</td>
<td>16.5</td>
<td>3.2</td>
<td>3.4</td>
</tr>
<tr>
<td></td>
<td>Moderate grass</td>
<td>194489</td>
<td>3.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Water body</td>
<td>Reservoir and ponds</td>
<td>160</td>
<td>0.6</td>
<td>6.0</td>
<td>6.0</td>
</tr>
<tr>
<td></td>
<td>River-beach</td>
<td>18528</td>
<td>2.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Built-up land</td>
<td>Urban built-up</td>
<td>4682</td>
<td>0.2</td>
<td>0.9</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Rural settlements</td>
<td>15601</td>
<td>1.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sandy land</td>
<td>Sandy land</td>
<td>14899</td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td></td>
<td>Others</td>
<td>6145</td>
<td>4.5</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
cover. A highly developed root system often associates with high fractional vegetation cover which has much stronger transpiration than that of spare canopy and moderate and low fractional vegetation cover areas. Considering that the comprehensive influences of meteorological factors (the relative humidity was merely 61%; atmospheric temperature was relatively high, daily maximal temperature was 34.4°C, daily mean temperature was 25.6°C. Both of them were the second maximum in June, 2005; the observed wind speed was also the second maximum value in June, 2005), the transpiration speed of woodland is rapid in general; therefore the amount of forest ET is higher.

The lower ET estimates occur in the northwest and southeast parts of the watershed. This is due to the fact that higher surface temperature results in a larger sensible heat flux based on the assumption that the temperature difference between surface and surface layer is taken to be proportional to the remotely sensed surface temperature. The built-up land with high impervious surfaces and low fractional vegetation cover, called non-evaporative surface, is dominant over the southeast part of the watershed, leading to the extraordinarily low soil moisture availability and high surface temperature in the daytime. Thereby, the energy transformation is mainly in the form of sensible heat exchange. A similar reason exists for the northwest part of the watershed which is predominantly sandy land and rocky bare hills. Moreover, low ET values are found for the dry cropland. They are probably related to the low soil moisture availability. According to the data recorded from meteorological and hydrological stations in the watershed, there were no rains before DOY 174 or little rain (there was 0.6 mm rainfall in Fengning meteorological station and no rainfall in other meteorological and hydrological stations), and both the daily maximum temperature and daily mean temperature from DOY 172 to DOY 174 were the highest values in June 2005. The conclusion is that dry air and rapidly decreasing soil moisture before DOY 174 caused low actual ET over dry cropland. Low ET values of river-beach result from extremely low discharges and bed material exposure.

As for the grassland, ET estimates of moderate grass (3.6 mm) exhibit counterintuitively slightly larger values than those of dense grass (3.2 mm), which could be explained by the limitations of SEBAL when applied to moderate grass having complicated ET processes. SEBAL pertains to the single-source model taking the soil and vegetation as a single big leaf and therefore cannot depict vegetation transpiration and soil evaporation separately. The single model simplifies the processes of heat flux exchanges between the surface layer and land surface by considering the exchange height of sensible and latent heat flux (roughness length for heat and water transfer) to be uniform and the components (vegetation and soil) to be isothermal, which may induce the deviation under sparse canopy or low vegetation fraction cover conditions. For instance, if the soil surface is very dry and the vegetation fraction cover is very low, the sensible heat source is always at the soil surface, whereas the water vapour source is close to the top of the canopy, indicating different roughness lengths for heat and water, and the difficulty in unambiguous definition of the roughness length for heat transfer. Although the model was designed to provide the land surface fluxes at regional scales using remote sensing and minimum ground data, the disparity between estimates and reality may be caused by the simplification of natural processes, specifically the physical mechanism of water–heat transfer in the soil—vegetation—atmosphere continuum.
7.2 **Relationship between ET and terrain**

The forest land, the only land cover type distributing in the tilted land in the watershed with various slopes and azimuth angles, is selected to explore the relationship between ET estimates and terrain. Figure 8 illustrates that there is a notable spatial heterogeneity in ET from different slopes and azimuths. Comparing with Figure 9, it is obvious that the net radiation is a primary manipulative factor of ET spatial distribution and the main driving force of ET processes.

For the tilted land whose azimuth angle ranges from $135^\circ$ to $315^\circ$, the maximum daily net solar radiation occurs in the slope ranging from $15^\circ$ to $20^\circ$. This range of

![Figure 8. The retrieved daily ET (mm) of forest land of various slopes ($\alpha$) and azimuth ($\beta$) (from due north, $\beta$ clockwise ranges from $0^\circ$ to $360^\circ$).](image)

![Figure 9. The daily solar net radiation ($\text{w m}^{-2}$) of various slopes ($\alpha$) and azimuth ($\beta$).](image)
slope is also called ‘the warmest slope’ related to the latitude, azimuth and solar declination. When the azimuthal direction is closed to the east or west, the warmest slope becomes larger.

If the slope was smaller than the warmest slope, the daily net solar radiation increases with increasing slope, showing the larger available solar radiation compared to the flat surfaces. If the slope was larger than the warmest slope, the daily net solar radiation decreases with increasing slope. For the azimuth angles ranging from 0° to 135° and 315° to 360°, the available solar radiation for all slopes is smaller than that of the flat surfaces. Compared to azimuths closest to the southern direction, decreases in speed and in amplitude of available solar radiation for azimuths closest to the northern direction are larger with increases in slope. Similarly, the ET distribution also follows this rule.

The differences in daily net solar radiation for various azimuths under certain slopes ranging from 0° to the warmest slopes. The daily net solar radiation for azimuths closest to the southwest is larger than any other azimuth, which coincides with the fact that the thermal inertia and temperature are highest in this specific period of time when the solar azimuth is closest to the southwest in a day.

The daily net solar radiation for azimuths closest to the north is not apparently small under gentle slopes because DOY 174 was nearly closest to the summer solstice in 2005 (DOY 172). On DOY 172, the sunrise azimuth was nearest to the northeast and the sunset azimuth was nearest to the northwest. Therefore, the potential sunshine duration for north azimuths was not short. However, the daily net radiation for steep slopes decreases rapidly with increasing slope. The spatial distribution of ET estimates is similar to that of the daily net solar radiation. ET estimates for azimuths closest to the southern direction are approximately two times the ET estimates for azimuths closest to the northern direction for steep slopes.

### 7.3 Relationship between ET and elevation

Exploring the relationship between ET and elevation is of importance to characterize the spatial distribution of ET, to better understand how mountainous ecosystems work, and to assess how land use interacts on the water quantity of watersheds. The ET estimates from SEBAL for woodland and grassland at all elevations are shown in figure 10. There exists no direct relationship between the elevation and ET estimates for the north facing woodland or grassland which corresponds to high availability of the soil moisture and insignificant difference in the daily net radiation. On the contrary, for the south facing woodland and grassland, the ET estimates, in general, tend to decrease at higher elevation, which is thought to be a consequence of deficiency in soil moisture and daily net radiation. The dry air condition mentioned previously is a probable reason to explain why the soil moisture of the south facing grassland is lost rapidly. The ET estimates of this south facing grassland at elevations ranging from 1200 m to 1600 m are rather lower than that at elevations ranging from 1600 m to 1800 m mainly due to the difference in the averaged LAI (for elevation ranging from 1200 m to 1400 m, the LAI is 0.58, from 1400 m to 1600 m, the LAI is 0.67, from 1600 m to 1800 m, the LAI is 0.78), which disturbs the tendency of decreasing ET estimates when elevation increases. The ET estimates for the south-oriented slopes are curiously lower than those for the north-oriented slopes. This may be due to the lower averaged evaporative fraction (0.635) for the south-oriented slopes (0.646 for the north-oriented slopes) even
though the daily net radiation for south-oriented slopes (174.39 $\text{W m}^{-2}$) is slightly higher than that for the north-oriented slopes (173.46 $\text{W m}^{-2}$). Thereby the soil moisture also contributes significantly to ET processes.

### 7.4 Uncertainty analysis

Despite SEBAL’s ability to capture major characteristics of ET distribution at the watershed scale, there are uncertainties arising from the assumptions and the operation of the model. The single source model cannot capture the transpiration of vegetation and evaporation from the soil surface separately, yielding discrepancies between retrievals and actual values, especially for sparse canopy or low fractional vegetation cover areas. The computation of sensible heat flux in SEBAL is independent of available energy; thus the uncertainties in wind speed observations and surface temperature retrieved directly from satellite lead to uncertainties in heat flux estimates. The assumption that there exists a linear relationship between the surface temperature and the temperature difference may lead to uncertainties. In addition, selection of the warmest pixel and the coldest pixel are significantly affected by factors such as thin cloud and hill shade. The extension for instantaneous latent heat flux to daily actual ET in terms of presumptive constant evaporative fraction may produce errors since the evaporative fraction varies with time to some extent in a day, especially for the tilted land with heterogeneity in underlying surfaces.

Inspecting all input and output maps, it was observed occasionally that some ET estimates for the north facing surface with steep slopes were larger than that for the south facing surface with gentle slopes. For the south facing tilted land surface, although the potential ET is generally large because the surface receives more energy, the soil moisture is rapidly lost under dry air conditions; hence ET values are not necessarily higher than those for the north facing surface for which the soil moisture is sufficient. The differences in turbulence exchange and soil heat fluxes for tilted land with various slopes and azimuths are more pronounced than those for flat
terrain and therefore advection and local circulation can be formed under conditions of complex terrain and heterogeneous underlying surface, for which the energy balance approach in the vertical direction is not effective in the case of significant advection and local circulation. As the distributions of wind speed and rainfall also vary with terrain, rendering significant variation in the soil moisture, each component of the energy balance equation becomes more complicated. Additionally, uncertainties associated with interpolation of both the wind speed and the air temperature, resampling for matching different data sources, and errors with respect to meteorological records, could be difficult to evaluate and will not be discussed in this paper.

8. Conclusions

The remote sensing technique provides a mean to capture the land surface characteristics with respect to models for land surface flux prediction, in which the surface temperature, surface albedo, surface emissivity and leaf area index are of significant importance to estimate ET at regional scales. Due to improvement of the parameterization schemes for calculating the instantaneous and daily net solar radiation required in SEBAL, combined with remote sensing and meteorological data, and DEM, the detailed spatial distribution of ET from complex terrains was obtained. Without quantifying the various sunrise and sunset angles of slope terrain for accurate determination of the instantaneous and daily net radiation, the retrieval of surface net radiation and ET in such detail would not be possible for complex terrains. The warmest slope was proven and demonstrated simultaneously by the spatial distribution of retrieved solar net radiation and ET estimates of woodland. The ET estimates of south facing woodland and grassland tended to decrease with increasing elevation, which reflected the fact that soil moisture plays an important role in ET processes. As ET encompasses the comprehensive effects of underlying surface, soil moisture, meteorological factors, the ET of the south facing surfaces is not necessarily larger than that of north facing surfaces. Due to the complicated characteristics of the land surface and the surface layer variables, the magnitude of spatial variation in ET across the diverse types of land cover were quite large, especially for the tilted surfaces where the difference in net radiation was considerably big, and advection and local circulation could not be neglected. The variation in ET estimates of woodland and grassland with different fractional vegetation cover implied that a change in fractional vegetation cover will lead to a change in processes and mechanisms of water–heat transfer, and the single-source model performs well for high fractional vegetation cover area but poorly for moderate and low covers. The explicit spatial distribution of ET could be used for various purposes, including estimating soil moisture and predicting runoff, and other water-related research on watersheds.

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