Energy and Water Cycle over the Tibetan Plateau
Surface Energy Balance and Turbulent Heat Fluxes

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Abstract This contribution presents an overview and an outlook of studies on energy and water cycle over the Tibetan plateau with an emphasis on the estimation of energy balance terms and turbulent heat fluxes. On the basis of the surface energy balance calculations, we show that the phenomena of energy imbalance exist in GAME/Tibet experiment data; although the explanations for the reasons are debated now and not resolved yet. We found that the derived latent heat flux is much higher than the measurements. However, the corrected measurements, which are calculated according to the hypothesis of the energy balance, compare very well with the estimation of SEBS. On this basis it is concluded that the deviation is caused by the energy imbalance of ground measurements in GAME/Tibet experiment area. The latent heat fluxes were likely underestimated.

Key words: Energy and water cycle, The Tibetan Plateau, Energy balance, Turbulent fluxes

1 Introduction
The Tibetan plateau is a significant heat and momentum source for the atmosphere; its energy and water cycles play an important role in the Asian Monsoon system which in turn is a major component of both the energy and water cycles of the global climate system. The Tibetan plateau contains the world's highest elevation relief features. Much of them exceed an altitude of 4000 m asl, some reaching into the mid-troposphere. Due to its topographic characteristics the plateau land surface absorbs a large amount of solar radiation and undergoes dramatic seasonal changes of surface heat and water fluxes. Land surface processes on the Tibetan plateau are multifaceted and complex. Due to the highly complex terrain and winter season conditions are characterized by irregular snow cover with extensive areas of frozen ground holding large quantities of moisture in the surface layers. Seasonal freezing and melting processes and their spatial distribution lead to timespace variations of surface wetness and to variations of the surface heat balance. Such variations have profound implications for follow-up monsoon behavior and global climate processes. In short, the land cover in Tibetan plateau shows spatial and temporal variability which will affect the distributions of sensible and
latent heat fluxes. Hence it is very important to investigate the interactions between the land surface and atmosphere over the Tibetan plateau so that we can understand the complete energy and water cycles and their effects on the Asian Monsoon system[1] and further on the global atmospheric circulations.

The atmospheric turbulent fluxes evapotranspiration when latent heat flux is expressed in water depth[1] at the land surface have long been recognized as the most important processes in the determination of the exchanges of energy and mass among the atmospheric and Biosphere[2]. Conventional techniques that employ point measurements to estimate the components of energy balance are representative only of local scales and cannot be extended to large areas because of the heterogeneity of land surfaces and the dynamic nature of heat transfer processes. Remote sensing is probably the only technique that can provide representative measurements of several relevant physical parameters at scales from a point to a continent. Techniques using remote sensing information to estimate atmospheric turbulent fluxes are therefore essential when dealing with processes that cannot be represented by point measurements only.

The Surface Energy Balance System[3] SEBS proposed by Su[4] is a validated algorithm among others to estimate atmospheric turbulent fluxes and evaporation fraction using satellite earth observation data in combination with meteorological information of Primary Boundary Layer in plain areas. The purpose of this study is to clarify whether SEBS is suitable to estimate the energy participation directly on the Tibetan plateau.

2 Energy balance terms and turbulent heat fluxes

2.1 Surface energy balance term

Neglecting advection and heat storage[5] the surface energy balance is commonly written as[6]

\[ R_n = G + H + L + E \]

where \( R_n \) is the net radiation[7], \( G \) is the soil heat flux[8], \( H \) is the turbulent sensible heat flux[9], and \( L \) is the turbulent latent heat flux.

The equation to calculate the net radiation is given by

\[ R_n = R_{dn} - R_{up} \]

where \( R_{dn} \) is the downward solar radiation, \( R_{up} \) the downward long wave radiation, \( g \) the emissivity of the surface, \( C \) the Stefan-Boltzman constant[10], and \( T \) the surface temperature.

The equation to calculate soil heat flux is parameterized as[11]

\[ G = \left( \frac{R_n - R_{dn}}{R_{up}} \right) \left( \frac{1}{1 + \frac{R_{dn}}{R_{up}}} \right) \]

in which we assume the ratio of soil heat flux to net radiation \( \frac{G}{R_n} \) of 0.05 for a full vegetation canopy[12] and \( \frac{G}{R_n} \) of 0.315 for bare soil[13]. An interpolation is then performed between these limiting cases using the fractional canopy coverage \( f \).

In order to derive the sensible and latent heat fluxes we use a model of similarity theory. In Atmospheric Surface Layer[14] the similarity relationships for the profiles of the mean wind speed \( u \) and the mean temperature \( T \) are usually can be written in integral form as

\[ u = \frac{u_0}{k} \left( \frac{z}{z_h} \right)^{1/3} \]

\[ T = \frac{T_0}{k} \left( \frac{z}{z_h} \right)^{1/3} \]

where \( z \) is the height above the surface, \( u_0 \) is the friction velocity, \( k \) is the surface shear stress, \( \rho \) is the density of air, \( k \) is 0.4 is von Karman's constant, \( u_0 \) is the zero plane displacement height, \( z_h \) is the roughness height for momentum transfer, \( T_0 \) is the potential temperature at height \( z_h \), \( z \) is the scalar roughness height for heat transfer, \( u_10 \) and \( u_20 \) are the stability correction functions for momentum and sensible heat transfer respectively, \( L \) is the Obukhov length defined as

\[ L = \frac{c_p u_20}{g} \left( z \right) \]

\[ \frac{1}{\theta} \frac{\partial \theta}{\partial z} = \frac{u_20}{L} \]

\[ \frac{\partial u}{\partial z} = \frac{u_20}{L} \]

where \( g \) is the acceleration due to gravity, \( \theta \) is the potential virtual temperature near the surface.

For field measurements performed at a height of a few meters above ground[15] clearly since the surface flux are related to surface variables and variables in the atmospheric surface layer[16] all calculations use the Mo-
non-Obukhov similarity functions given by Brutsaert \cite{brutsaert1982}. By replacing the MOS stability functions with the Bulk Atmospheric Boundary Layer (BAB) similarity functions proposed by Brutsaert \cite{brutsaert1982} the system of Eqs. (4-8) relates surface fluxes to surface variables and the mixed-layer atmospheric variables. The criterion proposed by Brutsaert \cite{brutsaert1982} is used to determine if MOS or BAB scaling is appropriate for a given situation.

2.2 An extended model for determination of the roughness length for heat transfer

In the above derivation\cite{brutsaert1982} the aerodynamic and thermal dynamic roughness parameters need to be known. When near surface wind speed and vegetation parameters are available the within-canopy turbulence model proposed by Massman \cite{massman1999} can be used to estimate the aerodynamic parameters $H$, the displacement height and $z_0$, the roughness height for momentum. This model has been shown by Smith et al. \cite{smith1999} to provide reliable estimates of the aerodynamic parameters. If only the height of the vegetation is available the relationship proposed by Brutsaert \cite{brutsaert1982} may be used. If a detailed land use classification is available the tabulated values of $W$ (in m) \cite{smith1999} can be used. However, since aerodynamic parameters depend also on wind speed and direction as well as the surface characteristics \cite{smith1999} the latter two approaches should be used only when the first method cannot be used due to lack of data.

The scalar roughness height for heat transfer $z_{h0}$ which changes with surface characteristics and also changes with the aerodynamic roughness of the surface $z_0$ can be derived from the roughness model for heat transfer proposed by Smith et al. \cite{smith1999}. However, their model requires a functional form to describe the vertical structure of the vegetation canopy to calculate the within canopy wind speed profile extinction coefficient $n_u$. In this study $n_u$ is formulated as a function of the cumulative leaf area index at the canopy top

\begin{equation}
\frac{n_u}{|2u^*|/h^3|z_0|} = 1.7 \quad W \text{here } C_{z_0} \text{ is the drag coefficient of the foliage elements assumed to take the value of 0.2, LAI is the one-sided leaf area index defined for the total area.}
\end{equation}

$q_h$ is the horizontal wind speed at the canopy top. The scalar roughness height for heat transfer $z_{h0}$ can be derived from

\begin{equation}
\frac{n_u}{|2u^*|/h^3|z_0|} = \frac{k_B}{\epsilon} \quad W \text{here } B^{\epsilon-1} \text{ is the inverse Stanton number, } \epsilon \text{ a dimensionless heat transfer coefficient. The model proposed by Su et al. \cite{su1994} and Su \cite{su1994} is used to estimate } k_B^{\epsilon-1}\end{equation}

\begin{equation}
k_B^{\epsilon-1} = \frac{k_C}{4C_{z_0}} \epsilon \left( 1 - \frac{n_u}{h_0} \frac{z_0}{h_0} \right) 
\end{equation}

\begin{equation}
2 \epsilon \frac{n_u}{h_0} = \epsilon \left( \frac{z_0}{h_0} \right) C_{z_0}
\end{equation}

\begin{equation}
W \text{ here } C_{z_0} \text{ is the fractional canopy coverage and } \epsilon \text{ is its complement. } C_{z_0} \text{ is the drag coefficient of the foliage elements assumed to take the value of 0.2. } C_i \text{ is the heat transfer coefficient of the leaf. For most canopies and environments } C_i \text{ is bounded as 0.058 } \leq C_i \leq 0.075 \epsilon W \text{ is number of sides of a leaf to participate in heat exchange.} \end{equation}

The horizontal wind speed at the canopy top. The heat transfer coefficient of the soil is given by $C_i = \frac{10^{-5}R_e^{1.0}P_r^{0.4}}{\epsilon}$ where $P_r$ is the Prandtl number \cite{massman1999} and the roughness Reynolds number $R_e = \frac{h_i u_i}{\nu}$ with $h_i$ the roughness height of the soil. The kinematic viscosity of the air is given by $\nu = 1.327 \times 10^{-5} P_i \frac{M_i}{T} \frac{h_i^{1.00}}{T}$ with $p_i$ and $T$ the ambient pressure and temperature and $P_i = 1013.25$ kPa and $T_i = 273.15$ K.

Physically and geometrically the first term of Eqn.\cite{brutsaert1982} follows the full canopy model of Choudhury and Monteith \cite{choudhury1988} the third term is that of Brutsaert \cite{brutsaert1982} for a bare soil surface. A quadratic weighting based on the fractional canopy coverage is used to accommodate any situation between the full vegetation and bare soil conditions. For bare soil surface $k_B^{\epsilon-1}$ is calculated according to Brutsaert \cite{brutsaert1982}

\begin{equation}
k_B^{\epsilon-1} = 2.46 \frac{2.46}{R} \left( 1.8 \frac{2.46}{R} \right) \left( 2.46 \frac{2.46}{R} \right)
\end{equation}

2.3 A new formulation for determination of evaporative fraction on the basis of energy balance in limiting cases

To determine the evaporative fraction \cite{brutsaert1982} to be defined below\cite{brutsaert1982} in case of energy balance considerations in limiting cases. Under the dry limit the latent heat or the evaporation becomes zero due to the lim-
the sensible heat flux is at its maximum value. From Eqn. 13, \[ \lambda E_{aw} = R \cdot G \cdot H_{aw} = 0 \text{per} \]

Under the wet limit where the evaporation takes place at potential maximum \( \lambda E_{aw} \) i.e. the evaporation is limited only by the energy available under the given surface and atmospheric conditions, the sensible heat flux takes its minimum value \( H_{aw} \) i.e.

\[ \lambda E_{aw} = R \cdot G \cdot H_{aw} = 0 \text{per} \]

The relative evaporation then can be evaluated as

\[ \Lambda = \frac{\lambda E}{\lambda E_{aw}} \]

Substitution of Eqns. 14 and 15 in Eqn. 17 and after some algebra

\[ \Lambda = \frac{H_{aw}}{\lambda E_{aw}} \]

The actual sensible heat flux \( H_{aw} \) defined by Eqn. 17 is constrained in the range set by the sensible heat flux at the wet limit \( H_{aw} \) and the sensible heat flux at the dry limit \( H_{aw} \). \( H_{aw} \) is given by Eqn. 17 and can be derived by combination of Eqns. 15 and 17. The combination equation similar to the Penman-Monteith combination equation \( \rho 14 \). Monteith \( \rho 14 \) showed that when the resistance (m s are grouped into the bulk internal or surface, or plant and external aerodynamic resistance) the combination equation can be written in the following form

\[ \lambda E = \Delta \cdot \left[ R \cdot G \cdot \frac{P_{v}}{T} \cdot \frac{\gamma}{\gamma + \Delta} \right] \]

\[ \Delta \] is the rate of change of saturation vapour pressure with temperature \( \rho \) is the psychometric constant \( \rho T \) and \( H_{aw} \) is the bulk surface internal resistance and \( H \) is the external or aerodynamic resistance. In the above equation it is assumed that the roughness length \( h \) for heat and vapour transfer are the same \( \rho 14 \). The Penman-Monteith equation is strictly valid only for vegetated canopy whereas the definition by means \( \rho 15 \) is also valid for soil surface with properly defined bulk internal resistance.

At the wet limit the internal resistance \( R_{w} = 0 \) by definition. Using this property in Eqns. 13 and changing the subscript correspondingly to select the wet limit condition, the sensible heat flux at the wet limit is obtained as

\[ H_{aw} = \frac{1}{k_{w}} \cdot \left[ \frac{\left( \Psi_{s} + \Psi_{v} \right)}{\left( \Psi_{w} \right)} \right] \]

The external resistance depends also on the Obukhov length \( L_{e} \) which in turn is a function of the friction velocity and sensible heat flux \( \rho 14 \). With the friction velocity and the Obukhov length determined by the numerical procedure described previously the external resistance can be determined from Eqns. 18 and 19

\[ \lambda E = \frac{1}{k_{u}} \cdot \left[ \frac{\left( \Psi_{s} + \Psi_{v} \right)}{\left( \Psi_{w} \right)} \right] \]

Similarly the external resistance at the wet limit can be derived as

\[ \lambda E = \frac{1}{k_{w}} \cdot \left[ \frac{\left( \Psi_{s} + \Psi_{v} \right)}{\left( \Psi_{w} \right)} \right] \]

The wet limit stability length can be determined as

\[ L_{w} = k_{g} \cdot 0.6 \cdot \frac{D^{1/3}}{R_{g} \cdot c_{p} / \lambda} \]

The evaporative fraction is finally given by

\[ \Lambda = \frac{\lambda E}{\lambda E_{aw}} \]

By inverting Eqn. 20, the actual latent heat flux \( \lambda E \) can be obtained.

Eqns. 1-21 constitute the formulation of SEBS \( \rho 14 \) its validation using four different data sets over the complex Tibetan plateau is the subject of the following sections also a brief description of the data used.

3 Data and materials

Humankind agriculture economics and the entire ecosystem of the Asian region seriously depend upon the monsoon climate and its predictability. More than \( \rho 14 \) of the Earth's population lives under the influence of this monsoon climate. As a part of Global Energy and Water Cycle Experiment (GEWEX Asian Monsoon Experiment (GAMES) \( \rho 15 \) was conducted to understand the role of the Asian monsoon in the global energy and water cycle and to improve the simulation.
and seasonal prediction of Asian monsoon patterns and regional water resources. To clarify the roles of the interactions between the land surface and the atmosphere over the Tibetan Plateau in the Asian monsoon system, two experiments using different scales were implemented in cooperation with Chinese TIBEX (Tibetan plateau Experiment of Atmospheric Sciences) as follows:

A Plateau-scale experiment (80°E-100°E, 27°37'N) using the north-east and south-west networks of one-dimensional observational stations. Systems for Fig. 1 Plateau-scale experiment include the special radiosonde Networks the AWS equipped with soil temperature-moisture measuring capability, the PBL tower and the precipitation sampling systems including those for isotope studies.

A Mesoscale experiment (91°E-92.5°E, 30°-33°N, the Nujiang basin) in the central plateau with two three-dimensional Doppler radars, mobile radio-controlled sonographs, the radiosonde network and the AWS network as shown in Fig. 2 Plateau-scale experiment. There are two spatial scales inherent to this basin. The catchment area of the overall basin is approximately 105 km²; this is the larger basin scale. A smaller basin about 103 km² is also embedded in the larger one. The characteristics of flow and ground vary over a wide range from continuous permafrost in the north to semi-arid conditions in the south. The distribution of land surface water is directly affected by the permafrost distribution.

The field data used in this thesis, including the surface temperature and other meteorological information, are collected during GAME/Tibet experiments, May to September 1998. The study area Fig. 3 Plateau-scale experiment in 1998 which is located between 90°E and 95°E and 29°N up to 36°N with a total area about 322 300 km². It includes two radiosonde stations Lhasa and Linzhi, three heat flux stations Anduo, Nampa and Nu Jiang, and three Automated Weather Stations AM - SD (E01010 and Toutoude) in this area.

3.1 Heat flux stations
3.1.1 Anduo-PBL
Anduo PBL Lat. 32.241°N Lon. 91.625°E and Elev. 4700 m within the central Tibetan plateau. The surface is essentially flat and open and partially covered by very short grasses in the monsoon season.

Systematic measurements were held at Anduo in GAME/Tibet intensive observing period from May to September 1998. The subsurface measurements comprised soil moisture at six depths (20 cm, 40 cm, 60 cm, 100 cm, 160 cm and 258 cm) measured by TDR system. Trime EKO-MT soil temperature at twelve depths (4 °C, 10 °C, 20 °C, 40 °C, 60 °C, 80 °C, 100 °C, 130 °C, 160 °C, 200 °C and 279 °C) by thermometers, Pt-100 thermocouples, soil heat flux G at 10 cm and 20 cm depths by heat plate EKO MF-81. The measurements in the atmospheric surface layer include downward shortwave radiation R_s and upward shortwave radiation R_u by EKO MS-601; downward longwave radiation L_s and upward longwave radiation L_u by Pirzey PIR and skin radiative temperature derived from R_s and R_u with emissivity e = 0.98 suggested by the observer; sensible heat flux Hs, latent heat flux LE, and momentum flux u° at 2.85 m level above the ground by a fast response system consisting of a 3-D sonic anemometer KaJo DA-304 and an infrared open-path hygrometer KaJo AN-300. In the post-processing of turbulence data, various corrections were made including cms wind and humidity effects on temperature, dynamic calibration of the sensors, and low frequency instability for humidity and Webb et al. (1981) correction for fluxes. Tanaka et al. (1983) indicated that the sensor of the infrared hygrometer did not work well during and several hours after precipitation. Wind speed, temperature T, by Pt-100 thermocouples and water vapour by electric capacitance R_s precipitation by tipping buckets.

3.1.2 NAFAM [N 83475] - North portable automated monsoon

The surface of NAFAM [N 83475] is at 31.926°N, Lon. 91.716°E and Elev. 5063 m, which is essentially flat and covered by snow and grass. Surface meteorological/hydrological observation is carried out during GAME/Tibet IOP for the understanding of surface-atmosphere direct interaction from May to September 1998. PAM III [FLUX-PAM] from NCAR National Center for Atmospheric Research [USA] was used. Surface eddy fluxes of mo-
ment; sensible heat and latent heat are measured in real-time as well as non-meteorological/hydrological parameters. Diurnal and seasonal variations of surface processes are evaluated. Systematic measurements include (1) soil temperature $T_s$ at six depths, 4 cm, 20 cm, 60 cm, 100 cm, 160 cm, and 260 cm measured by STR-10 REBSHT at the same depth; (2) soil moisture $q_s$ at the same depth; (3) soil heat flux $G$ at 1 cm depth below the ground by heatpower REBSHT. The measurements in the atmospheric surface layer include (4) downward shortwave radiation $R_s$ and upward shortwave radiation $R_u$, by kipp-schenk pyranometers CM-11 Radiation; downward longwave radiation $R_D$ and upward longwave radiation $R_U$ by Eppley FIR; and surface radiative temperature by Everest 400. 4G - Radiation thermometer E.5 sensible heat flux $H_s$ by GILL SAT-43A and corrected for water vapor flux; latent heat flux $E_s$ by GILL SAT-43A Bandpass THM Vaisala 50Y. (4) wind $U_p$ by propellervane Anemometer Model9101L and temperature $T_r$; relative humidity $R_h$; specific humidity $q$ by Vaisala 50Y. In the data processing, net radiation is found to be missing from 4 component measurements due to missing solar radiation; in the period 06/01-07/03 and 09/25-08/24, 06/01-08/30/August; 09/03-06/August; 06/07-09/August end of DOP. Net radiation is replaced by simple net radiation from REBSHT dataset during this period. They are roughly consistent during the DOP.

4 Results and discussions

Before validating of SEBS energy balance components, the investigation of ground measurements for the energy balance term is conducted.

4.1 Consistency of ground measurements

In this study, the dataset of two heat flux observation sites Anoudo and NFAM are used to validate the SEBS estimation. Data used are the ground measurements of the net radiation, sensible heat flux, latent heat flux, and soil heat flux at the different soil depths below the ground from June to August in 1998. The surface soil heat flux can be replaced by the measurements of the soil heat flux at 1 cm depth under ground at NFAM because the measurements close to the surface and it will be used to validate SEBS estimation directly. However, only measurements of the soil heat flux at 10 cm and 20 cm depth under ground are provided at Anoudo during GAME Tibet so it is needed to be noticed that lots of heat storage takes place between the land surface and measurement depth. The observation-derived value which is derived from observations at 20 cm depths under the ground is used to evaluate energy budget Eqn. [18]; it is calculated as follows.

$$ G_s = \frac{\lambda_s c_s^f (T_s - T_{atm})}{2} $$

The temperature gradient at the soil surface in Eq. (21) is derived with thermal diffusion equation [10].

$$ \frac{\partial^2 T}{\partial z^2} = \frac{\lambda_s}{\rho c_p} \frac{\partial T}{\partial t} $$

$$ \rho c_p \frac{\partial T}{\partial t} = 4.18 \times 10^5 \left( \frac{1}{0.75} + 0.65 \right) \frac{1}{0.65} \left( \frac{2.21}{0.92} \right) \frac{1}{0.21} $$

$$ \frac{\partial T}{\partial z} = \frac{\lambda_s}{\rho c_p} \frac{\partial T}{\partial t} $$

$\lambda_s$ is soil thermal conductivity, $c_p$ is soil heat capacity, $\rho$ is soil density, $T$ is temperature, and $T_{atm}$ is atmospheric temperature. $\alpha$ is approximated to the maximum soil water content and $q_{s}$ is volumetric water content which was observed in the fieldwork.

It is shown Fig. 3 that derivation of surface soil heat flux is greater than soil heat flux at 20 cm depth during daytime. In contrast, the surface soil heat flux is smaller than that of 20 cm depth below ground at night. The observation-derived value has a lag in comparison with those measured at 20 cm depth below ground although their phases are almost coincident. It indicates lots of heat can be evenly stored under the surface and it will take time for transportation of soil heat from surface to deep soil layers under ground. The lag of time is about two hours. Moreover, surface soil heat flux is greatly influenced by the surface temperature and time. The same results are obtained by the ground measurements in June and July at Anoudo in 1998 [see Fig. 4].

The lack of energy budget closure is particularly noticeable in the Tibetan plateau during Asian summer
monsoon based on the observations of the GAME-Tibet and the TIPEX projects 1998 Figs. 5-7. As showing in Fig.7 the average maximum residual energy $R_{\text{net}} = G_0 + H + \lambda E$ is approximately equal to 200 W/m$^2$ at NPAM in August 1998 and the average maximum residual energy has exceeded 200 W/m$^2$ at Anduo in August 1998. The same results are obtained by the ground measurements in June and July at Anduo and NPAM in 1998 Figs.6-7.

Yang et al. [23] indicated that the daily average incoming and outgoing energy fluxes at Naqu FX for 15 clear day from July and September and MS3478 of GAME-Tibet where the residual energy $R_{\text{net}} = G_0 + H + \lambda E$ often exceeded 1/3 of net radiation. Similar large residual was also found at other sites on the Tibetan plateau [24]. Ishikawa et al. [24] pointed out that the imbalance problem mainly occurred in the afternoon and the residual energy was greater than those reported.

Fig.4 The diurnal trend of both observation-derived soil heat flux and soil heat flux at 20 cm depth at the ground at Anduo in (a) June and (b) July 1998.

Fig.5 The diurnal trend of the energy residue at NPAM (a) and Anduo (b) in June of 1998. The solid curves are the average diurnal trend of the energy residue.

Fig.6 The diurnal trend of the energy residue at NPAM (a) and Anduo (b) in July of 1998. The solid curves are the average diurnal trends of the energy residue.
in the FIFE and the BOREAS. Miyazaki et al.\cite{26} suggested that the amount of residual energy in the plateau was larger in the monsoon period than in the pre-monsoon period. One reason of the energy imbalance can be explained by topographic heterogeneity\cite{26}, canopy energy storage or insufficient fetch. Considering the melting process of frozen soil water, Tanaka et al.\cite{22} pointed out that the soil heat flux might be underestimated. Wang et al.\cite{27} suggested that the measurement error of latent heat flux might greatly contribute to the energy closure problem at Naqu site of GAME/Tibet. The same result was obtained by Yang et al.\cite{25} at Anduo\cite{25} however, Kim et al.\cite{28} contested that net radiation flux and latent heat flux were least erroneous while soil heat flux and sensible heat flux were most erroneous for the same site. These opposite conclusions are resulted from different evaluation methods and incomplete observations. On the other hand some researchers wonder whether the imbalance is physically related to the heat transport by mesoscale convection\cite{1}, because the convective activities are so strong over the plateau in the summer season\cite{1}. For example, Kim et al.\cite{29} suggested considering the effect of some local and advection terms\cite{29} but quantitative estimation of these terms is still difficult based on observations.

In order to clarify this problem we analysed the behaviours of the evaporative fraction by inverting Eqsns.\ref{eq:25} and\ref{eq:26}:

\begin{align}
\text{ef}_1 &= \lambda_E \frac{R}{R} - \lambda_E \frac{G}{G} \\
\text{ef}_2 &= \lambda_E \frac{H}{H} + \lambda_E \frac{A}{A}
\end{align}

Fig. 7 shows the relationship between two evaporative fractions for using ground datasets at Anduo in August 1998. Obviously\cite{25} the evaporative fraction $\text{ef}_1$ is smaller than $\text{ef}_2$ because the measurement of net radiation is greater than sum of the soil heat flux and turbulent heat flux measurements. It also shows that the energy is not in equilibrium in the GAME/Tibet experiment areas. Same results can be obtained using the ground measurements of other months and other observation sites. In general\cite{25} the phenomena of the energy imbalance exist in GAME/Tibet experiment according to the analysis of the ground datasets\cite{25} although the explanations of energy in balance are debated. When validation is made using such measurement\cite{25} one should keep the imbalance fact of energy budget in mind.

\begin{align}
\text{ef}_1 &= \lambda_E \frac{R}{R} - \lambda_E \frac{G}{G} \\
\text{ef}_2 &= \lambda_E \frac{H}{H} + \lambda_E \frac{A}{A}
\end{align}

4.2 Validation of SEBS components

Due to lack of measurement terms\cite{25} some parameters in SEBS are estimated by actual land surface and
vegetation distribution in GAME /Tibet experiment area in this case. For example, canopy total leaf area index LAI and the fractional foliage coverage $f_c$ are assumed to take the value of 0.15. The surface emissivity for vegetation $e_v$ and bare soil $e_s$ are 0.965 and 0.95 [7], respectively. The surface albedo is obtained by ratio of upward solar radiation and downward solar radiation.

<table>
<thead>
<tr>
<th>Table 1</th>
<th>Statistics of SEBS estimated versus observed heat flux of dataset at Anduo in June of 1998 [8] MAD $\Delta H$ Mean Absolute Deviation $\text{RMSE}$ Root $H$ and Squared Error $e_s$</th>
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<td>$G$</td>
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<td>$H$</td>
<td>$G$</td>
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<th>Statistics of SEBS estimated versus observed heat flux of dataset at Anduo in August of 1998 [8] MAD $\Delta H$ Mean Absolute Deviation $\text{RMSE}$ Root $H$ and Squared Error $e_s$</th>
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<td>29.71</td>
<td>52.60</td>
</tr>
<tr>
<td>Number of data used: 382</td>
<td></td>
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</tbody>
</table>

The four energy balance terms predicted versus the measured values and their diurnal behaviors are also shown in Figs. 9 [8]. Figs. 10 [8], 11 [8], 12 [8], Plate 11 [8]. Table 1 to 3 show the predicted versus observed heat fluxes at Anduo in June, July, and August in 1998. Both MAD $\Delta H$ Mean Absolute Deviation and $\text{RMSE}$ Root Mean Squared Error of the predicted net radiation and sensible heat fluxes are generally less than 30 W/m². However, bigger biases exist between estimations of SEBS and measurements for surface soil heat flux and latent heat flux with the estimated SEBS larger than measurements. Similar results are obtained by the ground measurements in June, July and August at Anduo and NPAM in 1998.

At the Anduo site in August 1998 surface soil heat flux of SEBS estimated is larger than those of observation-derived during daytime. In contrast, it is smaller than that of observation-derived at night. The maximum estimation $229 W/m²$ of SEBS is also greater than that of observation-derived $90 W/m²$. Meanwhile, the phase of observation-derived has a lag about two hours in comparison with those of SEBS estimation. Similar conclusion can be obtained for the two observation sites in June and July in 1998. It indicates that surface soil heat flux is obviously related with the surface temperature and the time of downward heat transportation. Hence, the surface temperature is an important factor for parameterization of surface soil heat flux. However, some part of this is derived from homogeneous land surface in the plain regions and the surface temperature is negligible. It cannot be suitable to estimate surface heat flux in the GAME/Tibet experiment areas.

Estimations of latent heat flux are much higher than the measurements but their phases coincided very well. It shows that the diurnal trends of latent heat flux of SEBS estimates are comparable well with those measured. The basis of SEBS is that the energy components must be in balance. However, as mentioned above, the residual energy $R_s = G_t - H - L_e$ often exceeded $1/3$ of net radiation in GAME/Tibet experiment areas. To investigate the deviation of latent heat flux, the corrected measurements can be calculated by $R_s = G_t - L_e$ according to the assumption of the energy balance [8]. The results show that the corrected measurements compare very well with SEBS estimates in Fig. 21. It also certifies that the methodology to derive latent heat flux in SEBS is reliable. Hence, the main bias between SEBS estimation and those measured is generated by the surface energy imbalance of ground measurements rather than the method of SEBS estimation. We concluded therefore that the latent heat fluxes were under observed at these studied sites and are the main cause of the energy imbalance. The preliminary results of this study were obtained in Zhang [42] and the findings are also consistent with the results obtained for the Anduo site alone by Yang et al [43].
Fig. 9  SEBS estimated versus measured surface energy balance terms at Anduo in June of 1998

Fig. 11  SEBS estimated versus measured surface energy balance terms at Anduo in July of 1998
According to the assumption of the energy balance

\textbf{5 Conclusions}

The surface soil heat flux is larger than soil heat flux at deeper layer in the soil, so a lot of heat storage takes place between the land surface and the measurement depth in the ground and is non-negligible in surface energy balance calculation using observation taken under the surface. A lag of the average diurnal trend between the surface and deeper layer in the soil is caused by downward heat transportation.

The current parametrization of the surface soil heat flux is derived from homogeneous land surface in the plain areas and the surface temperature is neglected in Eqn. \eqref{eq:3}. It is not valid to estimate the surface soil heat flux directly in GAME/Tibet experiment area. Hence for the surface heat flux a new parameterisation is needed in which the surface temperature should be considered.

The phenomena of the energy imbalance exist in GAME/Tibet experiment data although the explanations of the energy imbalance have been debated and not resolved yet. The residual energy \( R = R_\text{net} - (G + H + \lambda E) \) often exceeded \( 1/3 \) of net radiation, the average maximum residual energy is approximately 200 W/m\(^2\) in August 1998.

Based on the calculated statistic, the SEBS predicted net radiation and sensible heat flux compare very well with those of ground measured ones.

Estimated latent heat fluxes are much higher than the measurements. However, the corrected measurement which are calculated according to the hypothesis of the energy balance compare with the estimation of SEBS very well. It clarifies that the deviation is caused by the energy imbalance of ground measurements in GAME/Tibet experiment area rather than the estimation of SEBS. Since other terms estimated with SEBS compare very well with observed ones, the latent heat fluxes were likely under-observed.

\textbf{References}


青藏高原地区能量水分循环:地表能量平衡和湍流热通量

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摘要: 文章给出了青藏高原能量水分循环研究的概况和总结，着重估计了能量平衡各分项和湍流热通量等。在能量平衡的计算基础上，尽管能量不平衡的原因解释仍有争论并且没有解决，但我们揭示了 GAME/Tibet试验观测资料中能量不平衡现象。我们发现估算的潜热通量比实际观测的要高许多。然而，根据能量平衡假设的计算结果和 SBE5 的估算一致性很好。在此基础上可以归纳出差异主要由 GAME/Tibet 试验观测资料中能量不平衡引起，潜热通量的实际观测可能偏小。

关键词: 能量水分循环；青藏高原地区；能量平衡；通流通量