THE PLATEAU SCALE LAND-AIR INTERACTION AND ITS CONNECTIONS TO TROPOSPHERE AND LOWER STRATOSPHERE

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THE PLATEAU LAND-AIR INTERACTION AND ITS CONNECTIONS TO TROPOSPHERE AND LOWER STRATOSPHERE

DISSENYATION

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This thesis is approved by
Prof.dr. Zhongbo Su, promotor
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These are things we know that we know. There are known unknowns. That is to say, there are things that we know we don’t know. But there are also unknown unknowns. These are things we don’t know we don’t know.

Donald Rumsfeld
February 12, 2002

“知之为知之，不知为不知，是知也。”

《论语》
Acknowledgements

The motivation for this thesis began in 2007 when I was helping Prof. Bob Su dig hole and manage to bury soil moisture sensors safely in Nagqu region at 4700 m above sea level. We had discussions about the parameterizations in land surface model during the field work. I still remember several questions raised by me at that moment when he was heavily suffering headache from high elevation oxygen shortage, ‘What is heat roughness length and momentum roughness length’?, ‘Is soil moisture really important’?......like a children asking what’s this what’s that, which he/she feels interesting or not interesting. Then during my third year of merged master-PhD study in 2009 at Institute of Tibetan Plateau Research, Chinese Academy of Sciences, I came to ITC and start my second PhD under the CAS-KNAW joint PhD training programme. I must be crazy to start the second PhD life without realizing I haven’t finished the first one! During my first and second year, I am a little bit nervous when I saw several my best mates have finished their study and work comfortably. The pressure makes me to discuss my work with Bob very frequently, even I took his time on the train when everybody was enjoying the excursion trip. I am grateful to Prof. Su for his valuable advice and deep knowledge. He help me crank out quickly from ‘huge huge troubles’ in my first year and make me learn to swim by myself in the last year. I thank him for introducing me to the broad scientific garden and teach me how to write a scientific paper. He has left his ‘ZS’ marks in my manuscript of 1.1, 2.1, 3.1.... at late night or during his holiday time. He always answers my request quickly and be trustful and helpful. When my head was hammered by my paper’s reviewers, he encouraged me to calm down to fight back. I would also like to thank my Chinese side supervisor Prof. Yaoming Ma. Without your support, I would not have the chance to study in the Netherlands. Without your courage, the thesis can’t be finished so quickly. You are always generous to mistakes I made. I devote the thesis to both of your great and successful supervisions.

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Enschede, The Netherlands
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ABBREVIATION

ABL — Atmospheric Boundary Layer
ASL — Above Sea Level
BAST — Bulk ABL Similarity Theory
CAMP/Tibet — CEOP Asia Monsoon Tibet Experiment
CBL — Convective Boundary Layer
CERES — Clouds and Earth’s Radiant Energy System
EF — Evaporation Fraction
ET — Evapotranspiration
ERA — European Centre for Medium-range Weather Forecasts Reanalysis data
GAME/Tibet — Global water and energy measurement experiment Asia monsoon Experiment/Tibet
GLDAS — Global Land Data Assimilation System
G0 — Ground heat flux
H — Sensible heat flux
HC — Canopy Height
IOP — Intensive Observation Periods
ISCCP — International Satellite Cloud Climatology Project
ITP — Institute of Tibetan Plateau Research, Chinese Academy of Sciences
JICA — Japan International Cooperation Agency
JRA — Japanese Reanalysis
LT — Local standard Time
LE — Latent heat flux
LRT — Lapse Rate Tropopause
LST — Land Surface Temperature
LWD — Downward Longwave radiation
LWU — Upward Longwave radiation
MB — Mean Bias
MT — Multi-Tropopause
MOST — Monin-Obukhov Similarity Theory
PBL — Planetary Boundary Layer
Q — Relative humidity
RMSE — Root Mean Square Error
Rn — Net radiation
SEBS — Surface Energy Balance System
STE — Stratosphere and Troposphere Exchange
SWD — Downward Shortwave radiation
SWU — Upward Shortwave radiation
Ta — Air temperature
TESEBS — Topographical Enhanced Surface Energy Balance System
TORP — Tibetan Observation and Research Platform
Tk2 — Eddy-covariance software developed by Bayreuth University
UTLS — Upper Troposphere and Lower Stratosphere
WS — Wind speed
As the largest and highest plateau in the world, the Tibetan Plateau exerts profound thermal and dynamic influences on the Asian monsoon and the northern hemispheric atmospheric circulation through land-atmosphere interactions (Hsu and Liu, 2003; Yanai et al., 1992). The plateau land surface processes not only affect the development of the local atmospheric boundary layer (ABL) but also change the water vapor distribution and temperature gradient in the air, which finally exert impacts on the weather and climate of East Asia (Ma et al., 2008b).

The interaction of water and energy between the land and the atmosphere is realized through the turbulent exchange of momentum, heat, and moisture in the atmospheric surface layer. A better characterization of surface energy and water exchange can lead to significant improvements in simulations and predictions of weather and climate (Balsamo et al., 2011). The complexity of land surfaces (lake, mountain, valley, wetland etc.) and vegetation types (meadow, forest etc.) on the Tibetan Plateau makes the parameterization of the land air interaction difficult. Conventional meteorological techniques that employ point measurements to estimate the components of the surface energy balance represent phenomena on a local scale (Su et al., 2006). Remote sensing enables us to estimate surface fluxes from local to plateau scales and offers the possibility of deriving the regional distribution of land surface heat fluxes over poorly instrumented and sparsely populated areas (Ma et al., 2006; Ma et al., 2011). Recent studies have explored remote sensing approaches to estimate the regional distribution of surface heat fluxes (Ma et al., 2008c; Ma et al., 2003; Oku et al., 2007). This PhD study uses ground measurement, the space remote sensing technique, a surface energy balance model, and atmospheric sounding to analyze the interactions between the plateau land surface and several important atmospheric layers (e.g. surface layer, ABL, upper troposphere and lower stratosphere (UTLS)). As the height of the plateau surface is close to the middle troposphere, the connections between the process in the ABL and the UTLS can be more frequent than that in other places (Skerlak et al., 2013).

1.1 Scientific background

It has been pointed out that the stratosphere-to-troposphere ozone flux into the ABL is the greatest on Earth over the Tibetan Plateau and the Rocky Mountains.
Introduction

(Skerlak et al., 2012; 2013). Chen et al. (2012) concluded that the Tibetan Plateau is the third of the three key source regions for transport from the boundary layer to the tropopause in the Asian monsoon region, after the South China Seas and the South Asian subcontinent. Both the ABL-to-tropopause and the stratosphere-to-ABL exchanges above the plateau are closely related to the structure and characteristics of the ABL and the UTLS. The Tibetan Plateau is a region of maximum multiple tropopause events and tropopause foldings (Randel et al., 2007; Añel et al., 2008; Chen et al., 2011), which can cause stratospheric air to be transported downward to the ABL via different processes (Johnson and Viezee, 1981).

Exposure to surface ozone can cause serious health problems in plants and people. Tropospheric ozone is known to have a damaging effect on carbon uptake in the terrestrial biosphere (Kvalevåg and Myhre, 2013). Stratospheric ozone can be brought into the troposphere during stratosphere-troposphere exchange (STE) events (Junge, 1962). Roelofs and Lelieveld (1997) pointed out that cross-tropopause transport of ozone is a significant factor in the tropospheric budget and distribution of ozone. In the subtropics, stratospheric ozone is transported downward through the tropopause breaks. Large-scale subsidence at these latitudes further transports ozone down to the surface (Roelofs and Lelieveld, 1997).

Ozone transported from the stratosphere and net photochemical formation in the troposphere are of comparable magnitudes (Roelofs and Lelieveld, 1997). Ozone from the stratosphere contributes significantly to surface ozone in winter and spring, when the photochemical lifetime of ozone is relatively long (Roelofs and Lelieveld, 1997). The cross-tropopause transport of ozone follows a seasonal cycle, with the maximum in spring in the northern hemisphere (Roelofs and Lelieveld, 1997). In winter and spring, downward cross-tropopause transports of ozone are relatively large (Roelofs and Lelieveld, 1997). Skerlak et al. (2013) pointed out that the Tibetan Plateau is a hotspot for stratospheric air reaching the PBL, especially in boreal winter and spring. The surface ozone concentrations around the Tibetan Plateau are likely to be influenced by deep stratospheric intrusions. Deep STE down to the surface can also contribute to enhanced ozone levels at the ground and affect plant and human physiology.
The structure variation of the UTLS will influence the ozone column over the plateau (Chen et al., 2011a). What is the influence of the thermal dynamics of the Tibetan Plateau on the stratospheric intrusion? The influence of the thermal dynamics of the plateau on the atmosphere will be analyzed using a newly developed high-resolution land surface heat flux product. Analyzing the relationship between the variation in surface heating and the STE process in the UTLS area will also be a part of this work.

1.2 Problem statement

Due to a lack of high-resolution radio-sonde data, the structure of the UTLS above the Tibetan Plateau is still insufficiently understood (Gettelman et al., 2010; Hegglin et al., 2010). Many previous studies have assumed that a simple tropopause exists over the Tibetan Plateau. Meanwhile, satellite observations of ozone have shown ‘ozone mini-hole’ events and ozone valley phenomena over the plateau (Zhou et al., 2005; Tobo et al., 2008; Bian, 2009). The dynamic mechanisms responsible for the low total ozone have not been disclosed.

PBL processes control energy, water, and pollutant exchanges between the surface and free atmosphere (Seidel et al., 2010). The ABL over the plateau is deeper than that over the lowlands (Fan et al., 2011; Li et al., 2006; Yang et al., 2004; Zuo et al., 2005). However, few studies have investigated the connection between the high ABL and the UTLS, even though the Tibetan Plateau is regarded to be a pathway of mass exchange between the troposphere and stratosphere (Zhou et al., 2006). The elevation of the Tibetan Plateau varies between 3000 and 8848 m above sea level (ASL). The top of the ABL may be as high as 9000 m ASL, which is close to the location of the tropopause. This is due to both the elevation of the plateau and the depth of the ABL. This may result in a stronger interaction between the UTLS and the ABL than in lowland areas (Skerlak et al., 2013).

ABL depth is higher in late winter daytime than during monsoon-onset and monsoon daytime over the Tibetan Plateau. Why is the ABL height higher in late winter time than during monsoon-onset and monsoon time over the plateau? Is it caused by changes in plateau land surface heating or other heating sources? These scientific questions will be addressed.
1.3 Statement of the objectives

This PhD study is designed to:

a) Investigates surface-flux parameterization and determines local water and energy flux over different land types of the Tibetan Plateau (e.g. alpine meadow, forest, lake, mountain, valley) based on measurements at the flux station and sites;

b) Determines land surface type and vegetation distribution by using satellite remote sensing; to adopt the appropriate turbulent flux parameterization method in surface energy balance system (SEBS) (Su, 2002) referring to the specific land surface; and to assess the performance of SEBS, which will firstly be driven and evaluated by the surface observation data.

c) Improves the SEBS model and use it to upscale energy and water exchange fluxes or parameters, integrating remote sensing data with surface meteorological data for the whole Tibetan Plateau area. Evaluates errors in which variable’s estimation will be the most significant factors influencing the model outputs.

d) Connects the land surface energy and water exchange with the process in ABL and upper air; to analyze the role of land-air interactions play on the characteristic layers, such as the ABL and the UTLS.

e) Uses the long time series of surface energy exchange data to quantitatively determine heat source/sink and water vapor source/sink; and to appraise climate change involving the variation in surface energy balance.

1.4 Methodology and observational data

Many previous conclusions about the Plateau land air interactions have been based on point measurements, or a meso-scale network. It is difficult to extend stations’ results to larger area due to limited coverage of measurement footprint. Remote sensing is probably the only technique that can provide observational information for land surface from point to continental or global scale (Su and Jacobs, 2001). Here we have developed a land surface energy balance model (after point evaluations and parameterization improvement) driven by remote sensing data and applied the model to provide land-air energy and water exchange information to study the role of Tibetan Plateau plays on the ABL and its connections with UTLS.
1.4.1 Land surface energy balance measurement

High-altitude observations are important for model physics evaluation and improvement (Bollasina and Benedict, 2004). Several experimental campaigns (e.g. QZPMEX (Zhang et al., 1988; Ye and Gao, 1979), GAME/Tibet (Koike et al., 1999), CAMP/Tibet (Ma et al. 2006), and JICA/Tibet (Zhang et al., 2012b)) have been carried out during the last thirty years. TORP (Ma et al., 2008b), a long-term surface observation network was set up in 2005. It is the biggest and most comprehensive surface energy balance observation network on the plateau. Datasets from other flux stations are also collected. Each site of the network represents a certain land surface type. More details will be given in Chapter 2 and 3. The dataset of each station consists of surface radiation budget components (downward/upward longwave/shortwave radiation), air temperature, humidity, wind speed and direction, turbulent fluxes measured by the eddy covariance technique, soil temperature, soil moisture, soil heat flux etc. In each case the new dataset was used in this PhD study.

1.4.2 Land surface energy balance model

In order to obtain surface energy balance map information at the plateau scale, it is essential to use remote sensing information to estimate land surface heat and energy fluxes over the plateau where ground observation data are highly scarce. The methodologies used in this study are described in Fig. 1.1. The surface energy balance system (SEBS) (Su, 2002) requires three sets of information as inputs. The first dataset consists of land surface albedo, emissivity, temperature, vegetation coverage leaf area index, and the height of vegetation from satellite data (e.g. MEdium Resolution Imaging Spectrometer (MERIS), Moderate Resolution Imaging Spectroradiometer (MODIS) and Advanced Along-Track Scanning Radiometer (AATSR)). When vegetation information is not explicitly available, NDVI is used as a surrogate. The second dataset includes air pressure, temperature, humidity, and wind speed at a reference height. This dataset can also be obtained from reanalysis data, e.g. National Centers for Environmental Prediction (NCEP) (Kalnay et al., 1996), European Centre for Medium-Range Weather Forecasts (ECMWF) interim re-analysis (ERA-Interim) (Dee et al., 2011), Modern-Era Retrospective Analysis for Research and Applications (MERRA) (Rienecker et al., 2011). The third dataset includes downward short-wave radiation and downward long-wave radiation, which can be provided by present radiation product, e.g. International Satellite Cloud Climatology Project.
Introduction

(ISCCP), Clouds and Earth’s Radiant Energy System (CERES), Global Land Data Assimilation System (GLDAS).

Fig.1.1 Methodology and technique used in land surface energy analysis in the study

The model evaluation and parameterization improvement was realized in two ways:

a) The model was driven by station point measurement at locations in different climate environments. The point flux simulation output was compared with the flux measurement. The sensitivity of the output flux to input variables and parameters was analyzed with bias between the measured ‘true values’ and modeled values. This technique overcomes the influence of uncertainties in the forcing data on the model evaluations.

b) The usage of the advanced SEBS model was up-scaled to regional and continental areas. The forcing data for continental studies comes from remote sensing and meteorological reanalysis data. Both inputs and outputs were evaluated based on the station observations. This method can identify the error sources in the present calculation of land surface fluxes and evapotranspiration when using the energy balance model at large scale.
It is easy to obtain trustable values for land surface temperature, normalized difference vegetation index (NDVI), and canopy information at one point. When the applications of the input variables are expanded to continental or global scale, errors in their values are all incorporated into the model computations. By evaluating the influence of errors in input variable estimations on the outputs can help us to choose more reliable input datasets for this study area, and it can also suggest the enhancement of specific variable precision when using remote sensing to quantitatively study the water and energy exchange over the Tibetan Plateau.

1.4.3 Observations of atmospheric boundary layer depth

In the last two decades several comprehensive field experiments have been carried out to analyse the land surface processes and the structures of the ABL over the Tibetan Plateau (Li et al., 2006; Ma et al., 2009; Sun et al., 2007; Xu et al., 2002; Yanai et al., 1992; Yanai and Li, 1994; Yang et al., 2004). It was discovered that the depth of the ABL over the plateau is larger than it is over the lowlands (Fan et al., 2011; Li et al., 2006; Yang et al., 2004; Zuo et al., 2005). Because of the harsh natural conditions, there were few intensive radiosonde observations over the Tibetan Plateau before 2008 – the year when a regional radiosonde observation network was implemented through a Sino-Japan joint cooperation project (Xu et al., 2008). In our research we analyzed the seasonal variations in the ABL over the plateau and related them to the surface energy balance and process in the UTLS.

1.4.4 Observations of upper troposphere and lower stratosphere

So far the lack of high-resolution data in the atmosphere has hampered our knowledge of tropopause characteristics, limiting our ability to evaluate model performance in the UTLS (Hegglin et al., 2010). Many previous studies have assumed that there was only one thermal tropopause over the Tibetan Plateau (Fan et al., 2008; Tian et al., 2008). Actually observations in this thesis indicate that a multiple tropopause (MT) often occurs over the plateau in winter. Several references have analyzed global MT events with coarse vertical resolution data based on GPS radio occultation (Son et al., 2011), ERA-40 data (Randel et al., 2007a) and the Integrated Global Radiosonde Archive database (IGRA) (Añel et al., 2008). The high-resolution radiosonde data of 2008 are an important source of information for studying the thermal structure of the UTLS. This thesis also
reports on an analysis of the structure of UTLS and provides observations of stratosphere and troposphere exchange (STE) over the plateau. We also made a comparison to MT events derived from other low quality dataset.

1.5 Structure of the Thesis

Chapter 1 (this chapter) gives an introduction to the scientific question, the methodology, and the specific objectives. An analysis of observed annual and seasonal variations in land surface energy and water fluxes is presented in Chapter 2. Chapter 3 describes our improvement and development of the land surface energy balance model. This task is carried out by making comparisons between model output and flux measurements. While most of the studies using SEBS to derive surface energy balance items relate to flat areas (Su et al., 2005; Yang et al., 2010a), none of them considers topographic influence. When applying SEBS to the high-resolution satellite dataset, the topographic influences become increasingly important. Chapter 4 introduces a topography-based land surface radiation model into the surface energy balance model and demonstrates the applicability of the surface energy balance model to complex mountain areas. Chapter 5 upscales the application of the land surface energy balance model to a continental area to analyze the spatial distribution of each land surface energy balance item. Chapter 6 analyzes the diurnal and seasonal variations in the top of the convective boundary layer over the Tibetan Plateau, check the relationship between land surface energy partitioning and ABL height, and the potential of high ABL’s connection with the upper air process. The process in UTLS is studied in Chapter 7. Special attention is given to the tropopause folds over the Tibetan Plateau. The possibility of coupling between the high ABL and tropopause is discussed in Chapter 8. The consistent variations in the ABL and tropopause folds are also analyzed. The final chapter summarizes this PhD study and gives an outlook for the future.
Chapter 2 Analysis of land-atmosphere interactions based on ground observations

2.1 Abstract

To better understand the basic characteristics of the land surface energy budget, yearly continuous measurements of land surface energy fluxes have been analyzed. Seasonal and annual variations of micrometeorological measurements and land surface energy balance were analyzed. The results disclosed that the high soil moisture in summer is corresponded with low albedo. The ratio between sensible heat and net radiation (H/Rn) can be as high as 0.49 when the soil is driest. The ratio (H/Rn) decreases to 0.14 with the soil moisture increased to the highest value. On the contrary, the ratio between latent heat flux and net radiation (LE/Rn) is increased when soil moisture rising. The highest value of LE/Rn can be as high as 0.5 when soil moisture was between 15% and 20%. After defining the effects of different soil moisture level on partitioning of surface available energy into sensible and latent heat fluxes, it does benefit to qualify how much the sensible heating is decreasing and the latent heating is increasing under current plateau environment changes of warming and moistening.

* This chapter is based on:
2.2 Introduction

Due to the elevation of the Tibetan Plateau, energy and water are exchanged between the surface and the middle troposphere above it. The growing interest in the interaction between atmosphere and plateau surface is driven by the realization that the thermal effects of the plateau may influence atmospheric circulation in the northern hemisphere (Tao and Ding, 1981; Wang et al., 2011). The lack of observational knowledge on the interactions between the plateau and the atmosphere limited our understanding of the land-air interaction in this region (Chen et al., 2012c). Here we used a new ground in-situ network to analyze the land-air interaction signal.

A large area of the permafrost and seasonal frozen soil are widely distributed on the 2.5 million km² area of the plateau (Yang et al., 2010b). The Tibet Plateau is one of the most sensitive areas to global change (Liu and Chen, 2000). The Tibet Plateau present trends towards warm and wet conditions (Li et al., 2010; Zhong et al., 2011). The seasonal frozen soil and permafrost on the Tibet Plateau has already been impacted by global warming and changing now (Yang et al., 2010b; Zhao et al., 2004). The thawing process will lead to dramatic changes in the soil moisture, which in turn exert a profound influence on the entire energy and water exchange (Tanaka et al., 2003). Here we invested to formulate the relations between soil moisture and land surface energy departing based on measurements over a bare soil land surface.

2.3 Ground measurement and sites

Data used in this thesis were collected from several surface point stations on the Tibetan Plateau. It includes Qomolangma Station for Atmospheric and Environmental Observation and Research, the Chinese Academy of Sciences (hereafter QOMS/CAS), Nam Co for Multisphere Observation and Research (hereafter Namco), Linzhi Station for Alpine Environment Observation and Research (hereafter Linzhi station), Ngari Desert Observation and Research Station (hereafter Ali station) and Maqu station. The locations and detailed information about each site are listed in Table 2.1. The landscape around each site’s equipment is flat and covered with very sparse and short grass in the monsoon season.

The QOMS/CAS is located 30 km away from Mount Everest. The dataset of the QOMS/CAS station consists of surface radiation components, air temperature,
humidity, wind speed and direction, and turbulent fluxes. Sensors for wind speed, wind direction, air temperature, and relative humidity are installed at five different levels (1.0, 2.0, 4.0, 10.0, and 20.0 m) of a 40 m PBL tower. An open path eddy covariance turbulent measurement system is set up at 3.5 m height and runs continuously at a sampling frequency of 10 Hz. The surface of the site is formed by sandy soil with small rocks. Sparse vegetation up to 0.2 m in height is found scattered around the station in summer time.

<table>
<thead>
<tr>
<th>TABLE 2.1 Eddy covariance site information and parameters used in TK2</th>
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<tr>
<td>Latitude [degree]</td>
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<td>Longitude [degree]</td>
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<tr>
<td>Elevation [m]</td>
</tr>
<tr>
<td>Land cover</td>
</tr>
<tr>
<td>Eddy covariance measurement height [m]</td>
</tr>
<tr>
<td>Direction of Licor7500 [degree]</td>
</tr>
<tr>
<td>Horizontal distance (CSAT3-Licor7500) [m]</td>
</tr>
<tr>
<td>Direction of CSAT3 [degree]</td>
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<tr>
<td>Sample frequency</td>
</tr>
</tbody>
</table>

The surface at Namco station is relatively smooth and covered with very sparse and short grass in the monsoon season. The meteorological measurements (sensor height, depth, and types) taken at Namco and QOMS/CAS station are the same (http://www.namco.itpcas.ac.cn/namcoenglish.html). During winter, the station surface is continuously covered by a homogeneous, flat, and open snow field.
Linzhi Station is located in the southeast of the Tibetan Plateau. The site’s surface is mainly covered with a dense canopy, and the site is surrounded by forest. This canopy is the highest and densest of the four sites. The meteorological sensors are the same as those at QOMS/CAS and Namco station.

Maqu site is located on the northeast side of the Tibetan Plateau. The Maqu site has a relatively smooth surface and is covered with short grass in the monsoon season. Sensors of wind speed, wind direction, air temperature, and relative humidity are installed at five different levels (1.4, 3.2, 4.4, 8.4, and 18.4 m) on a PBL tower. Four components of radiation are installed 1.5 m above ground. The information for the flux tower measurements at each station was given in Table 2.1.

### 2.4 Data processing

Before using turbulent fluxes of the eddy covariance system, a data quality assessment and controlling process was applied to the turbulent data. The high frequency turbulent data were processed to produce half-hourly fluxes. Calibrations of the eddy covariance signals were also executed. Meteorological and soil observation data were sampled every 10 minutes. All datasets were then resampled to 30 minute intervals by nearest interpolation method.

The application of the eddy-covariance technique is based on a few theoretical assumptions, such as horizontal homogeneity, steady-state, and non-advective conditions (Gockede et al., 2004). Therefore, to gain high quality flux data, quality assessment and quality control have to be established in the processing of the eddy-covariance data. TK2 (Mauder and Foken 2004) was used here to process the turbulent data of QOMS/CAS, Namco and Linzhi. The spike detection algorithm of (Vickers and Mahrt, 1997) was used to remove physically or electronically unfeasible values in the high frequency turbulent data. The time delay between sonic anemometer (CSAT3) and gas analyzer (Licor-7500) signals was eliminated by maximizing covariance. Tilt-correction was done with the planar fit method (Wilczak et al., 2001). Correction of spectral loss was done according to Moore (1986)’s method. The sonic temperature was converted into actual temperature. WPL-corrections (Webb et al., 1980) were also performed in the process. The TK2 input parameters of each eddy covariance system are listed in Table 2.1.
2.5 Results

2.5.1 Seasonal and diurnal variation

Fig. 2.1 Diurnal and seasonal variations of temperature (a), relative humidity (b), wind direction (c), wind speed (d), and water vapor content (e) in the year of 2007 at QOMS/CAS station. The unit of x-axis is day.
A year of air temperature, humidity, wind speed, and wind direction in 2007 at QOMS/CAS was taken as an example to shown the seasonal and diurnal variations in meteorological signals (Figure 2.1). The x-axis is the date of year and the y-axis is the daytime. Air temperature showed a significant seasonal variation. The air temperatures are found to vary in the range of -20.4 to 26.1°C. The difference between the highest air temperature and lowest air temperature was 46.5°C. The variation of humidity near the surface indicates quite well the physical processes regulating the moisture content of the air at the ground. The diurnal variation in relative humidity (RH) is out of phase with the temperature variation. High water vapor content are observed at night time during the monsoon, which is associated with the night heavy rains during this season, as pointed out by Liu et al. (2002). Low values of water vapor content are observed at daytimes. The diurnal variability of water vapor content is greater during monsoon.

2.5.2 The impacts of soil moisture on albedo

The surface albedo is one of the most important parameters to reflect the land surface characteristics. Land surface albedo is calculated as the ratio of the total upward solar radiation (the reflected solar radiation) and the total downward solar radiation reaching the surface. Surface albedo is strongly influenced by solar elevation angle and surface conditions, such as soil moisture, vegetation cover, roughness, and so on. In order to eliminate the influence of vegetation, the measurements of QOMS/CAS were used because the station is surrounded by a flat bare soil. Here we chose observations from 1 Jun to 30 September without snow covers. It is necessary to first examine the influence of solar zenith angle on surface albedo (Guan et al., 2009; Wang et al., 2005). The solar zenith angle can be calculated from the longitude and latitude of the site, Julian day and mean measurement time. Our analysis shows that solar zenith angle has little influence on the surface albedo when the solar zenith angle varies below 30 degree (not included here). Therefore, when the solar zenith angle varies from 0 to 30 degree, the surface albedo were used to study its relations with top soil moisture. The soil moisture coordinates very well with every rain event (Fig. 2.2). Soil moisture in shallow layers is strongly affected by precipitation. Meantime, the observed albedo dropped down during each rain event.
Fig. 2.2. Variation of albedo (a) and soil moisture with precipitation (b) from 1 June to 30 September 2007 at QOMS/CAS.

Fig. 2.3 The scatter plot and fitting line between top soil moisture and albedo at QOMS/CAS.
The relationship between surface albedo and surface soil moisture content (volume per volume) in the top soil in Fig. 2.3 was presented by equation 2.1. One can see that albedo decreases with increasing soil moisture, with the data fitting the following equation:

\[
\text{Albedo} = -0.00027 \times \text{SM}^2 - 0.00113 \times \text{SM} + 0.24497
\]  

(2.1)

Soil moisture is the key parameter not only in controlling variation of albedo, but also influencing the ratio of net radiation partitioning into latent and sensible heat fluxes. In the next section, we will give analysis on this issue.

### 2.5.3 The impacts of soil moisture on the land-atmosphere energy exchange

Soil moisture is one of the most important factors affecting the partitioning of available energy into sensible and latent heat fluxes between the ground surface and the atmosphere. Soil moisture has been shown to play an important role for the occurrence of climate extremes (Hirschi et al., 2011; Jung et al., 2010). The role of soil moisture on the weather and climate change is accomplished by adjusting the land-air energy and water exchanges. In order to evaluate the influence of soil moisture variations on the surface energy flux, we divide top soil moisture (SM) into different level, level 1 (in red color in Fig. 2.4) 0-2.7%, level 2 (green color in Fig. 2.4) 2.7-6.1%, level 3 (blue color in Fig. 2.4) 6.1-9.5%, level 4 (yellow color in Fig. 2.4) 9.5-12.9%, and level 5 (pink color in Fig. 2.4) 12.9-16.3%. Based on the SM level, we plot both sensible heat flux and latent heat flux scatter points with net radiation (Fig. 2.4a, 2.4b). The turbulent heat fluxes from 10:00 to 16:00 were used here, when stronger heat and water coupling between the land-surface and the air occurs. Taking the points distribution into account, linear fitting lines were used. The difference in the intercepts of the five fitting lines is relatively small. The ratio of H/Rn on SM level 1 can be as high as 0.41, then it is gradually decreased to 0.13 at SM level 5. The scatter points of LE and Rn demonstrate that the slope between LE and Rn is rising, when the soil moisture changes from level 1 to level 5 and becomes wet. The ratio of LE and Rn is increased from 0.17 to 0.5 by rising soil moisture from level 1 to level 5. The Evaporative Fraction (EF) (Nichols and Cuenca, 1993), which characterizes the partition of the surface energy budget, is defined as the ratio between the latent heat flux and available energy at the surface.

\[
\text{EF} = \frac{\text{LE}}{\text{LE} + \text{H}}
\]  

(2.2)
LE scatter points with H+LE are drawn in Figure 2.4c. EF increases with rising of the soil moisture content. The intercepts of five linear lines in Figure 2.4c are small and have little difference. Thus EF of each soil moisture level is taken as the corresponding slope value. EF is increased from 0.22 on SM level 1 to 0.78 on level 5. The slopes of level 4 and 5 have little discrepancy. The high EF on level 4 and 5 suggest that the soil was saturated, and a large part of available energy at the surface was used for evaporation.
We pick out the slopes of the fitting lines in Figure 2.4 and the SM level, and plot them in Figure 2.5. It shows that the ratio of sensible heat flux and net radiation (H/Rn) decreases from 0.55 to a lowest value of 0.14 when soil moisture increased from 0 to 15%. On the contrary, the ratio of latent heat flux and net radiation (LE/Rn) increases from 0.10 to 0.40 when soil moisture increased from 0 to 15% (Fig. 2.5(b)). The scatter points in Figure 2.4(a) and 2.4(b) were fitted by the following exponential functions:

\[
\frac{H}{Rn} = 0.54998 - 0.11947 \times SM + 0.0109 \times SM^2 - 0.00032 \times SM^3 \\
\frac{LE}{Rn} = 0.09936 - 0.01931 \times SM + 0.00427 \times SM^2 - 0.0001 \times SM^3
\]

These two equations give us a reference estimation of the soil moisture variations’ effects on the surface energy. As scientists have disclosed the thawing process of frozen soil is accelerating (Yang et al., 2010b). The thawing will increase the soil moisture in short time. Researchers have discussed that the increasing soil moisture will decrease sensible heat flux and strength latent heat effects, but without dependable observational results. The results in equation 2.3 and 2.4 give us a quantitatively estimation of influence of top soil moisture on the land-atmosphere heat and water exchanges. The soil moisture induced variations in energy and water exchanges maybe cause alternations of the energy and water circle around the Plateau. When the soil is dry, the evaporative fraction is typically 0.2, demonstrating that only 20% of the energy transferred from the surface to the atmosphere is via latent heating (Fig. 2.5(c)). In contrast, the evaporative fraction is ~0.8 when the soil is wet. A more gradual increasing of EF sensitivity with a decrease in soil moisture was shown in the picture.
Fig. 2.5. (a) Scatter plot and spline fitting functions between the ratio of sensible heat flux (H) and net radiation (Rn) and soil moisture level, (b) Scatter plot and spline fitting functions between the ratio of latent heat flux (LE) and net radiation (Rn) and soil moisture level, (c) Scatter plot and spline fitting functions between evaporative fraction (LE/(H+LE)) and soil moisture.
2.6 Conclusion and discussions

Here we took the bare soil at QOMS/CAS as a representation of Tibetan Plateau. It is found that the variation of evaporative fraction over typical Tibetan bare soil as a function of soil moisture. A relationship between different soil moisture level and normalized surface turbulent energy fluxes was also formulated by this study.

Matthew (2010) estimates that the soil moisture anomalies over the TP increased at a rate of 0.03 to 0.13% per year. According to our equation between soil moisture and albedo over the Qomolangma region, a 0.08% anomaly of soil moisture will cause albedo a variation of 0.0008, in case of a 15% s soil moisture. The annual shortwave can be taken as 300 W/m². The net radiation will has an anomaly of 0.24 W/m². The value of H/Rn and LE/Rn at a 15% soil moisture is about 0.21 and 0.47 separately; Finally will cause sensible and latent heat flux anomalies of 0.05 and 0.11 W/m² separately. Using the equations of 2.1, 2.3 and 2.4, impacts of soil moisture anomalies $\delta$(SM) on the surface energy partitioning ($\delta$(LE) and $\delta$(H)) can be connected by the following differential analysis: $\delta$(SM)$\rightarrow$ $\delta$(albedo)$\rightarrow$ $\delta$(net radiation) $\rightarrow$ $\delta$(LE) and $\delta$(H). Even though the soil moisture-net radiation feedback proposed by Eltahir (1998) was not taken into account and the derived empirical formulas are limited to local scale, our result bridge the gap between increasing of soil moisture and the partitioning of energy fluxes at the land surface into sensible and latent heat flux, thus help to understand the water cycles changes in terms of soil moisture’s climatic effects. This chapter shows quantitative measurement of water and energy exchange at station point scale.
Chapter 3 An Improvement of the Surface Energy Balance System (SEBS) over the Tibetan Plateau*

3.1 Abstract

The technique of remote sensing enables us to estimate surface fluxes to bridge the gap between local scale flux measurements and regional scale land-atmosphere exchanges. Here we evaluated the performance of SEBS (surface energy balance system, which is popular used to remote sensing data to derive geographical distribution of surface energy balance items) over different land cover based on time series of measurements at four Tibetan sites which have land covers of bare soil, sparse canopy, dense canopy, and snow surface. Both under- and over-estimation of sensible heat fluxes by SEBS was discussed. Sensitivity analyses identified that biases are related to the bare soil’s $k B^{-1}$ parameterization in SEBS. The $k B^{-1}$ of bare soil in SEBS was updated from Brutsaert (1982) to Yang et al. (2002) and the evaluation results show that the revised model performs better than the original one.

* This chapter is based on:
3.2 Introduction

The Surface Energy Balance System (SEBS) developed by Su (2002) can estimate turbulent heat fluxes from point to continental scale with reasonable accuracy using satellite and meteorological data (Oku et al., 2007; Su et al., 2005). While several progresses have been made to SEBS (Gokmen et al., 2012; Kwast et al., 2009), further efforts are needed to improve the accuracy of SEBS over the Tibetan Plateau regions (Su et al., 2006).

Hereby in this chapter, the output produced by SEBS which was driven by meteorological measurement, was compared with flux measurements at four stations. The impacts of model related parameters on estimated sensible heat flux were analyzed. Sensitivity analysis technique was conducted by comparing the sensible flux output of varying input parameters with measurement. A yearly meteorological data and ancillary parameters were compiled at the four sites to perform the sensitivity analysis. Then the turbulent heat parameterization method in SEBS was improved, specifically the equation for $kB^{-1}$.

3.3 Model description and its bias in sensible heat flux simulation

The sensible heat flux ($H$) was computed by means of Monin–Obukhov similarity theory (MOST):

$$H = k u_* \rho C_p (\theta_0 - \theta_a) [ \ln \left( \frac{z-d}{z_{0h}} \right) - \Psi_h \left( \frac{z-d}{L} \right) + \Psi_h \left( \frac{z_{0h}}{L} \right) ]^{-1},$$

(3.1)

where $k$ is von Karman constant; $u_*$ is the friction velocity; $\rho$ is air density; $C_p$ is specific heat for moist air; $\theta_0$ is the potential temperature at the surface; $\theta_a$ is the potential air temperature at height $z$; $d$ is the zero plane displacement height; $\Psi_h$ is the stability correction functions for sensible heat transfer (Brutsaert, 1999), and $L$ is the Obukhov length. Detailed information about $u_*$, and $L$ calculation is referred to Su (2002) and Chen et al. (2013c).

The roughness height for heat transfer ($z_{0h}$) must be accurately determined before the computation of sensible heat flux. $z_{0h}$ is calculated with the following equation:

$$z_{0h} = \frac{z_{0m}}{\exp(kB^{-1})},$$

(3.2)
Chapter 3

The quantity of $kB^{-1}$ has been the subject of study (Beljaars and Holtslag, 1991; Brutsaert, 1982; Kanda et al., 2007; Owen and Thomson, 1963; Sheppard, 1958; Su et al., 2001; Zeng and Dickinson, 1998). Few formulas are able to describe the observed diurnal variation in $kB^{-1}$ over the Tibetan Plateau (Ma et al., 2009b). A $kB^{-1}$ scheme that can account for the diurnal variations in thermal roughness length performs better in the turbulent heat flux estimations and land surface modeling (Yang et al., 2008). On the basis of fractional canopy coverage, Su (2002) presented $kB^{-1}$ as follows (this method will be abbreviated as Su02 hereinafter):

$$kB^{-1} = f_c^2 * kB_c^{-1} + f_s^2 * kB_s^{-1} + 2 * f_c * f_s * kB_m^{-1},$$

where $f_c$ is the fractional canopy coverage and $f_s$ is fraction of bare soil; $kB_c^{-1}$ is the $kB^{-1}$ of the canopy; $kB_s^{-1}$ is that of the bare soil; $kB_m^{-1}$ is $kB^{-1}$ for mixed bare soil and canopy. As the most important parameter in MOST based calculation of sensible heat flux, $kB^{-1}$ will be updated by this study.

The roughness height $z_{om}$ for momentum transfer in equation 3.2 is derived from canopy height ($HC$), leaf area index ($LAI$) and canopy momentum transfer model (Massman, 1997):

$$z_{om} = HC * (1 - d/HC) * \exp(-k \beta),$$

where $d$ is displacement height, which is derived from $HC$ and wind speed extinction coefficient (Su, 2002; Su et al., 2001).

To run SEBS at a point scale, the air temperature, relative humidity, wind speed at 10m height, land surface temperature, pressure, and four-component radiation datasets from the four stations were used. The four stations in this chapter include Qomolangma Station for Atmospheric and Environmental Observation and Research, the Chinese Academy of Sciences (hereafter QOMS/CAS), Nam Co for Multisphere Observation and Research (hereafter Namco), Linzhi Station for Alpine Environment Observation and Research (hereafter Linzhi station), and Maqu station. The introduction of each station is described in section 2.3 of Chapter 2. The data processing of turbulent flux measured by eddy covariance is introduced in section 2.4 of Chapter 2. The detailed information of the input data to run SEBS is listed in Table 3.1.
TABLE 3.1 Meteorological measurements and parameters used to run SEBS at the four station points

<table>
<thead>
<tr>
<th></th>
<th>QOMS/CAS</th>
<th>Namco</th>
<th>Linzhi</th>
<th>Maqu</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Wind speed &amp; direction</strong></td>
<td>height: 10.0 m;</td>
<td>height: 10.0 m;</td>
<td>height: 5.0 m;</td>
<td>height: 8.4 m;</td>
</tr>
<tr>
<td><strong>Air temperature &amp; humidity</strong></td>
<td>height: 10.0 m;</td>
<td>height: 10.0 m;</td>
<td>height: 5.0 m;</td>
<td>height: 8.4 m;</td>
</tr>
<tr>
<td><strong>Surface radiation</strong></td>
<td>CNR-1, Kipp &amp; Zonen</td>
<td>CNR-1, Kipp &amp; Zonen</td>
<td>CNR-1, Kipp &amp; Zonen</td>
<td>CNR-1, Kipp &amp; Zonen</td>
</tr>
<tr>
<td><strong>Pressure</strong></td>
<td>PTB220B,</td>
<td>PTB220B,</td>
<td>PTB220B,</td>
<td>PTB220B,</td>
</tr>
<tr>
<td><strong>HC_{min}</strong></td>
<td>0.002 m</td>
<td>0.002 m</td>
<td>0.002 m</td>
<td>0.002 m</td>
</tr>
<tr>
<td><strong>HC_{max}</strong></td>
<td>0.1 m</td>
<td>0.01 m</td>
<td>0.8 m</td>
<td>0.3 m</td>
</tr>
</tbody>
</table>

The result in figure 3.1 shows that Root Mean Square Error (RMSE) changes from 30 to 35 Wm\(^{-2}\), all the correlation coefficients R\(^2\) are higher than 0.9, and Mean Bias (MB) is below 9.2 Wm\(^{-2}\). The linear fitting line is depicted by a dashed line in figure 3.1. The slopes in figure 3.1 demonstrate that SEBS produces a slightly lower estimation of sensible heat flux at QOMS/CAS and Namco. Figures 3.1c and 3.1d show some sensible heat fluxes simulated by SEBS to be systematically higher than the observations (see the dashed circle), while others are systematically lower (see the solid circle). Even with these systematic deviations in panel (c) and (d), the fitting line and statistical results are still comparable with panel (a) and (b).
Fig. 3.1 Scatter plot of sensible heat flux ($H$, unit of $Wm^{-2}$) between the measurements of eddy covariance (ED) and outputs of SEBS (Su02) at QOMS/CAS (a), Namco (b), Linzhi (c), and Maqu (d) stations. The thick line is the 1:1 line. The linear fitting line is presented as a dashed line. RMSE is the Root Mean Square Error. $R^2$ is the Correlation coefficient. MB is the Mean Bias.

3.4 Method and results

3.4.1 Parameter sensitivity analysis

To calculate the sensible heat flux by means of a similarity theory, the roughness height for heat transfer must be accurately determined. The primary objective of this study is to evaluate the roughness height parameterization method for heat transfer in SEBS. The scalar roughness height for heat transfer ($z_{oh}$) is closely related to the following equations:

$$
\beta = C_1 - C_2 \exp(-C_3 C_d * LAI), 
$$

(3.5a)
An Improvement of the Surface Energy Balance System (SEBS) over the Tibetan Plateau

\[ n_{ec} = \frac{C_d \cdot LAI}{z\beta}, \]  

(3.5b)

\[ d_0 / HC = 1 - \frac{1}{2n_{ec}} \cdot (1 - e^{-2n_{ec}}), \]  

(3.5c)

where \( C_1, C_2, \) and \( C_3 \) are model constants related to the bulk surface drag coefficient (Massman 1997), whose values have been tested over several canopies (Cammalleri et al., 2010; Chen et al., 2013c) and evaluated as one of the best solution for canopy turbulent flux parameterization (Cammalleri et al. 2010). \( C_d \) is the drag coefficient, typically equals to 0.2 (Goudriaan, 1977). LAI is leaf area index, derived from NDVI; \( n_{ec} \) is the wind speed profile extinction coefficient in the canopy; \( d_0 \) is the displacement height; \( z_{0m} \) is the roughness height for momentum transfer; \( \beta (=0.4) \) is the von Karman constant; \( HC \) is the height of the canopy (unit m), which is computed from the NDVI using following equation:

\[ HC = HC_{\text{min}} + \frac{HC_{\text{max}} - HC_{\text{min}}}{(NDVI_{\text{max}} - NDVI_{\text{min}})} \cdot (NDVI - NDVI_{\text{min}}) \]  

(3.6)

where \( HC_{\text{max}} \) and \( HC_{\text{min}} \) are the measured maximum and minimum canopy height; \( HC_{\text{min}} \) is set to 0.0012; and \( HC_{\text{max}} \) is the highest measured value for canopy height in one year.

The uncertainties in \( C_1, C_2, C_3, C_d, HC, \) and \( NDVI \) can directly influence \( d_0, z_{0m}, \) and \( z_{0h} \), which will ultimately determine the accuracy of canopy turbulent flux simulations. The values of \( C_1, C_2, C_3, \) and \( C_d \) have been given by Massman (1997) to parameterize the surface drag coefficient. To date nobody has tested their applicability over the Plateau canopy. Variations in the fitting line’s slope and intercept, RMSE, \( R^2 \), and MB in predicted heat fluxes with \( C_1, C_2, C_3, \) and \( C_d \) are given in our paper. Here we use the eddy covariance observed turbulent fluxes as reference. The slope and intercept, RMSE, \( R^2 \), and MB for all four sites are plotted to analyze their influence on sensible heat flux for different environments and land surfaces.

Figure 3.2 shows that the slope can increase from 0.5 to 1.2, while RMSE and \( R^2 \) vary little, when \( C_1 \) increases from 0.15 to 0.4. The intercept becomes stable when \( C_1 \) is above 0.32. Most of the MB values are negative, which means increasing or decreasing \( C_1 \) will not solve the problem in figure 3.1. Bache (1986) suggests that the \( C_1 \) varies between 0.15 and 0.4 for different canopy types. Based on the sensitivity evaluations, we suggest that the value of \( C_1 \) is not adjusted.
Fig. 3.2 Variations of intercept (a), slope (b), RMSE (c), $R^2$ (d), and MB (e) derived from observed and simulated sensible heat fluxes due to $C_1$ changes from 0.1 to 0.4. The vertical line corresponds to $C_1=0.32$. 
The results in figure 3.3 demonstrate that all five variables vary little regardless of changes in $C_2$. This shows that $C_2$ is not a principal parameter, and therefore its original value has not been altered. Concerning $C_3$ (fig. 3.4), the intercept and $R^2$ change little with changes in $C_3$, while slope, RMSE and MB are sensitive to $C_3$ changes when $C_3$ falls below 15. On the contrary, when $C_3$ is above 15, the statistical variables become stable. The values of slope may change with $C_3$ while the MB remains negative. This means there are more points distributed below the 1:1 line. This result also displays the inability to solve problems of SEBS sensible heat flux by adjusting $C_3$. Based on the above sensitivity analysis of the five statistical variables, the original values of $C_1=0.32$, $C_2=0.26$, and $C_3=15.1$ have been used here. We believe they are suitable for the canopy of the Tibetan Plateau.

$Cd$ is the foliage drag coefficient as a function of height within the canopy, which is assumed to be uniform throughout the low canopy. The influence of $Cd$ on the five statistical variables was estimated, as shown in figure 3.5. At Namco all five variables changed little with $Cd$. The slopes were around 0.75, and the average MB was 10 Wm$^{-2}$. Thus changing $Cd$ will not resolve the lower $H$ estimation problem in SEBS at Namco. When $Cd$ increased from 0.1 to 0.5, the slope of QOMS/CAS was enhanced from 0.75 to 0.92, and the MB was reduced to zero. Thus, $Cd$ changes can solve the problem of SEBS at QOMS/CAS. The MB values at Linzhi and Maqu were still negative when $Cd$ increased. The slopes of Linzhi and Maqu station decreased persistently from 1.2 and 1, respectively, with $Cd$ increasing. From these results, we can see that increasing $Cd$ has a different impact on the four stations’ slope and MB results. This study disclosed that the lower estimation of sensible heat flux at Namco is not caused by the $Cd$ value. While for the dense canopies at Linzhi and Maqu, the influence of $Cd$ becomes important. Based on these sensitivity analyses, we can conclude that the systematic errors of the SEBS sensible heat flux at the four stations cannot be solved by adjusting the values of $C1$, $C2$, $C3$, and $Cd$. 

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An Improvement of the Surface Energy Balance System (SEBS) over the Tibetan Plateau
Fig. 3.3 Variations of intercept (a), slope (b), RMSE (c), $R^2$ (d), and MB (e) derived from observed and simulated sensible heat fluxes due to $C_2$ changes from 0.1 to 1. The vertical line correspond to $C_2=0.26$. 
Fig. 3.4 Variations of intercept (a), slope (b), RMSE (c), R² (d), and MB (e) derived from observed and simulated sensible heat fluxes due to $C3$ changes from 5 to 50. The vertical line corresponds to $C3=15.1$. 

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Fig. 3.5 Variations of intercept (a), slope (b), RMSE (c), $R^2$ (d), and MB (e) derived from observed and simulated sensible heat fluxes due to $Cd$ changes from 0.1 to 0.5. The vertical line correspond to $Cd=0.2$. 

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Fig. 3.6 Variations of intercept (a), slope (b), RMSE (c), R² (d), and MB (e) derived from observed and simulated sensible heat fluxes due to $HC_{max}$ changes from 0.01 to 0.5. The vertical lines from left to right correspond to $HC_{max}$ equals to 0.01, 0.1, 0.3, and 0.8.
Kwast et al. (2009) have pointed out that the canopy height ($HC$) can cause large deviations in the modeling of sensible heat flux. As shown in Equation 3.6, the magnitude of $HC$ is determined by $HC_{\text{max}}$ and $NDVI$. $HC_{\text{max}}$ was also used in the sensitivity analysis for $C1$, $C2$, $C3$, and $Cd$, with results presented in figure 3.6. All the slope values at the four stations increased from below 1 to above 1 and the MB values decreased from positive to negative with increasing canopy height. The figure reveals that the accuracy of the estimated canopy height is important for the final fluxes. The errors in canopy height input values can be settled by measuring the height of the canopy at each station. The measured $HC_{\text{max}}$ values which have been used in SEBS were listed in Table 3.1. Alternatively, the sensitivity test of $HC_{\text{max}}$ can also be used to appraise the measured canopy height. The practical canopy height has been estimated by in situ site inspection and used in this study. Thus, $HC_{\text{max}}$ cannot trigger SEBS to produce above regular sensible heat flux bias. As other variables, such as $NDVI$, $LAI$, wind speed, and air temperature, are also taken from measurements, we assume their accuracy is influenced by the equipment. Their sensitivity will not be discussed further. A final option rests with the parameterization of $kB^{-1}$.

3.4.2 Improvement of roughness height for heat transfer

From time series comparison with measurement, we found that lower estimation of sensible heat fluxes occurred during winter (Fig. 3.7). Based on the time series comparison analysis, the significant lower estimations were at daytime. A diurnal pattern exists in the difference between observed and simulated sensible heat fluxes. In Equation 3.3, $kB^{-1}$ was divided into three parts: bare soil, canopy, and mixed canopy and soil. In winter, coverage of the land surface was dominated by bare soil. The role of bare soil on $kB^{-1}$ should be an important part. As a test over the bare soil of the Tibetan Plateau (Chen et al., 2010; Zhang, 2010), the heat roughness height parameterization method of Yang et al. (2002) was introduced into the $kB^{-1}$ of SEBS (abbreviated as SY02). The $kB_{s}^{-1}$ of bare soil in Equation 3.3 was revised according to the following two equations:

$$z_{oh} = \frac{70.\theta}{u_{*}} \exp(-7.2 \cdot u_{*}^{0.5} \theta_{s}^{0.25})$$  \hspace{1cm} (3.7a)

$$kB_{s}^{-1} = \log \left( \frac{z_{am}}{z_{oh}} \right)$$  \hspace{1cm} (3.7b)

where $\theta$ is the kinematic viscosity of air ($1.5\times10^{-5}$ m$^2$s$^{-1}$), $u_{*}$ is the surface friction velocity (m s$^{-1}$), and $\theta_{s}$ is the surface friction temperature (K).
Fig. 3.7 Scatter plot of sensible heat flux (H, unit of W m⁻²) between the eddy covariance and outputs of SEBS (Su02) sensible heat flux data for different months at Maqu station. Other abbreviations are similar to those for Fig. 3.1.
As $z_{0h}$ is computed with $u_*$ and $\theta_*$ in Equation 3.7a, the diurnal change information in $u_*$ and $\theta_*$ will be transported to $z_{0h}$, and then to $kB_{-1}$. The diurnal change information of $z_{0h}$ and $kB_{-1}$ is important in the calculation of surface flux through $z_{0h}$ or $kB_{-1}$ in MOST. This makes $kB_{C_{-1}}$ value in Equation 3.7b better than the fixed $kB_{C_{-1}}$ value of Equation 3.3. After upgrading $kB_{C_{-1}}$ in SEBS ($kB_{C_{-1}}$ and $kB_{m_{-1}}$ are the same as in Su et al. (2001)), the results are depicted in the figures 3.8. The statistical variables before and after the improvement are all listed in Table 2.3 for comparison. The scatter distributions in figure 3.8 are found closer to the 1:1 line than in figure 3.1. The monthly scatter plots of Maqu station also demonstrate an enhancement in SEBS performance during winter time (Fig. 3.9).

After introducing the equation of $z_{0h}$ by Yang et al. (2002), the linear fitting lines are situated closer to the 1:1 line (Fig. 3.8). Other statistical values show little change (Table 3.2). The results show a better performance of the new $kB_{-1}$. The method is preferable for the Tibetan Plateau.

<table>
<thead>
<tr>
<th>TABLE 3.2 Statistics results for simulated and observed sensible heat flux</th>
<th>QOMS/CAS</th>
<th>Namco</th>
<th>Linzhi</th>
<th>Maqu</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Su02</td>
<td>SY02</td>
<td>Su02</td>
<td>SY02</td>
</tr>
<tr>
<td>Slope</td>
<td>0.9</td>
<td>1.0</td>
<td>0.8</td>
<td>1.1</td>
</tr>
<tr>
<td>Intercept(Wm$^{-2}$)</td>
<td>3.9</td>
<td>5.7</td>
<td>0.7</td>
<td>4.5</td>
</tr>
<tr>
<td>RMSE(Wm$^{-2}$)</td>
<td>35.4</td>
<td>34.5</td>
<td>35.5</td>
<td>38.4</td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.9</td>
<td>0.9</td>
<td>0.9</td>
<td>0.9</td>
</tr>
<tr>
<td>MB (Wm$^{-2}$)</td>
<td>1.7</td>
<td>-6.1</td>
<td>9.9</td>
<td>-8.3</td>
</tr>
</tbody>
</table>
Fig. 3.8 Scatter plot of sensible heat flux (H, unit of Wm^-2) between the measurements of eddy covariance and outputs of SEBS (SY02) at QOMS/CAS (a), Namco (b), Linzhi (c), Maqu station (d).
Fig. 3.9 Scatter plot of sensible heat flux (H, unit of Wm⁻²) between the eddy covariance and outputs of SEBS (Su02) sensible heat flux data for different months at Maqu station. Other abbreviations are similar to those for Fig. 3.1.
3.4.3 Further test of the $kB^{-1}$ scheme for the Tibetan Plateau

For the snow surface, where $f_c$ in equation 3.3 equals 0, SY02 SEBS will use equations 3.7a and b to compute sensible heat flux. The $z_{om}$ for snow surface is given a value of 0.0001 (Arck and Scherer, 2002). A snow depth of 0.4 m was manually measured between November 25 and December 31, 2006 at Namco station. The measurements during this period were used to evaluate the performance of SY02 for the snow surface (figures 3.10 and 3.11). The atmospheric surface layer over snow covered land is often dominated by a very stable stratified layer. Under stable atmospheric stratification, turbulent events are rather intermittent. The intermittence of the turbulence causes violations of the eddy-covariance assumptions (steady state conditions, flux-variance similarity), and yields a significant misestimation of the sensible heat flux (Lüers and Bareiss, 2010). It is well documented that stable atmospheric stratification presents numerous challenges to the interpretation of surface fluxes measured using the eddy-covariance technique, e.g. Aubinet (2008), and Helgason and Pomeroy (2011). In our study, we also found abnormal turbulent fluxes measured by the eddy covariance over the snow surface. Especially in cases of calm wind, most of the measured fluxes were abnormal. Measuring eddy covariance is more difficult during very stable air temperature stratification, low heat fluxes, and low wind velocities (Foken and Wichura, 1996; Marks et al., 2008). When a gust of wind occurs, indicated by periods of yellow background in figure 3.10, the mechanical turbulence caused by the wind shear assists the heat exchange from the air to the snow. It may also increase the depth of the surface layer, which rises to above the sensors of the eddy covariance. During such disturbance periods, sensible heat fluxes measured by eddy covariance become reasonable. In order to obtain good quality data, a standard of friction velocity $> 0.2 \text{ ms}^{-1}$ was set to determine time periods of strong wind shear. The selected data of eddy covariance were used to make a comparison with the values simulated by SEBS. The scatter points and fitting line are shown in figure 3.11. This result testifies the validity of the present roughness height for heat transfer over a snow surface. Guo et al. (2011) also supported the reliable sensible heat flux estimated with Yang et al. (2002)’s method over a Tibetan glacier snow surface. All these results indicate that our new revised heat parameterization is superior when applied to the Tibetan Plateau’s bare soil, canopy or even snow surface.
Fig. 3.10. Comparison of friction velocity ((a) $U_{\text{star}}$, unit of ms$^{-1}$), sensible heat flux ((b) $H$, unit of Wm$^{-2}$) between eddy covariance and SEBS (SY02) from November 26, 2006 to December 31, 2006 at Namco site when the land surface is covered by snow. The periods of yellow background provide a good comparison between SEBS (SY02) and eddy covariance measurements.

Fig. 3.11 Scatter plot and linear fitting of sensible heat flux (Wm$^{-2}$) at the Namco site between eddy covariance and SEBS (SY02).
In this study, SEBS was driven by a time series of meteorological observation data at point scales. The performance of SEBS has been evaluated by comparisons between its simulated sensible heat fluxes and a high quality dataset of observed turbulent fluxes. The measurements were performed over typical Tibetan land-cover units. The results show that SEBS captured the diurnal and seasonal variation of sensible heat flux well. But it also showed that the Su02 formulation of $kB^{-1}$ tends to produce lower estimates for sensible heat flux over bare soil. Using sensitivity analyses, we have identified the most critical factor in the determination of effective roughness height. The wind speed profile extinction coefficient within canopy proposed by Massman (1997) is applicable over the Plateau canopy structure. The practical solution to replace the soil part of the $kB^{-1}$ parameterization with the one by Yang et al. (2002) was proposed. The new parameterization equation of $kB^{-1}$ in SEBS has been tested and found to be an improvement over the original one for bare soil, and low canopies. Observations also evaluated the validity of the new heat roughness height parameterization of $kB^{-1}$ over snow surfaces. The revision of heat roughness length parameterization provides SEBS with an improved accuracy and applicability over complex surfaces. The new scheme of turbulent heat flux parameterization method will be useful for land-surface interactions in the Tibetan Plateau climate models. This work helps to improve applicability of the model over typical land covers of the Plateau, and helps us to analyze the possibility and suitability of SEBS to generate surface turbulent heat flux maps over the whole plateau area by remote sensing technique. Evaluated here over different land surface types, the sensible heat parameterization method can also be used in other regions of the Tibet Plateau. It is also worth to assess the generality of the approach presented here for other regions.

In practice, $kB^{-1}$ was derived from the bulk transfer formulation using measurements of other quantities. Any uncertainties associated with these measurements will cause uncertainties in the evaluation of $kB^{-1}$ and the final estimated sensible heat flux. The flux and meteorological variables may have different up-wind source area due to different measurement heights. This also explains the scatter points in figure 3.1 and 3.7. The surface temperature was determined using radiometers with a limited field of view. An assumption was made to regard this temperature as representative of the eddy covariance system’s fetch area. The discrepancy shown by the scatter points is believed to
be related to differences in footprint of the sensors and caused by effects of the inhomogeneous terrain.

SEBS has lower estimated momentum flux over the snow cover. This can be explained because it was given an ideal roughness length value of flat snow cover. Actually the constructions around the measurement can also influence its representative value. The roughness length in the direction of a building on site is bigger than the pre-established value in the model (Fig.5 in Zhou et al. (2010)). The heterogeneity of the land surface around the eddy covariance device makes the actual roughness bigger than the ideal snow roughness length. Secondly, MOST uses an iterative scheme to calculate sensible heat flux. The initial value of $u_*$ is given a value of neutral condition, without stability correction. The initial value is already lower than the true value. All these explain why the simulated momentum flux is lower than observations. Here we revised the method for heat roughness length, which also has dependency on stability. Improving stability-dependent profile functions in MOST may also help us to get more accurate surface fluxes.
An Improvement of the Surface Energy Balance System (SEBS) over the Tibetan Plateau
Chapter 4 Estimation of surface energy fluxes over a mountainous area of the Tibetan Plateau

4.1 Abstract

As most part of the Tibetan Plateau topography was dominated by mountains, knowledge of the spatial distribution of solar radiation in mountainous area is therefore vital for the energy exchange process between the atmosphere and the plateau mountain land surface. This study developed a DEM based radiation model to estimate instantaneous clear sky solar radiation for surface energy balance system to obtain accurate energy absorbed by the mountain surface. Efforts to improve spatial accuracy of satellite based surface energy budget in mountainous regions were made in this chapter. Based on eight scenes of Landsat TM/ETM+ (Thematic Mapper / Enhanced Thematic Mapper+) data and observations around the Qomolangma region of the Tibetan Plateau, the topographical enhanced surface energy balance system (TESEBS) was tested for deriving net radiation, ground heat flux, sensible heat flux and latent heat flux distributions over the heterogeneous land surface. The land surface energy fluxes over the study area showed a wide range in accordance with the surface features and their thermodynamic states. The model was validated by observations at QOMS/CAS site in the research area with a reasonable accuracy. The mean bias of net radiation, sensible heat flux, ground heat flux and latent heat flux is lower than 23.6 Wm$^{-2}$. The surface solar radiation estimated by the DEM based radiation model developed by this study has a mean bias as low as -9.6 Wm$^{-2}$. TESEBS has a decreased mean bias about 5.9 Wm$^{-2}$ and 3.4 Wm$^{-2}$ for sensible heat and latent heat flux, respectively, compared to SEBS.

* This chapter is based on:
4.2 Introduction

Mountainous area covers about one-fifth of the earth’s continental areas (Yang et al., 2011). Accurate surface solar radiation estimations are essential for studies of solar energy resource, hydrological processes, and climate change. Solar radiation exerts strong control on available energy exchanges at the surface. Knowledge of the spatial distribution of solar radiation in mountainous area is therefore vital for the energy exchange process between the atmosphere and the mountain land surface. Terrain determines whether a surface receives direct radiation or if it is shaded. In zones of complex topography, variability in elevation, surface slope and aspect create strong spatial heterogeneity in solar radiation distribution, which determines air temperature, soil temperature, evapotranspiration, snow melt and land--air exchanges. The spatial and temporal distribution of surface radiation exerts a fundamental control on mass and energy exchange between air and land. The mountainous areas are often remote and inaccessible to carry out measurement of land--air interactions. The zones of complex topography therefore form interesting but little studied areas for land-air exchange studies.

Recent studies have explored approaches to estimate the regional distribution of surface heat fluxes with observational data of different satellite sensors (Ma et al., 2009a; Ma et al., 2006; Ma et al., 2011; Oku et al., 2007). Remote sensing based turbulent flux algorithms can be divided into two-source (like TSEBS of Anderson et al., 2008) and single source models (like SEBS of Su, 2002). Yang et al. (2003) pointed out that a single source heat transfer model is applicable on the Tibetan plateau; SEBS was adopted here. The Surface Energy Balance System (SEBS) developed by Su (2002) has been designed to estimate energy partitioning by using satellite and meteorological data. While most of the studies using SEBS derive surface energy balance items located at flat areas (Su et al., 2005; Yang et al., 2010a), none of them consider the influence of topography. With the development of satellite sensor grid resolution, when applying SEBS to the high resolution satellite dataset, the topographic influences become increasingly important. Terrain controls how much sky is visible and therefore influences incident diffuse and reflected sky radiation. Since surface solar radiation measurement is very sparse in the mountainous region, the knowledge of the terrain is thus important for the radiation balance and further for the surface energy balance in complex terrain (Aguilar et al., 2010; Long et al., 2010; Tovar-Pescador et al., 2006). The aim of this research
was to combine a topographically corrected solar radiation (the amount of shortwave radiation received under clear-sky conditions) with SEBS over the Tibetan Plateau mountain area. A topographically enhanced surface energy balance system (TESEBS) was developed to generate a series of distributions of surface energy balance in a meso-scale area on the north area of Mt. Qomolangma over the Plateau. Small lakes, rivers, glacier, and surfaces with short canopies are all included in the study area (Fig. 4.1).

The surface energy balance analysis around Mt. Everest was studied with measurement at point scale (Zhong et al., 2009; Zou et al., 2009). The aim of this research is to upscale in situ point observations of land surface variables and land surface heat fluxes over regional scale using high resolution remote sensing data. In mountainous regions, due to the complex topography, high-
resolution data are needed. Landsat TM/ETM+ sensors include optical and thermal sensors with a relative high image resolution. Here we use Landsat data to determine regional land surface heat fluxes around the area.

The shortwave radiation reaching the surface of the earth is divided into direct, diffuse or reflected radiation. Direct radiation reaches the surface of the earth from the solar beam without interactions with particles in the atmosphere. Diffuse radiation is scattered out of the solar beam by gases and aerosols before reaching the surface. Reflected radiation is mainly reflected to the surface from surrounding terrain and is therefore important in mountainous areas. A knowledge of the values for each component is often required when considering the topographic effects on each radiation component separately (Aguilar et al., 2010). To get an accurate incoming solar radiation flux in mountainous terrain, a radiation model which considers the shading and reflecting effects of adjacent features is needed by SEBS. At each point, the direct, diffuse, and reflected solar radiations were estimated. The global radiation was obtained by adding the direct, diffuse and reflected radiation. The intention of this study is to compute the instantaneous solar radiation with the above three radiation variables for various slopes and azimuth terrains in order to make each component of the energy balance system more accurate.

4.3 Model formulation

The surface energy balance equation is written as

\[ R_n = G_0 + H + LE, \]

where \( R_n \) is the net radiation; \( G_0 \) is the ground heat flux; \( H \) is the turbulent sensible heat flux, and \( LE \) is the turbulent latent heat flux. Latent heat flux \( LE \) is computed as the last item of surface energy balance equation after derivations of other three variables in Equation 4.1. The sensible heat flux was computed by means of Monin–Obukhov similarity theory (MOST) theory which is described in last chapter.

4.3.1 The instantaneous net radiation

Net radiation is a critical input variable in the energy balance equation and the most sensitive variable in latent heat flux estimate (Zhang et al., 2005). Therefore, the accuracy of retrieved net radiation determines the accuracy of estimates of latent heat flux and ET (evapotranspiration) to some extent.
the computation of surface energy balance in complex terrain, a highly detailed surface radiation balance model is necessary. Topography is well known to alter the shortwave radiation balance at the surface. In order to use the surface energy balance equation over the complex topography of the Plateau, further efforts are needed to improve spatial accuracy of satellite based surface energy budget in mountainous regions. A detailed radiation balance model is therefore required by this study. Thus here we introduce the surface radiation model as follows.

The net radiation flux is computed with the following equation:

$$ R_n = (1 - \alpha) SWD + LWD - LWU, $$

(4.2)

Where $\alpha$ is the broadband albedo, SWD is downward surface shortwave radiation. LWD and LWU are downward and upward surface longwave radiation. On flat terrain and under clear-sky conditions, the downwelling shortwave radiation is nearly the same from point to point over relatively large areas and so one measurement can be taken to be representative of the entire regional area (Bastiaanssen, 2000; Kwast et al., 2009). However, direct measurements are rarely available to represent the shortwave radiation over most mountainous areas. Therefore, in mountainous regions a detailed solar radiation balance solution is required by surface energy balance equation. Parameterization models are often used to make predictions of individual solar radiation components under clear sky conditions (e.g., Yang et al. (2001), Liang et al. (2012)). Meanwhile, the topographic effects are rarely considered. Here we employ surface radiation parameterization models and solar radiation transfer above an inclined surface to account for the impact of complex terrain, which follow the simple form of the Angstrom–Prescott model (Prescott, 1940), and its inputs (air temperature and relative humidity) are easily accessible from routine surface meteorological observations.

### 4.3.2 The instantaneous downward shortwave radiation

The surface downward shortwave (solar) radiation is divided into three parts over complex terrain: direct radiation ($I_d$), diffuse ($I_a$) and reflected ($I_r$) insolation.

$$ SWD = I_b + I_d + I_r. $$

(4.3)

The downward shortwave radiation varies in response to altitude, surface slope and aspect. The parameterization schemes for calculating the instantaneous solar radiation were improved by accounting for the three part variations to slope and azimuth of land surface and terrain shadow in mountainous areas. Studies have described how to use a digital elevation map dependent model to
compute direct solar radiation, diffuse and reflected insolation (Kumar et al., 1997). According to the knowledge, the method used to compute distribution of downward shortwave radiation over mountainous areas is as follows:

\[
I_b = I_0 \cdot \tau_c \cdot \cos \theta, \quad (4.4)
\]

\[
I_d = I_0 \cdot \tau_d (\cos s)^2 / (2 \sin a), \quad (4.5)
\]

\[
I_r = r \cdot I_0 \cdot \tau_r (\sin s)^2 / (2 \sin a), \quad (4.6)
\]

where \( \theta \) is solar incidence angle; \( a \) is the solar altitude angle; \( s \) is the tilt angle of the surface (slope); \( \tau_c \) solar beam radiative transmittance; \( \tau_d \) solar diffuse radiative transmittance; \( \tau_r \) is the reflectance transmittance; \( r \) is the ground reflectance.

The applications of optical remote sensors by SEBS are limited to conditions of cloudless sky; Therefore, in this part only atmospheric transmittances under conditions of cloudless sky was considered. The clear sky radiative transmittance is based on local geographical and meteorological conditions. The clear sky surface solar radiation is affected by a number of extinction processes in the atmosphere. Although Kumar et al. (1997) suggested a method for computation of \( \tau_c, \tau_d \) and \( \tau_r \), their method did not consider the difference in Rayleigh scattering, aerosol extinction, ozone absorption, water vapor absorption and permanent gas absorption. Yang et al. (2001) developed a broadband radiative transfer model based on Leckner’s (1978) spectral model. As evaluated as one of the best broadband models (Gueymard, 2003a; Gueymard, 2003b), here we used the model to calculate solar beam radiative transmittance \( \tau_c \), and solar diffuse radiative transmittance \( \tau_d \), under clear skies. \( \tau_c \) is computed as function of radiative transmittance due to ozone absorption, water vapour absorption, permanent gas absorption, Rayleigh scattering, and aerosol extinction, respectively. The detailed solution of \( \tau_c \) is described in Appendix A. \( \tau_d \) and \( \tau_r \) are computed with \( \tau_c \). The solutions for \( I_d \) and \( I_r \) are presented in Appendix B.

The influence of tilted surface on surface radiation is expressed by solar incidence angle, solar altitude angle and topographic information shown in equations of 4.4 - 4.6. A high resolution DEM map (obtained from the U.S. Geological Survey Earth Resources Observation and Science center) of SRTM (Shuttle Radar Topography Mission) in the region was used to calculate slope and aspect of each pixel. The slopes and aspects were then used in subsequent executions to generate solar radiations in complex mountainous area.
4.3.3 The instantaneous downward and upward surface longwave radiation

It is relatively easy to estimate incoming longwave radiation under clear sky conditions. Different parameterizations for atmospheric longwave radiation were tested for clear sky periods (Brunt, 1932; Iziomon et al., 2003; König-Langlo and Augstein, 1994; Prata, 1996), but Brutsaert’s (1975) method was among the best performance in the computations of incoming longwave radiation (Iziomon et al., 2003; Kimball et al., 1982; Kustas et al., 1994). Brutsaert’s (1975) parameterization method is expressed as following:

\[
LWD = \varepsilon_a \sigma T^4, \tag{4.7}
\]

\[
\varepsilon_a = 1.24 e^{0.14286}, \tag{4.8}
\]

where \(\sigma\) is the Stefan-Boltzmann constant \(5.67 \times 10^{-8} \text{ Wm}^{-2} \text{K}^{-4}\). Air emissivity \(\varepsilon_a\) is determined by actual water vapor pressure \(e\) (hPa) and air temperature \(T\) (K).

Longwave emission from different terrains was taken as isotropic here. The upward longwave radiation is computed using the Stefan–Boltzmann equation:

\[
LWU = \varepsilon_a LST^4, \tag{4.9}
\]

where \(\varepsilon\) is the “broad-band” land surface emissivity, derived from a satellite based method referred to Chen et al. (2013). When \(\text{NDVI} < 0\) and \(\alpha < 0.47\), the pixel was taken as water surface, where \(\varepsilon = 0.985\); If \(\alpha \geq 0.47\) the pixel was assumed as snow, where \(\varepsilon = 0.99\). LST is land surface temperature.

4.3.4 The instantaneous ground heat flux

The regional ground heat flux \(G_0\) in equation 4.1 cannot directly be mapped from satellite observations. The ground heat flux, as an indirect variable in surface energy balance, was calculated through net radiation according to a different surface dependent ratio value. The relationship between \(G_0\) and \(Rn\) (Kustas and Daughtry, 1990) over bare soil in this area is

\[
G_0 = 0.315 Rn \tag{4.10}
\]
For water area (NDVI < 0 and α < 0.47), we use the equation: 

\[ G_0 = 0.5 \times Rn \] (Gao et al., 2011); For glacier area, \( G_0 \) is negligible according to Yang et al. (2011a) and we use an equation of 

\[ G_0 = 0.05 \times Rn. \]

The glacier is distinguished according to LST \( \leq 273 \) K. Daughtry et al. (1990) investigated that the midday \( G_0/Rn \) ratio is predictable from vegetation indices. For canopy coverage area, the following equation is adopted:

\[ G_0 = Rn \left( f_c \times r_c + r_s \times (1 - f_c) \right) \] (4.11)

\( rs \) and \( r_c \) are the ratio between ground heat flux and net radiation for bare soils and surfaces with fully covered vegetation separately. The \( rs \) in this area is given an value of 0.315 (Kustas and Daughtry, 1990), and \( r_c \) has a value of 0.05 (Monteith, 1973). The fractional vegetation cover \( f_c \) is determined using the normalized difference vegetation index NDVI in equation 4.16.

### 4.3.5 The instantaneous sensible heat flux density

The updated equations for sensible heat flux in Chapter 3 was used in this study. Here we will not describe it again.

### 4.4 Study site and data processing

#### 4.4.1 Study area

The Himalaya, as the south barrier of the Plateau with large area of high mountains directly pierces to the middle troposphere on the earth, it has great influence on the weather and climate over there (Bollasina and Benedict, 2004; Gao, 1981; Ueno et al., 2008; Ye and Gao, 1979; Zhong et al., 2009; Zhou et al., 2008; Zou et al., 2009). The Himalaya exerts profound thermal and dynamical influence on atmospheric circulation (Bollasina and Benedict, 2004). The Himalaya mountains provide the water sources for Indus, Ganges, and Brahmaputra rivers, which supply water to billions of people in Asia (Immerzeel et al., 2010). Considering the importance to Asian monsoon (Boos and Kuang, 2010), Bollasina and Benedict (2004) pointed out that land-atmosphere interactions over Himalaya needs particular attention. The study was conducted in 28°0'42"N to 29°0'54.9"N and 86°9'5"E to 87°1'42.8"E, located around QOMS/CAS station (Figure 4.1), a comprehensive observation and research station on the north slope of Himalaya. Elevation in the study area changes between 3700 to 7107 m. This region has been chosen because it is a
representative of a high-alpine environment and glaciers, lakes, rivers, and short canopies present in this area (see bottom left picture of Figure 4.1).

### 4.4.2 Field measurements

The Qomolangma Station for Atmospheric and Environmental Observation and Research, Chinese Academy of Sciences (QOMS/CAS) locates at 28°21.63’N, 86°56.93’E, with an elevation of 4276 m, and 30 km away from Mount Everest. The dataset of the QOMS/CAS station consist of surface radiation budget components (CNR-1, Kipp & Zonen), vertical profiles of air temperature, humidity, wind speed and direction (MILOS520, Vaisala), turbulent fluxes measured by eddy correlation technique. Sensors of wind speed, wind direction, air temperature, and relative humidity at five levels (1.0, 2.0, 4.0, 10.0, and 20.0 m) are installed on a 40 m PBL tower. The soil temperature was measured at depths of 0, 10, 20, 40, 80 and 160 cm. The soil moisture content was measured at depths of 10, 20, and 30 cm. The soil heat flux was measured using soil heat flux plates (HFP01) buried at a depth of 10 cm. The soil heat flux at the surface is calculated by adding the measured flux at 10cm depth to the energy stored in the layer above the heat flux plates.

The accuracy of the model was evaluated using Mean Bias ($MB$) and Root Mean Square Error ($RMSE$). These statistical indicators are defined as:

$$RMSE = \sqrt{\frac{\sum_{i=1}^{N}(x_i - obs_i)^2}{N}}$$  \hspace{1cm} (4.12)

$$MB = \frac{\sum_{i=1}^{N}(obs_i - x_i)}{N}$$  \hspace{1cm} (4.13)

where $x_i$ is simulations of TESEBS, $obs_i$ is observations, and $N$ is the sample number.

### 4.5 Remote sensing data preparation

To capture the heterogeneity of the land surface over the domain, high-resolution satellite data are required. The Landsat TM/ETM+ data can provide high-resolution information on the land surface temperature, land cover classification, albedo, and the NDVI. Eight scenes of Landsat TM/ETM+ datasets were collected on 10 March 2008, 26 March 2008, 11 April 2008, 29 May 2008, 02 September 2008, 20 October 2008, 21 November 2008, and 9 April 2010 with cloudless skies. The fraction of cloud cover is not more than 3%
on these days. In mountainous areas, topography also strongly influence the signal recorded by space-borne optical sensors. The topographic influence on the satellite received signal were eliminated by the method of Richter et al. (2009). The reflectivity for each band ($\rho_\lambda$) is calculated as

$$\rho_\lambda = \frac{\pi L_\lambda}{E_{\text{SUN},\lambda} d_r \cos \theta} \quad (4.14)$$

where $L_\lambda$ is the spectral radiance for each band; $E_{\text{SUN},\lambda}$ is the mean solar exo-atmospheric irradiance for each band; $\theta$ is solar incidence angle; $s$ is the surface slope. The solar exo-atmospheric irradiance for TM and ETM+ 1, 2, 3, 4, 5, 7 band in Markham and Barker (1987) and Chander and Markham (2003) were used separately. $\theta$ is the solar zenith angle (from nadir), and $d_r$ is the inverse squared relative Earth-Sun distance. The surface albedo ($\alpha$) for shortwave radiation is retrieved from converting narrowband to broadband planetary reflectivity which is obtained as the total sum of the different narrow-band reflectivity according to weights for each band. The weights for the different bands are given by Teixeira (2010). Broadband shortwave surface albedo was calculated from the normalized reflection values of channels 1, 2, 3, 4, 5 and 7, using the following equation:

$$\alpha = 0.293 \cdot \rho_1 + 0.274 \cdot \rho_2 + 0.233 \cdot \rho_3 + 0.157 \cdot \rho_4 + 0.033 \cdot \rho_5 + 0.011 \cdot \rho_7, \quad (4.15)$$

$\rho_1, \rho_7$ are the reflectivity for band 1 to 7. The spectral radiance of band 6 is converted into a brightness temperature applicable at the top of the atmosphere by inversion of Plank’s law (Teixeira, 2010). The LST is calculated by monowindow algorithm (Qin et al., 2001; Sobrino et al., 2004). The NDVI is computed as the ratio of the differences in reflectivities for the near-infrared band ($\rho_4$) and the red band ($\rho_3$) to their sum. The vegetation fractional coverage is estimated using formulation:

$$f_c = \frac{\text{NDVI}(x,y) - \text{NDVI}_{\text{imin}}}{\text{NDVI}_{\text{imax}} - \text{NDVI}_{\text{imin}}}, \quad (4.16)$$

The value of $\text{NDVI}_{\text{imin}}$ and $\text{NDVI}_{\text{imax}}$ is about 0.2 and 0.5 (Sobrino et al., 2004). In the case of NDVI<0.2, $f_c=0$. The broad band emissivity $\varepsilon$ is used to calculate total long wave radiation emission from the surface. This broadband emissivity was calculated from the NDVI according to the method of Sobrino et
To maintain a spatial consistency, NDVI, albedo and other data were interpolated to corresponding thermal infrared band using a linear technique.

### 4.6 Meteorological data

To compute surface fluxes over the coverage area of one satellite image, the spatial distribution of meteorological data (air temperature, atmospheric pressure, relative humidity etc.) at PBL-height or near near-surface height at satellite pixel scale is required. Spatial interpolation method is often used to get these meteorological data from meteorological stations or atmospheric reanalysis data (Ma et al., 2006; McVicar et al., 2007; Oku et al., 2007). Xie et al. (2007) pointed out that meteorological elements above Mt. Everest coincided with measurements at a meteorological station 60 km away. Then we assume the meteorological measurement at QOMS/CAS station can represent the synoptic situation of our research area during clear sky conditions. Air temperature for each grid cell was adjusted with respect to elevation, assuming a standard air temperature lapse rate of 6 K/km (Chen et al., 2007). Atmospheric boundary layer height of 600 m is used according to the results of Ma et al. (2009b). Due to the elevation changes from 4000 to higher than 8000 m, the corresponding surface pressure changes significantly. Thus the surface pressure is estimated by

\[
p_s = p_0 \exp(-z/8430),
\]  

where \( p_0 = 101325 \) Pa, \( z \) is DEM data in a unit of m.

### 4.7 Evaluations of TM/ETM+ based TESEBS results

Instantaneous surface energy balance items at the satellite overpass time are highly dependent on the estimation of key variables, namely, land surface temperature, albedo, downward and upward shortwave/longwave radiation, etc. Hence, we evaluate these variables by comparison with site-specific ground-based measurements. The correlation coefficient of downward shortwave radiation (SWD) between TESEBS simulated and measured by radiometers is as high as 0.99, with MB of -9.62 Wm\(^{-2}\) and RMSE of 45.4 Wm\(^{-2}\). This demonstrates that the DEM based radiation model performs very well over the Tibetan mountainous region. The LST has a mean bias of 1.46 K with a high R value. The albedo derived from TM/ETM+ is a little bit lower than the in-situ true values. The SWU of TESEBS has a mean value of 42.55 Wm\(^{-2}\) lower than
the observation. The final surface energy balance items were also evaluated with in-situ measurement (Fig. 4.3). The MB values of $R_n$, $H$, $G_0$ and LE is about -23.6, -15.8, 7.7 and -6.8 Wm$^{-2}$ separately. Overall, the values derived from TM/ETM+ by TESEBS agree well with ground measurements. The comparison of TESEBS and SEBS statistical variables were listed in Table 4.1. Both RMSE and MB of TESEB are lower than SEBS, which shows the better performance of TESEBS.

![Fig. 4.2 Comparison of derived results with field measurements for the (a) land surface temperature (LST, unit K), (b) albedo, (c) downward shortwave radiation (SWD, unit Wm$^{-2}$), (d) downward longwave radiation (LWD, unit Wm$^{-2}$), (e) upward shortwave radiation (SWU, unit Wm$^{-2}$), (f) upward longwave radiation (LWU, unit Wm$^{-2}$).](image-url)
### Table 4.1 The comparison between TESEBS and SEBS statistical variables

<table>
<thead>
<tr>
<th></th>
<th>TESEBS</th>
<th>SEBS</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Sensible heat flux (Wm(^{-2}))</strong> RMSE</td>
<td>31.1</td>
<td>34.6</td>
</tr>
<tr>
<td></td>
<td>MB</td>
<td>-15.8</td>
</tr>
<tr>
<td><strong>Latent heat flux (Wm(^{-2}))</strong> RMSE</td>
<td>25.0</td>
<td>25.1</td>
</tr>
<tr>
<td></td>
<td>MB</td>
<td>-6.8</td>
</tr>
</tbody>
</table>

![Graphs showing comparison](image)

**Fig. 4.3** Comparison of derived results with field measurements for the (a) net radiation (Rn, unit Wm\(^{-2}\)), (b) sensible heat flux (H, unit Wm\(^{-2}\)), (c) latent heat flux (LE, unit Wm\(^{-2}\)), (d) ground heat flux (G0, unit Wm\(^{-2}\))
Figure 4.4 shows the distribution maps (1605×1882 pixels) of each surface energy balance item based on TM remote sensing data obtained on 9 April, 2010. The overpass time is 10:32 LT (local time), the sun azimuth angle is 127.2 degree, and the sun elevation is 58.15 degree. To maintain consistency in spatial resolution, the remote sensing data of band 1, 2, 3, 4, 5, 7 were interpolated to 120 m ×120 m. The experimental area presents extreme variability characterized not only by steep slopes and altitude variations of thousands of meters, but also by a variety of land surfaces such as grassy marshland, several small rivers and lakes, bare soil and glaciers. Therefore, these derived parameters show a wide range due to the strong contrast of surface features. The sensible heat flux over the water body is lower than in other places around it, while the latent heat flux is higher (Fig. 4.4 (a)). The sensible heat flux over the glacier area is dominated by negative values. Slopes which are exposed to the east consequently have a higher net radiation and sensible heat flux than that of slopes with an exposition to the west.

The processes and mechanisms of energy and mass transfers become complicated under the complex terrain due to the substantial differences in radiation availability caused by various slopes and azimuths of surfaces. The surface net radiation in the region changes from -155 to 461.5 Wm\(^{-2}\), sensible heat flux from -25.2 to 265.6 Wm\(^{-2}\), and latent heat flux from -4.5 to 257.9 Wm\(^{-2}\). The surface flux maps reflect distinct mechanisms of energy partition and water evaporation of various land cover types, showing differences in the spatial distribution pattern of surface turbulent heating. The derived distribution maps over our study area were found to be reasonable.

West facing slopes are receiving only about half of the shortwave radiation of the eastern slopes (Fig. 4.6 (a)). Such variations would surely have a significant effect on the heat budget of different places, thus influencing latent and sensible heat fluxes. It can be seen that the values in the region present a huge difference in the distributed values of the variable, but also noticeable was the wide range of global values in the region, when topographic factors are taken into account, as shown by the maximum, minimum values. The spatial gradient in each part of solar radiation is evident (Fig. 4.5). It can be seen that the locations receiving more radiation are those in the highest part of the region, with an east-facing orientation that remains unobstructed during the hours of satellite pass over. While the west-facing slopes which are shaded by the terrain at the satellite pass-over time receive a relative low solar beam radiation (Fig. 4.5b), and a
relative high diffuse radiation (Fig. 4.5c). The surface energy balance has the greatest influence on environmental processes, especially if snow is present. Over the snow covered glacier area on the bottom left corner of the map (Fig. 4.5d), the reflected solar radiation can be around 100 Wm$^{-2}$, which is significant compared to low values of other surface energy variables over there. This also demonstrate that the glacier topography does play a fundamental role in determining the surface energy balance (Arnold et al., 2006).

![Fig. 4.4 Distribution of each surface energy balance item over the study area at 10:30 (LT) on 09 April 2010, the cross line is the location of the station. The geographical information is the same as bottom-left picture of figure 4.1.](image)
Fig. 4.5 Distribution of each surface solar radiation item over the study area at 10:30 (LT) on 09 April 2010, the cross shows the location of the station. The geographical information is the same as bottom-left picture of figure 4.1.
Fig. 4.6 Distribution of each surface radiation balance item over the study area at 10:30 (LT) on 09 April 2010, the cross line is the location of the station, SWD and SWU are downward and upward surface shortwave radiation. LWD and LWU are downward and upward surface long-wave radiation. The geographical information is the same as bottom-left picture of figure 4.1.
4.8 Conclusion

Dealing with regional land surface heat fluxes over heterogeneous landscapes is not an easy job. In order to analyze the interactions between the complex land surface and the atmosphere above it, a Topographical Enhanced Surface Energy Balance System (TESEBS) was developed to upscale energy and turbulent heat fluxes from point to a meso-scale mountainous area. When using high resolution satellite data over mountainous areas to get the surface energy balance items, the terrains effects must be considered. In this study, a radiation parameterization scheme for grid topography accounting for shading, and terrain reflections is used to get the surface radiation and energy balance system. DEM information was used to characterize the topographic role in the spatial distribution of surface energy balance items in Qomolangma region’s complex topography. Each radiative flux is parameterized individually as a function of slope, sun elevation angle, and albedo. We quantify the topographic impacts on each individual shortwave radiation (solar beam, diffuse and reflected radiation) with real topographies. Variations in atmospheric transmissivity resulting from actual column water vapor, ozone, and aerosol have been considered in our clear-sky satellite applications.

TESEBS was evaluated from site point and a regional scale. Firstly TESEBS was forced by a long time meteorological observation data at point scale. The performance of TESEBS has been evaluated by comparisons between its output fluxes and a high quality dataset of observed turbulent fluxes. Then TESEBS was expanded to a regional scale, where glaciers, bare soil and small lakes all present. It’s difficult to get spatial distributions of meteorological inputs, hereby we spatialized the meteorological measurements at the station to present synoptic situation over the complex region. The distributions of surface energy balance correspond well with the land surface class. The significant incidence of topography on the values of surface energy balance throughout the region has been demonstrated by the proposed topographic solar radiation algorithm. Results indicate that surface flux predictions from TESEBS perform well at local scales on a flat terrain, when assessed against in situ flux measurements derived from eddy covariance approaches, and provides realistic outputs at regional scale for more complex topography. Due to lack of measurement on different aspect and slopes, it is impossible to do fully evaluations for this complex topography.
This work helps us to analyze the possibility and suitability of TESEBS to model surface turbulent heat flux over typical land covers of the Plateau by remote sensing technique. The performance of TESEBS over the glaciers also makes it possible to study the energy balance of the snowpack, and validate snowmelt runoff model in future. Opportunities also exist for improving the performance of both models via data assimilation and model calibration techniques that integrate remote sensing based surface energy flux predictions. However, the topography effects on the roughness length still remain a blank area at present. Advection could be formed under complex terrain, which may complicate the energy balance at point scale. In the future, further validation of the parameterization method needs to be made over the Plateau water surface and other land covers.
Appendix A. Direct radiation

Due to scattering processes by molecules (Rayleigh scattering) and by aerosols (Mie scattering) as well as due to absorption processes by different components of the atmosphere only a fraction of solar radiation is received as global radiation at the surface. Simple broadband transmittance functions for each atmospheric constituent are therefore commonly applied to solar radiation in order to obtain the spectrally integrated direct and diffuse sky radiation components. Solar radiation is attenuated as it passes through the atmosphere and, in a simplified case:

$$I_0 = S_0 \times (1 + 0.0344 \times \cos(\frac{\pi d\text{oy}}{365})) \times \sin \alpha,$$

(A1)

$S_0$ (1367 Wm$^{-2}$) is solar constant; $d\text{oy}$ is day of the year; $\alpha$ is the solar altitude angle. The equation accounts for variation in the solar irradiance at the top of the atmosphere throughout the year.

The last stage is to calculate the solar radiation on a tilted surface. Incident global radiation is defined as the sum of incident direct (beam) radiation ($I_b$), incident diffuse sky radiation ($I_d$) due to scattering processes in the atmosphere, and incident radiation received from the surrounding terrain due to reflections ($I_r$).

$$I_b = I_0 \times \tau_c \times \cos \theta,$$

(A2)

$$\tau_c = \max(0, \tau_{oz} \times \tau_w \times \tau_g \times \tau_r \times \tau_a - 0.013),$$

(A3)

$$\tau_{oz} = \exp(-0.0365 \times (m \times l)^{0.7136}),$$

(A4)

$$\tau_w = \min[1, 0.909 - 0.036 \times \ln(m \times w)].$$

(A5)

$$\tau_g = \exp(-0.0117 \times m_c^{0.3139}),$$

(A6)

$$\tau_r = \exp[-0.008735m_c(0.547 + 0.014m_c - 0.0038m_c^2 + 4.6 \times 10^{-6} m_c^3)^{-4.08},$$

(A7)

$$\tau_a = \exp[-m \beta(0.6777 + 0.1464m \beta - 0.00626(m \beta)^2)^{-1.3}].$$

(A8)
\[ m = \frac{1}{\sin h + 0.15(57.296h + 3.885)^{-1.253}}, \]  
(A9)

\[ m_c = m \frac{p_s}{p_0}, \]  
(A10)

where \( \tau_{oz} \), \( \tau_w \), \( \tau_g \), \( \tau_r \), and \( \tau_a \) are the radiative transmittance due to ozone absorption, water vapour absorption, permanent gas absorption, Rayleigh scattering, and aerosol extinction, respectively. \( m \) is the air mass, \( m_c \) the pressure-corrected air mass, \( h \) (radian) the solar elevation, \( p_s \) surface pressure (given by equation 4.15) and \( p_0 = 1.013 \times 10^5 \) Pa. \( l \) is the thickness of the ozone layer (unit cm or 1000 Dobson Units), \( \beta \) Angstrom turbidity coefficient. \( w \) is the precipitable water defined as the amount of water in a vertical column of atmosphere. Humidity profile measurements of the atmosphere are needed to calculate the precipitable water needs, which is usually unavailable at surface meteorological stations. In this model, the precipitable water \( w \) (cm) is estimated from surface relative humidity \( \text{Rh} \) (%) and air temperature \( T \) (K) by a semi-empirical formula (Yang et al. 2006)

\[ w = 0.00493 \text{Rh} T^{-1} \exp(26.23 - 5416T^{-1}), \]  
(A11)

The ozone optical depth used in this study was computed using the determinations of total ozone columnar concentration from data obtained NASA/ GSFC Ozone Processing Team. The equation for calculation of \( \cos \theta \) is described in Appendix C independently.
Appendix B. Diffuse and reflected solar radiation

Diffuse solar radiation ($I_d$) was calculated using the method suggested by Gates (1980),

$$I_d = I_0 \tau_d (\cos s)^2 / (2 \sin a), \quad (B1)$$

where $\tau_d$ is the diffuse radiation transmissivity. $\alpha$ is the solar altitude angle. $s$ is the tilt angle of the surface (slope).

$$\sin \alpha = \sin L \sin \delta_s + \cos L \cos \delta_s \cos h_s, \quad (B2)$$

$L$ is the latitude, solar declination ($\delta_s$) and hour angle ($h_s$).

The magnitude of reflected radiation depends on the slope of the surface and the ground reflectance coefficient. The reflected radiation here is the ground-reflected radiation, both direct sunlight and diffuse skylight, impinging on the slope after being reflected from other surfaces visible above the slope’s local horizon. The reflecting surfaces are considered to be Lambertian. Here reflected radiation ($I_r$) was calculated based on the method of Gates (1980):

$$I_r = r I_0 \tau_r (\sin s)^2 / (2 \sin \alpha). \quad (B3)$$

where $r$ is the ground reflectance. $s$ is slope. $\alpha$ is the solar altitude angle. $\tau_r$ is the reflected radiation transmissivity. $\tau_r$ can be related to $\tau_e$ by the relationship in the following equation (Kumar et al., 1997):

$$\tau_r = 0.271 + 0.706 \tau_e, \quad (B4)$$
Appendix C. Calculation of the cosine of the solar incidence angle ($\cos \theta$)

The solar incidence angle is the angle between the solar beam and a line perpendicular to the land surface. In the flat model, we assume that the land surface is horizontal and the calculation of $\cos \theta$ is very simple and is a constant over the area of interest. In the Mountain area, $\cos \theta$ is different for each pixel depending on the slope and aspect of the land surface. The following equations are used to compute $\cos \theta$:

$$
\cos \theta = \sin \delta \sin \phi \cos s - \sin \delta \cos \phi \sin s \cos \gamma + \cos \delta \cos \phi \cos s \cos \omega + \cos \delta \sin \phi \sin s \cos \gamma \cos \omega + \cos \delta \sin \phi \sin s \sin \omega,
$$

where $\delta$ is declination of the earth (in radians, positive in summer in northern hemisphere)

$$
\delta = 0.409 \sin\left[\frac{2 \pi \times \text{day}}{365} - 1.39\right]
$$

$\phi =$ latitude of the pixel (in radians, positive for northern hemisphere). $s =$ slope (radians) where; $s=0$ is horizontal and $s=\pi/2$ is vertical downward ($s$ is always positive and represents a downward slope in any direction).

$\gamma =$ surface aspect angle (in radians) where; $\gamma = 0$ for due south, $\gamma = +\pi/2$ for east, $\gamma = -\pi/2$ for west and $\gamma = \pm \pi$ for north.

$\omega = \pi \times (t-12)/12$, hour angle (in radians). $t$ is the local standard time. $\omega = 0$ at solar noon, $\omega$ is negative in morning and $\omega$ is positive in afternoon.
Chapter 5 Development of a 10-year (2001-2010) high resolution land surface energy balance product over the Tibetan Plateau

5.1 Abstract

In the absence of high resolution estimations of the components of the surface energy balance over the Tibetan Plateau, we developed an algorithm based on surface energy balance system (SEBS) to generate a dataset of land surface energy and water fluxes on a monthly time scale from 2001 to 2010 at a 0.1 × 0.1 degree resolution by using multi-satellite and meteorological forcing data. A remote-sensing based method was developed to estimate canopy height which was used to calculate roughness length and fluxes dynamics. The land surface flux dataset was validated by using ‘ground-truth’ observations at 11 flux tower stations in China. The estimated fluxes correlate well with the stations’ measurement for different vegetation types and climate conditions and present an average bias of 15.3 Wm$^{-2}$, RMSE of 26.4 Wm$^{-2}$. The results show that our method is a preeminent one for producing a high resolution surface energy flux for Tibetan Plateau landmass with satellite data. The validation results demonstrated that more accurate downward long-wave radiation datasets are needed in order to accurately estimate turbulent fluxes and evapotranspiration when using surface energy balance model. A trend analysis of the land surface radiation and energy exchange fluxes revealed that the Tibetan Plateau area is experiencing relative stronger climate changes than other parts of China during last 10 years. The capability of the dataset to provide spatial and temporal information on water cycle and land-atmosphere interactions for China landmass is examined.

* This chapter is based on:
5.2 Introduction

Growing evidence has shown that a striking climate warming occurred over the TP during the second half of the twentieth century (Liu and Chen, 2000; Zhu et al., 2001). Understanding of the interaction between land and atmosphere and its response to climate change will enable us to confront and hopefully solve the environmental problems. The variability of surface energy balance and its partitioning may also have an important impact on local climate variability. Changes in the plateau land surface energy fluxes alter the differential heating of land and ocean, which in turn affects the intensity of the East Asian monsoon (Zhou and Huang, 2008). Understanding of variation in plateau energy fluxes is important for the study of climate change. While it is of critical importance to understand the partitioning of water and energy distribution across the plateau’s terrestrial surface, accurate monitoring their spatial variation is difficult.

Several field experiments are being carried out to monitor selected land cover over the plateau by using ground-based eddy covariance measurements (Ma et al., 2008b; Wang et al., 2010; Yu et al., 2006). However, these results only represent small areas around the locations where the measurements are being made. Establishment of an eddy covariance flux network cannot provide a complete land surface heat flux picture for the entire plateau area.

There are several product related to land surface energy fluxes. Jung et al. (2009) generated global spatial fluxes fields by using a network up-scaling method, however their flux network does not include flux stations on the plateau. The Global Soil Wetness Project 2 (GSWP-2) (Dirmeyer et al., 2006) produced a global land surface product on a 1 × 1 degree grid from 1986 to 1995, The Global Land Data Assimilation System (GLDAS) (Rodell et al., 2004) can provide a global coverage in the form of 3 hourly, 0.25 degree data. European Centre for Medium-Range Weather Forecasts (ECMWF) interim re-analysis (ERA-Interim) (Dee et al., 2011), National Centers for Environmental Prediction (NCEP) (Kalnay et al., 1996), Modern-Era Retrospective Analysis for Research and Applications (MERRA) (Rienecker et al., 2011) and other reanalysis data can also provide a temporal continuous but coarse spatial resolution dataset of land surface fluxes. When these products were applied at large scales, the different approaches have resulted in large difference (Jiménez et al., 2011; Mueller et al., 2011; Vinukollu et al., 2011a). The problems met by using present flux data in climate studies in China have been reported by Zhou
Zhu et al. (2012) reported that summer sensible heat flux from 8 dataset (including NCEP, ERA, GLDAS etc.) in Tibetan Plateau region differ from one to another in spatial distribution. In addition, all the flux datasets mentioned above is based on model simulations, which has different disadvantages for studying changes in the water cycle and land-air interactions over the Tibetan Plateau (Chen et al., 2013d; Ma et al., 2008a; Su et al., 2013; Wang and Zeng, 2012).

A spatially and temporally explicit estimate of surface energy fluxes is of considerable interest for meteorological, climatological investigations and hydrologic assessment (Norman et al., 2003). Satellite-based surface variables can be used to produce maps of heat and water fluxes at different scale (Li et al., 2012; Liu et al., 2010; Vinukollu et al., 2011b; Wang and Liang, 2008). Remote-sensing approaches to estimate surface heat and water fluxes have been often used on a regional scale over the plateau (Ma et al., 2011) but plateau-scale water and energy fluxes are still scarce. Indeed, there is no analysis of satellite-derived data currently underway to produce a complete, physically-consistent, decadal land surface heat flux dataset (Jiménez et al., 2009) for the plateau area. The use of remotely-sensed data offers the potential of acquiring observations of variables such as albedo, land surface temperature, NDVI and so on. Fig. 5.1 gives an example of NDVI map over China land. Since surface fluxes cannot be directly detected by satellite-borne sensors, an alternative for estimating continental water and energy fluxes can be derived by applying aerodynamic theory of turbulent flux transfer (Ma et al., 2011) or by building statistic relationships between related satellite observations and land surface fluxes (Jiménez et al., 2009; Wang et al., 2007). Most remote-sensing-born latent heat flux or evapotranspiration product has problems at pixels with non-canopy coverage (Jiménez et al., 2009; Mu et al., 2007; Wang et al., 2007). This is due to physical process was not included in the application of their method. Statistical methods build the relationship between the satellite observations (e.g. NDVI, LST, albedo) and land surface fluxes through different fitting technique (Wang et al., 2007). The built simple relations cannot give reasonable approximation over extreme conditions, like bare soil or other non-canopy land covers (e.g. water, desert), because land covers behave significantly different in land surface energy flux partitioning. Meanwhile, the turbulent flux transfer parameterization can overcome the shortcomings of statistical method and can produce spatially continuous distributions of land surface energy fluxes with
prepared meteorological information. That’s why we choose a physical based method – turbulent flux parameterization to produce the dataset.

The challenge in turbulent flux parameterization approach is in the transition from regional to plateau and continental scales, because meteorological data at a high resolution (i.e. 1 -10 km) are not easily accessible for a large region. Recently, Chinese scientists have produced a high resolution meteorological forcing data, which makes possibilities of this study. Another issue is the complexity met with the method when combining different spatial and temporal sampling input variables. The detailed consideration in this area will be informed in the section 5.4. The last difficulty which has surrounded the applications of turbulent flux parameterization method to remote sensing dataset at continental area scale during last 10 years is the geophysical information of roughness length. Here we developed a remote sensing mixing technique to estimate canopy height in a continental area and used the canopy height dataset to derive dynamical variations of surface roughness length for the plateau. This study sets out to estimate turbulent heat fluxes simulated with energy balance and aerodynamic parameterization formulas based on a revised surface energy balance system (SEBS) model (Chen et al., 2013b; Chen et al., 2013c; Su, 2002; Timmermans, 2011), which has been tested with a better performance and improvements in case of bare soil, short canopy and snow surface of the Tibetan area (Chen et al., 2013b; Chen et al., 2013c). Sensible heat flux in SEBS was derived from the difference between surface temperature and air temperature by using Monin–Obukhov similarity theory and bulk atmospheric boundary layer similarity (Brutsaert, 1999), which parameterizes ground surface momentum and heat transfer coefficient maps to take into account surface roughness, canopy height, vegetation cover, and meteorological stability (Chen et al., 2013c; Su, 2002; Su et al., 2001). The latent heat flux can then be estimated from an energy balance model, assuming surface net radiation and ground flux are known (Allen et al., 2011; Ma et al., 2002; Vinukollu et al., 2011b).
Fig. 5.1 A NDVI map for China landmass based on SPOT satellite. The terminology is: Tibetan Plateau (TP), northwestern China (NWC), inner Mongolian Plateau (MP), Loess Plateau (LP), north China Plain (NP), northeastern China Plain (NEP); Pearl River delta (PRD), Sichuan (SCB), Yinchuan (YCB), the inner Mongolian basins (IMB), Lasha (LB), Tarim (TRB), Junggar Basin (JB); Himalaya mountain (HM), Gangdes mountain (GM), Kunlun mountain (KLM), Karakorum mountain (KRM), Tianshan mountain (TM), Nyainqentanglha mountain (NQM) and Qilian mountain (QLM). The short names of the flux stations are in yellow color. Blue lines show several biggest rivers in China. Black lines give borders of each province.
Development of a 10-year (2001-2010) high resolution land surface energy balance

Complex topography makes it very difficult to obtain a picture of energy and heat fluxes distribution, especially with a high spatial resolution over a relatively long time for the plateau area. To derive the surface energy fluxes for the plateau area, we used a high resolution reanalysis data, which merges model outputs, remote-sensing observations, and in-situ measurements. In addition, we also assessed the accuracy of the surface energy balance terms (net radiation, sensible heat, latent heat, and ground heat fluxes) and their climatic trends in recent 10 years. After describing the equations of SEBS model and required input dataset, we evaluated the fluxes dataset. Further, we assess the capacity of the remote-sensing-based product to reproduce the range and variability of measured fluxes by comparing them with in-situ measurements, followed by trend analysis of the spatial patterns of the fluxes.

5.3 Model description and development

The surface energy balance model known as SEBS (Su, 2002) uses aerodynamic resistance to create a spatially estimate of land heat fluxes. Some model inputs can be obtained from remote sensing data while others can be obtained from meteorological forcing data (e.g. GLDAS, ERA and NCEP reanalysis data). The model’s equations and the required forcing variables are described in the remainder of this section.

The surface energy balance equation can be expressed as:

\[ R_n = G_0 + H + LE, \]  

(5.1)

where \( R_n \) is the net radiation flux; \( G_0 \) is the ground heat flux, which is parameterized by relationship with \( R_n \) (Su et al., 2001); \( H \) is the turbulent sensible heat flux; and \( LE \) is the turbulent latent heat flux. \( LE \) is computed by using the evaporative fraction after deriving the other three variables in equation 5.1 and taking into consideration of energy and water limits (Su 2002). Since the fluxes were produced with a monthly average temporal resolution, energy storage in vegetation is not considered.

Net radiation flux is computed with the following equation:

\[ R_n = (1 - \alpha) \ SWD + LWD - LWU, \]  

(5.2)
where $\alpha$ is broadband albedo; SWD is downward surface shortwave radiation; LWD and LWU are downward and upward surface longwave radiation, respectively.

Here satellite observed albedo is used. LWU is derived from land surface temperature (LST) using the Stefan–Boltzmann law. Land surface emissivity is derived as described in Chen et al. (2013b). LWD and SWD values are obtained from meteorological forcing data.

The equation of sensible heat flux ($H$) is included in Chapter 3. The new $kB^{-1}$ was directly used in this Chapter. Here we will not repeat it.

As Chen et al. (2013c) have pointed out that canopy height (HC) is vital for the sensible heat simulations, making an accurate estimation of HC is important for this study. A remote-sensing-based method (Chen et al., 2013c) was further developed to estimate China land canopy height distribution. A global forest canopy-height map was produced by Simard et al. (2011) using data from the Geoscience Laser Altimeter System (GLAS) aboard ICESat (Ice, Cloud, and land Elevation Satellite). However, short canopy (e.g. maize, rice, wheat) height information cannot be provided by laser technique. Short canopy height usually has a seasonal variation every year, as crops are planted in spring and harvested in autumn. To accommodate the seasonal variations in short canopy height, we calculated short canopy height used an NDVI-based equation from Chen et al. (2013c):

$$HC = HC_{\text{min}} + \frac{HC_{\text{max}} - HC_{\text{min}}}{NDVI_{\text{max}} - NDVI_{\text{min}}} \times (NDVI - NDVI_{\text{min}}), \quad (5.3)$$

where $HC_{\text{max}}$ and $HC_{\text{min}}$ are the maximum and minimum short canopy height. $HC_{\text{min}}$ is set to 0.0012; and $HC_{\text{max}}$ is set to 2.5 m, corresponding to the highest height of seasonal crops. $NDVI_{\text{min}}$ and $NDVI_{\text{max}}$ are minimum and maximum NDVI during the 10 years. Each short canopy pixel was given a $NDVI_{\text{min}}$ and $NDVI_{\text{max}}$ value to calculate the canopy height. This NDVI-based short canopy-height equation was used to fill the pixels with forest canopy heights lower than 10 m. The high canopy heights (greater than 10 m) were assumed to be constant with seasonal change. By merging high canopy heights and variable short-canopy data, we constructed monthly dynamic maps of canopy heights for the period of 2000 – 2010. These maps were used to calculate heat fluxes.
5.4 Input dataset

The approach uses a variety of satellite-sensor products and meteorological forcing data to estimate monthly energy and water fluxes in China. The forcing data can be from satellite based or reanalysis dataset. Due to the influence of cloud, satellite sensed visible and thermal band data (e.g. NDVI, albedo, LST) often have spatial and temporal gaps in a daily data. Various temporal and spatial gap-filling algorithms have been developed to produce continuous monthly data for satellite-sensed variables (Chen et al., 2004; Moody et al., 2005). In order to avoid both spatial and temporal gaps in the final product, we selected some specific satellite-sensed datasets for this study. Due to the input dataset not only covers the area of Tibetan Plateau but also China landmass, we determine to produce a flux product for China landmass. Table 5.1 lists all input data sets used in this study.

The longest period covered by the forcing dataset is approximately 31 years; the shortest is about 10 years. Spatial resolution of the dataset varies from 0.01 to 0.25 degree and its sample frequency from 3 hours to 1 month. The meteorological forcing data developed by the Institute of Tibetan Plateau, Chinese Academy of Sciences (hereinafter referred to as ITPCAS forcing data) (He, 2010) was constructed to study meteorological variations in China. ITPCAS forcing data covers the entire land mass of China and has the highest temporal resolution of all the input datasets. Other variables such as LST and albedo, for example, have coarser temporal resolutions monthly and global coverage. When combing data at different spatial and temporal resolution, both spatial and temporal scaling issues need to be addressed.

Estimates of land surface energy flux can be subject to large errors, due to bias in the meteorological forcing inputs. The spatial distribution of meteorological variables is closely related to topography. When interpolating meteorological input variables to finer scales, these effects have to be accounted for (Sheffield et al., 2006), which is beyond the scope of this study. Therefore, it is better to resample satellite products of high spatial resolution to a lower spatial resolution that matches the resolution of the meteorological input data. ITPCAS forcing data provides us with the highest spatial resolution among present available meteorological forcing data (e.g. ERA-interim, NCEP, GLDAS, MERRA). Taking into account all these effects, our aim was to produce monthly product of 0.1 × 0.1 degree resolution land surface heat fluxes, which
contains neither spatial nor temporal gaps and can be used to study seasonal and inter-annual variability in the hydrological and energy cycles in China.

Table 5.1 Input data sets used for computing China land flux estimates (see Sect. 5.3 and 5.4 for explanation of abbreviations).

<table>
<thead>
<tr>
<th>Variables</th>
<th>Data source</th>
<th>Temporal resolution</th>
<th>Availability</th>
<th>Domain</th>
<th>Spatial resolution</th>
<th>Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>SWD</td>
<td>ITPCAS</td>
<td>3 hours</td>
<td>1979-2010</td>
<td>China land</td>
<td>0.1 deg.</td>
<td>Reanalysis</td>
</tr>
<tr>
<td>SWU</td>
<td>ITPCAS &amp; GlobAlbedo</td>
<td>3 hours</td>
<td>1982-2009</td>
<td>China land</td>
<td>0.1 deg.</td>
<td>Satellite &amp; Reanalysis</td>
</tr>
<tr>
<td>LWD</td>
<td>ITPCAS</td>
<td>3 hours</td>
<td>1979-2010</td>
<td>China land</td>
<td>0.1 deg.</td>
<td>Reanalysis</td>
</tr>
<tr>
<td>LWU</td>
<td>MOD11C3</td>
<td>1 month</td>
<td>2000-2012</td>
<td>China land</td>
<td>0.05 deg.</td>
<td>Satellite</td>
</tr>
<tr>
<td>Ta</td>
<td>ITPCAS</td>
<td>3 hours</td>
<td>1979-2010</td>
<td>China land</td>
<td>0.1 deg.</td>
<td>Reanalysis</td>
</tr>
<tr>
<td>Q</td>
<td>ITPCAS</td>
<td>3 hours</td>
<td>1979-2010</td>
<td>China land</td>
<td>0.1 deg.</td>
<td>Reanalysis</td>
</tr>
<tr>
<td>Ws</td>
<td>ITPCAS</td>
<td>3 hours</td>
<td>1979-2010</td>
<td>China land</td>
<td>0.1 deg.</td>
<td>Reanalysis</td>
</tr>
<tr>
<td>P</td>
<td>ITPCAS</td>
<td>3 hours</td>
<td>1979-2010</td>
<td>China land</td>
<td>0.1 deg.</td>
<td>Reanalysis</td>
</tr>
<tr>
<td>LST</td>
<td>MOD11C3 V5</td>
<td>1 month</td>
<td>2000-2012</td>
<td>Global</td>
<td>0.05 deg.</td>
<td>Satellite</td>
</tr>
<tr>
<td>$h_e$</td>
<td>GLAS &amp; SPOT VEGETATION</td>
<td>1 month</td>
<td>2000-2012</td>
<td>China land</td>
<td>0.01 deg.</td>
<td>Satellite</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>GlobAlbedo</td>
<td>1 month</td>
<td>2000-2010</td>
<td>Global</td>
<td>0.05 deg.</td>
<td>Satellite</td>
</tr>
<tr>
<td>NDVI</td>
<td>SPOT VEGETATION</td>
<td>10 days</td>
<td>1998-2012</td>
<td>Global</td>
<td>0.01 deg.</td>
<td>Satellite</td>
</tr>
</tbody>
</table>

5.4.1 Meteorological forcing data

In studies previous to this one, reanalysis data have been applied in many different ways, for example, to construct land surface forcing data (Sheffield et al., 2006), to detect climate trends (Taniguchi and Koike, 2008), and to investigate water and energy cycles between land and atmosphere at regional and continental scales (Roads and Betts, 2000). Another application of reanalysis data was in deriving global terrestrial evapotranspiration and gross primary production by remote sensing community (Mu et al., 2007; Yuan et al.,
2010). However, few studies have used them with remote sensed ground status to derive global land energy flux (sensible heat flux, latent heat flux, net radiation etc.).

Researchers have developed several kinds of reanalysis data. Inter-comparisons and evaluations of these reanalysis products with in-situ observations have been performed for individual sites, specific region, and global land (Decker et al., 2011; Wang and Zeng, 2012). It is well known that inaccuracies existing in the reanalysis forcing data may have substantial impacts on the simulation of land surface energy partitioning. It is difficult to make an appraisal on which reanalysis data is better when they are used as a forcing data. Additionally, the spatial resolution of all of the above reanalysis/forcing datasets is not as high as remote sensing data. The ITPCAS forcing dataset was produced by merging a variety of data sources. In particular, this dataset benefits from the merging information from 740 weather stations operated by China Meteorological Administration that have not been used in other forcing data. The dataset has already been used to run land surface models and has been shown to be more accurate than other forcing datasets (Chen et al., 2011b; Liu and Xie, 2013). ITPCAS meteorological forcing data include instantaneous near surface air temperature (Ta), near surface air pressure (P), near surface air specific humidity (Q), near surface wind speed (Ws) at a temporal resolution of 3 hours, 3-hourly mean surface downward short-wave (SWD) and downward long-wave (LWD) radiation. The time period coverage is from 1979 to 2010, with a grid-size spatial resolution of 0.1 × 0.1 degree.

5.4.2 MOD11C3 land surface temperature

MODIS sensors have been used to produce several global and continental scale LST datasets. MOD11C3 V5 products (Wan, 2009) are validated over a range of representative conditions with an average bias of less than 1 K (Coll et al., 2009; Wan and Li, 2008). The MOD11C3 V5 LST product has a 0.05 degree grid size, a monthly temporal resolution without gaps, and covers the period March 2000 to October 2012. It provides monthly daytime and nighttime LST values. Daytime and nighttime values were averaged by this study to represent monthly means.

After interpolation MOD11C3 V5 LST to 0.1 × 0.1 degree resolution, we picked out LST values of pixels including the 11 flux stations used in this paper. The pixel values were validated against in-situ LST measurements. The detailed
information of each station was described in section 5.5.1. The linear correlation ($R = 1.0$), RMSE ($= 1.9$ K) and MB (mean value of the satellite data minus in-situ observation, $= 0.5$ K) indicate that the quality of the remotely-sensed LST data in China is high. It also shows that MOD11C3 V5 LST captures the in-situ LST variability at different elevations and land surfaces, which will be shown in section 5.5.2.

### 5.4.3 Albedo

Land surface albedo determines the fraction of short-wave radiation absorbed by the ground, thus influences the surface energy budget. Studies of land surface energy balance need temporal and spatial gapless albedo input data. Several research teams have been devoted to produce long-term time series of surface albedo from various satellite-borne sensors (Liu et al., 2012; Muller et al., 2012; Riihel et al., 2013). However most of the albedo products do not provide time series gap-filled albedo maps. Taking MODIS (Moderate-resolution Imaging spectroradiometer) MCD43B albedo product as an example, 20 to 40% of the global land pixels miss valid albedo values every year (Liu et al., 2012). A 20% invalid values in albedo input data will cause same amount empty value in heat flux output. This point does limit albedo data which can be used in this study. After checking several albedo product (including GlobAlbedo (Muller et al., 2012), CMSAF cLouds, Albedo and RAdiation Surface Albedo (CLARA-SAL albedo) (Riihel et al., 2013), and MCD43B), we chose to use GlobAlbedo which does not include spatial and temporal gaps. The albedo dataset is based on a monthly sample and has a spatial resolution of 0.05 degree, which was then interpolated to a 0.1 degree resolution for our study.

### 5.4.4 NDVI

The Normalized Difference Vegetation Index (NDVI) is regarded as a good indicator of vegetation parameters such as vegetation coverage, and leaf area index. NDVI has been widely used to explore vegetation dynamics and their relationships with environmental factors (Piao et al., 2006). NDVI data from the Systeme Pour l’Observation de la Terre (SPOT) VEGETATION sensor, distributed by Vito (http://free.vgt.vito.be/), has a spatial resolution of 1 km × 1 km and a temporal resolution of 10 days (synthesized on days 1, 11 and 21 of each month). In order to reduce noise resulting from clouds, the maximum NDVI value in a month for each pixel is selected to represent the canopy status of that month.
5.4.5 Canopy fraction

Canopy fraction \( f_c \) is defined as the fraction of ground surface covered by the vegetation canopy (varying from 0 to 1). \( f_c \) in SEBS is used to distinguish the contributions of vegetation and soil to the roughness parameterization. Here \( f_c \) was derived from NDVI data using the following equation,

\[
f_c = \frac{NDVI - NDVI_{min}}{NDVI_{max} - NDVI_{min}}
\]

5.5 Evaluations with flux network

5.5.1 Flux network measurement

The product generated by our model needs to be validated by comparing it with an independent observational dataset. The energy balance measurement system (eddy covariance, four component radiation and ground heat flux) at flux sites is widely accepted as a method for direct measurement of energy and fluxes and it is widely applied for assessing global evapotranspiration products (Fisher et al., 2008; Jung et al., 2011; Yan et al., 2012; Zhang et al., 2010a).

To validate the product, we collected a dataset from 11 China flux stations including bare soil, alpine meadow, forest, cropland, orchard, grassland, and wetland land covers. Site elevations range from 5 m to 4800 m. The observational dataset included data from three plateau station, such as Maqu (MQ) (Chen et al., 2013c; Wang et al., 2013), Bijie (BJ), Ali (AL) (Ma et al., 2006), and low elevation site, such as Wenjiang (WJ) (Zhang et al., 2012a), Miyun (MY) (Liu et al., 2013a), Daxing (DX) (Liu et al., 2013a), Guantao (GT) (Jia et al., 2012; Liu et al., 2011; Liu et al., 2013a), Yucheng (YC) (Flerchinger et al., 2009), Dongtan (DT) (Zhao et al., 2009), SACOL (SC) (Guan et al., 2009; Huang et al., 2008; Wang et al., 2010), and Weishan (WS) stations (Lei and Yang, 2010a; Lei and Yang, 2010b). Detailed information about each site is listed in Table 5.2.

The half-hourly fluxes were processed using standardized quality control procedures, which have been described in the references of each station. The half hourly \( H \), \( LE \), and four component radiation were then averaged to obtain monthly values. Monthly average values derived from less than 70% of the total data in that month were excluded to minimize uncertainty.
Chapter 5

Table 5.2 Flux sites used for product validation.

<table>
<thead>
<tr>
<th>Lat[deg]/Lon[deg]</th>
<th>Land cover</th>
<th>Eddy covariance</th>
<th>Radiometer</th>
<th>Measurement period</th>
<th>Elevation</th>
</tr>
</thead>
<tbody>
<tr>
<td>WJ 30.4200N/103.5000E</td>
<td>Crop</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Mar, 2008-Aug, 2009</td>
<td>539 m</td>
</tr>
<tr>
<td>MQ 33.8872N/102.1406E</td>
<td>Alpine meadow</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Apr, 2009-May, 2010</td>
<td>3439 m</td>
</tr>
<tr>
<td>AL 33.3905N/79.7035E</td>
<td>Bare soil</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Jul, 2010-Dec, 2010</td>
<td>4700 m</td>
</tr>
<tr>
<td>BJ 31.3686N/91.8986E</td>
<td>Alpine grass</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Jan, 2008-Dec, 2010</td>
<td>4520 m</td>
</tr>
<tr>
<td>MY 40.6038N/117.3233E</td>
<td>Orchard</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Jan, 2008-Dec, 2010</td>
<td>350 m</td>
</tr>
<tr>
<td>DX 39.6213N/116.4270E</td>
<td>Crop</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Jan, 2008-Dec, 2010</td>
<td>100 m</td>
</tr>
<tr>
<td>GT 36.5150N/115.1274E</td>
<td>Crop</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Jan, 2008-Dec, 2010</td>
<td>30 m</td>
</tr>
<tr>
<td>YC 36.9500N/116.600E</td>
<td>Crop</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Oct, 2002-Oct, 2004</td>
<td>13 m</td>
</tr>
<tr>
<td>DT 31.5169N/121.9717E</td>
<td>Wetland</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Jan, 2005-Dec, 2007</td>
<td>5 m</td>
</tr>
<tr>
<td>SC 35.955N/104.133E</td>
<td>Dry land</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Jan, 2007-Dec, 2008</td>
<td>1965 m</td>
</tr>
<tr>
<td>WS 36.6488N/116.0543E</td>
<td>Winter wheat</td>
<td>CSAT3, Licor7500</td>
<td>CNR-1</td>
<td>Jan, 2006-Dec, 2008</td>
<td>30 m</td>
</tr>
</tbody>
</table>

5.5.2 Evaluation results

In order to analyze the source of flux calculation errors, variables related to surface radiation fluxes were all validated with measurement. Table 5.3 shows that H and LE have a RMSE value not higher than 22 W/m² which is lower than that of other statistical method produced product (Table 7 in (Wang et al., 2007), Table 5 in (Jiménez et al., 2009)). Kalma et al. (2008) assessed 30 published LE validation results by using ground flux measurements and reported an average RMSE value of about 50 W/m² and relative errors of 15–30%. We also compared our validation results with that of other similar product produced by previous SEBS version, like Vinukollu et al. (2011b) produced a global land surface fluxes with a RMSE value of 40.5 W/m² (sensible flux) and 26.1 W/m² (latent flux) (calculated from Table 4 in (Vinukollu et al., 2011b)) which are bigger than that of this study. Table 5.3 also lists the statistical value for GLDAS (which has the highest spatial resolution than other available land energy flux dataset) based on the same China flux stations measurements in this study. According to the mean values of statistical variables, the quality of our
Development of a 10-year (2001-2010) high resolution land surface energy balance

The development of a 10-year (2001-2010) high resolution land surface energy balance flux dataset is comparable to GLDAS’ model and data assimilation results. These accuracy comparisons demonstrate that our revised model is a preeminent one for producing a high resolution China land surface energy flux dataset.

### Table 5.3 The comparisons between measurements and modeled fluxes

<table>
<thead>
<tr>
<th></th>
<th>Energy flux</th>
<th>Radiation flux</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>H (Wm⁻²)</td>
<td>LE (Wm⁻²)</td>
</tr>
<tr>
<td>This study</td>
<td>0.39</td>
<td>0.9</td>
</tr>
<tr>
<td>Intercept</td>
<td>-0.5</td>
<td>-6.1</td>
</tr>
<tr>
<td>RMSE</td>
<td>21.5</td>
<td>21.9</td>
</tr>
<tr>
<td>MB</td>
<td>14.7</td>
<td>10.1</td>
</tr>
<tr>
<td>R</td>
<td>0.41</td>
<td>0.85</td>
</tr>
<tr>
<td>Sample</td>
<td>280</td>
<td>284</td>
</tr>
<tr>
<td>GLDAS</td>
<td>Slope</td>
<td>0.77</td>
</tr>
<tr>
<td>Intercept</td>
<td>20.83</td>
<td>5.1</td>
</tr>
<tr>
<td>RMSE</td>
<td>26.6</td>
<td>20.6</td>
</tr>
<tr>
<td>MB</td>
<td>-15.8</td>
<td>0.75</td>
</tr>
<tr>
<td>R</td>
<td>0.46</td>
<td>0.80</td>
</tr>
<tr>
<td>Sample</td>
<td>249</td>
<td>250</td>
</tr>
</tbody>
</table>

Net radiation has the relative highest RMSE and MB values than other energy fluxes in the dataset, because its accuracy is dependent on the accuracy of other variable estimates (albedo, LST, SWD, LWD, LWU, etc.). Any errors in these variables can cause bias in the net radiation. LWD has a linear fitting slope value of 0.9, with most points located around the fitting line. The correlation coefficient is as high as 0.98. It demonstrates that there is still room for improvement of the LWD algorithms, as the LWD algorithms in ITPCAS were developed based on measurements over the Tibetan Plateau. The LWD algorithms may not be accurate for other parts of China (these contents are discussed with Dr. Kun Yang). This reminds that more accurate LWD radiation fluxes are needed in order to get more accurate turbulent fluxes and evapotranspiration.
Fig. 5.2 Time series comparison with SEBS input and output variables and its measurement at Yucheng station.
In addition to the statistical evaluation of model results against observations, seasonal and inter-annual changes in model results need to be checked. Yucheng station, which is an agricultural experimental station with winter wheat and summer maize as dominant crops, was taken as an example. Crops at Yucheng station mature twice per year, which is representative of warm, temperate farming cropland, typical for the Northern China Plain. Two years’ observations of monthly average fluxes and related variables were used to compare with values extracted from our model-derived product (Fig. 5.2). The inter-annual and seasonal LST and LWU data closely match the in-situ observations. The SWD term also successfully captures seasonal variations. LWD shows a similar trend, although it is systematically lower than observations. The LE produced at Yucheng station not only captures seasonal variations, but also responds consistently at step stages, for example when the wheat is harvested or maize seeds just sown (from June to August). The observed increased sensible heat and decreased latent heat flux in July 2003 is caused by the wheat harvest, however this signal change is not captured by the model result. The simulated sensible and latent heat produced by SEBS has one month lag than the reality. This phenomena should be caused by maximum monthly NDVI value which produces a failure representation of canopy status changes in June.

The Semi-Arid Climate and Environment Observatory of Lanzhou University (SACOL) is situated on the China Loess Plateau, at 1965.8 m above sea level. Annual mean precipitation there is 381.1 mm and annual evapotranspiration is 1528.5 mm (Huang et al., 2008). As a typical station under arid conditions, flux measurements from SACOL were compared with the time series of the grid point extracted from the model product (Fig. 5.3). The land surface around the station was covered by snow from 19 January to 20 February in 2008. The albedo has a consistent high value for February. It does show a relative low albedo for January, which could be caused by the coarse temporal sample of the satellite sensor. The calculated monthly sensible heat and latent heat in January of 2008 has a MB value of -11.7 (with an observed monthly mean value of 15 W/m²) and -7.6 W/m² (with an observed monthly mean value of 4.8 W/m²) separately. The relative large bias over snow covers of SC station may be caused by the mix pixel around the station.
Fig. 5.3 Time series comparison with SEBS input and output variables and its measurement at SACOL station.
The results of other stations were included in the supplementary materials. These comparisons show that the model estimates of surface energy balance variables adequately match the magnitude and seasonal variation observed at the stations in several contrasting eco-systems. Comparison of the flux tower and modeled fluxes at the station show that latent fluxes were more accurate than the sensible fluxes. The revised heat roughness parameterization method also performance better than the original version according to the flux measurement in this paper.

We also do comparisons of statistical values from other studies (Table 5.4). It shows that the accuracy of our dataset is one of the best among high resolution land surface fluxes dataset.

<table>
<thead>
<tr>
<th>Research area</th>
<th>Method</th>
<th>Statistical parameters</th>
<th>H (Wm⁻²)</th>
<th>LE (Wm⁻²)</th>
<th>Flux network</th>
<th>Note</th>
</tr>
</thead>
<tbody>
<tr>
<td>This study</td>
<td>China landmass</td>
<td>SEBS</td>
<td>RMSE</td>
<td>23.1</td>
<td>21.9</td>
<td>Flux towers in China</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>MB</td>
<td>16.8</td>
<td>8.3</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>R</td>
<td>0.6</td>
<td>0.8</td>
<td></td>
</tr>
<tr>
<td>Wang et al., 2007</td>
<td>Southern Great Plains of USA</td>
<td>Regression method</td>
<td>RMSE</td>
<td>×</td>
<td>29.8</td>
<td>Flux towers in Southern Great Plains of USA calculated from table 9 in the reference</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>MB</td>
<td>×</td>
<td>12.17</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>R</td>
<td>×</td>
<td>0.91</td>
<td></td>
</tr>
<tr>
<td>Jiménez et al., 2009</td>
<td>Globe</td>
<td>Statistical method</td>
<td>RMSE</td>
<td>×</td>
<td>×</td>
<td>AmeriFlux calculated from table 5, and 7 in the reference</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>MB</td>
<td>-5.23</td>
<td>7.9</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>R</td>
<td>0.68</td>
<td>0.76</td>
<td></td>
</tr>
<tr>
<td>Vinukollu et al., 2011b</td>
<td>Globe</td>
<td>SEBS</td>
<td>RMSE</td>
<td>40.5</td>
<td>26.1</td>
<td>AmeriFlux calculated from table 4 in the reference</td>
</tr>
</tbody>
</table>

5.6 Spatial distribution of land surface energy fluxes

Based on yearly average maps of land surface radiation and energy fluxes, we analyzed the spatial pattern of radiation and energy fluxes in China land and compared it with other product, like GLDAS. The highest values of downward surface solar radiation (Fig. 5.4a) was located in the south-west of the Tibetan Plateau, while the lowest occurred in the Sichuan Basin (SB). The highest level of upward short-wave radiation (Fig. 5.4c) occurred around the snow-covered peaks of the Himalaya (HM), Karakorum (KRM) and Kunlun (KLM), Qilian (QLM) and Nyainqentanglha (NQM) mountains. The strongest net solar
radiation (SWD minus SWU) on the Chinese land mass occurred in the southern part of the Tibetan Plateau (see supplementary materials). The downward and upward long-wave radiation (Fig. 5.4b and 5.4d) on the Tibetan Plateau was the lowest for all China landmass. Southern China has the highest levels of upward and downward long-wave radiation. The highest values of net long-wave radiation (LWU minus LWD) occurred on the southern and western parts of the Tibetan Plateau (see supplementary materials).

Figure 5.5a shows that northwestern China (NWC), the western Tibetan Plateau (TP), the inner Mongolian Plateau (MP) and the Loess Plateau (LP) have the highest yearly average values for surface sensible heat flux. Cropland of the north China Plain (NCP, including the lowlands of Shandong, Henan, and Hebei provinces) and the northeastern China Plain (NEP, including lowlands of Liaoning, Jilin, and Heilongjiang provinces) have low yearly average value for sensible heat flux. The Pearl River delta (PRD) and Tarim (TRB) and Sichuan (SCB) basins also have low levels of sensible heat flux, as do the Yinchuan (YCB) and the inner Mongolian basins (IMB) along the Yellow River. The spatial distribution is consistent to GLDAS result (see supplementary).

Fig. 5.4 Yearly average maps of (a) downward shortwave radiation (SWD), (b) downward longwave radiation (LWD), (c) upward shortwave radiation (SWU), (d) upward longwave radiation (LWU) from 2000 to 2010.
Simulated annual latent heat fluxes (Fig. 5.5b) exhibited a southeast to northwest decreasing gradient, which is consistent to other study (Liu et al., 2013b). The southeast of Tibetan Plateau have high annual latent heat fluxes. The Gobi desert in northwest of China (NWC), has the lowest annual latent heat flux, followed by the western Tibetan Plateau and the inner Mongolian Plateau (MP). Lake regions along the Yangtze River and basin region along Yellow River have relatively high levels of latent heat flux. The highest levels of yearly averaged surface net radiation (Fig. 5.5c) can be found around southwestern China and Lasha basin (LB). The lowest levels occur in the Sichuan (SCB) and Junggar basins (JB). The highest levels of yearly averaged ground heat flux (Fig. 5.5d) are to be found in the western China, which is due to large incoming solar radiation under the dry conditions. The monthly average of G0 is negligible compared with other fluxes. These spatial distribution maps of each factor in the surface energy balance give important background information and demonstrate the impact of plateau topography on energy, water and climate change (Yao et al., 2013).

Fig. 5.5 Yearly average maps of (a) sensible heat flux (H), (b) latent heat flux (LE), (c) net radiation (Rn), (d) ground heat flux (G0) from 2000 to 2010.
Figure 5.6 contains seasonal comparisons of $H$ between boreal winter (DJF), spring (MAM), summer (JJA) and autumn (SON). The largest area of positive sensible heating occurs in spring. Lee et al. (2011) shows that the sensible heat flux contrast between the Chinese land mass and the ocean during the pre-monsoon period affects the monsoon development in East Asia. Figure 5.6(a) shows that the sensible heating sources in spring are found to be around the Tibetan and other plateaus in China. Variations in sensible heating over these high and dry plateau bodies are likely to have important impacts on the summer monsoon (Boos and Kuang, 2010; Qiu, 2013; Wu et al., 2012). During summer, the highest sensible heat fluxes are found on the western Tibetan Plateau, the eastern Loess Plateau (LP) and the Northwestern China (NWC).

Fig. 5.6 Averaged maps of sensible heat fluxes for (a) March-May (MAM), (b) June-August (JJA), (c) September-November (SON), and (d) December-February (DJF) from 2000-2010.

LE in summer has the largest area of high latent heating, followed by spring, autumn and winter (Fig. 5.7). Southeastern and southern China have the highest levels of latent heat in summer, which is related to abundant rainfall in these regions. Irrigated land, such as that found in Yinchuan (YB), the inner Mongolian basin (IMB) and the downstream basins of the Tianshan (TM) and Kunlun (KLM) mountains, has a level of high latent heat and evapotranspiration
due to the ample water sources in summer. Latent heat fluxes in autumn and winter were significantly lower than that of the other two seasons. The magnitudes and spatial patterns of LE in China are generally consistent with other reports (Jung et al., 2010; Mu et al., 2007; Yao et al., 2013).

Net radiation in summer has the highest values among the four seasons. Most of the Chinese land mass acts as a surface energy source for atmosphere (Fig. 5.8). The energy balance predict a negative value of net radiation, but close to 0, which may be likely due to uncertainties in long-wave radiation from the reanalysis data, as well as inaccurate estimation of emissivity.
5.7 **Trend analysis**

Capturing the inter- and intra-annual variation for each land surface energy variables is interesting to the monsoon and climate change studies (Zhu et al., 2012). Besides, understanding these variations is essential for us to study the climate change and water resource related issues. We calculated annual average values of each flux. The nonparametric Mann-Kendall test (MK) is one of the most widely used methods for hydro-meteorological time series analysis (Gan, 1998; Liu et al., 2013c). The MK method was applied to the series of annual average fluxes to check variations during the last ten years. The resulting slope indicates that surface downward short-wave radiation increased from 2001 to 2010 over the majority of the Tibetan Plateau (Fig. 5.9). The upward short-wave radiation over peaks of the Himalaya (HM), the Gangdes (GM), the Karakorum (KRM), and the Qilian (QLM) and Nyainqentanglha (NQM) mountains increased over last 10 years period, which may be caused by the glacial retreat in these areas (Scherler et al., 2011; Yao et al., 2004). Lasha basin (LB) has the steepest rising trend in the LWU signal, which may be caused by relatively greater anthropogenic (e.g. urbanization) activity in this area. Downward long-wave radiation did not show any clear spatial pattern in the trend analysis. The
Development of a 10-year (2001-2010) high resolution land surface energy balance

net radiation over several high mountains ranges (including the Himalaya, the Gangdedise, the Karakorum and the Qilian and Nyainqentanglha mountains) increased by approximately 5 W/m² throughout the 10 years (Fig. 5.10). The central part of the Tibetan Plateau had the strongest increase in the net radiation. As pointed out by Matthew (2010), soil moisture in the central Tibetan Plateau showed an increasing trend from 1987 – 2008. Wetting soil can cause the ground surface to absorb more net radiation and increase the latent heat flux. Moreover, wetter soil can increase soil heating capacity (Guan et al., 2009) and further increase ground heat flux. The increased net radiation and soil moisture can also explain a rising trend in latent heat in the central Tibetan Plateau. The Tibetan Plateau is experiencing accelerated environmental change (Salama et al., 2012; Zhong et al., 2011). The land surface radiation and energy trend analysis also shows that the plateau is experiencing a relative stronger change in land surface radiation (verified by Tang et al. (2011)) and energy exchange than other parts of China.

Fig. 5.9 Spatial trends of (a) SWD (downward shortwave), (b) LWD (downward longwave), (c) SWU (upward shortwave), and (d) LWU (upward longwave radiation) in China land from 2001-2010.
5.8 Conclusion

In view of China’s highly fragmented landscape, high-resolution land surface heat flux maps are necessary for hydrological studies. China includes arid, semi-arid, humid, and semi-humid regions. Quantifying the water and energy budgets is a challenge. We have developed the surface energy balance system further to produce land surface heat flux dataset at a continental scale of higher resolution than other methods. In summary, a data product at monthly interval for land surface heat fluxes assessment has been developed using remote sensing data and surface meteorological information. We have validated the remote-sensing-based approach with in-situ observations from 11 flux stations in China. Taking into account the limitations of available spatial data and computing resources, we applied the model to the area of Chinese land mass using a 0.1 degree grid meteorological dataset, MODIS LST, vegetation indices and other variables to generate a climatological dataset of land surface energy balance for a 10-year period. The modeling results at both pixel point and spatial distribution demonstrate that this approach meets our aims in terms of being robust across a variety of land cover and climate types and performs well for the temporal and spatial scales of interest. These spatial distribution maps of
each terms of surface energy balance give important background information on terrestrial hydrological and energy cycle. This product also demonstrates the impact of topography and climate condition on land-air energy, water exchanges in China.

The applicability of remote-sensing-based estimates of land fluxes is hampered by limited temporal coverage of satellite sensors (Ryu et al., 2012). Remote sensing data are snapshots of the land surface status at a particular point in space and time (Ryu et al., 2011). It is a challenge to compare remote-sensing-based monthly flux data with ground measurements which are made on time scales ranging from half-hour to monthly scale. The energy flux product has a spatial resolution of approximately 10 km, while flux towers have a footprint of tens to hundreds meters. The tower footprint may not be representative of the larger pixel of the product, and this mismatch will result in errors if the mean of the satellite pixel is different from that of the tower flux footprint. Remote-sensing-based studies stress that direct comparison is a challenge because scale mismatch (Norman et al., 2003) and heterogeneity of the land surface reduces the spatial representativeness of ground-site measurements (Mi et al., 2006). Another challenge is validating the grid-box-based simulation results on the China land scale, since reliable observations of flux data are only available from a few sites in the simulated region.

Potential effects of changes in turbulent heat fluxes on the monsoon over East Asia (Lee et al., 2011) as a result of China’s recent urbanization can be studied further using this product. As an independently satellite-based product, it can also be used as a data source for evaluating land surface models. The product can be downloaded from the URL:

https://drive.google.com/folderview?id=0B7yGrB1U9eDec2JFbnA5eldlVHc&usp=sharing
Chapter 6 Deep atmospheric boundary layer over the Tibetan Plateau

6.1 Abstract

The greatest observed height of the atmospheric boundary layer (ABL) over the Tibetan Plateau, reaches altitudes as high as 5 km above the ground (~9.4 km above sea level). Its daily structure and evolution are investigated and compared for three different periods of the year; namely late winter, the onset of the monsoon, and the monsoon period itself. We find that the vertical profiles of several variables relevant to the ABL are quite distinct during the three periods, and that the daytime convective boundary layer (CBL) in February and March is much deeper than it is during the monsoon period. The dry condition of both soil and atmosphere in late winter does support a high CBL development. Also, the effect of the horizontal wind shear on the development of the CBL is important.

*This Chapter is based on:
Chen X., Skerlak, B., Añel JA, Ma Y., Su Z., Rotach, M.W. A record 9.4 km measurement of top of the atmospheric boundary layer over the Tibetan Plateau (in revision).
6.2 Introduction

The Tibetan Plateau is the largest and highest plateau on Earth, and it influences the atmospheric boundary layer (ABL) over it via thermal, dynamical and turbulent processes (Ma et al., 2002; Xu et al., 2008; Yang et al., 2004). The ABL processes are crucial for the exchange of momentum, water, and other substances between the plateau's surface and the troposphere. Good knowledge of the ABL above the plateau is vital to the understanding of the thermal and dynamical effects of the Tibetan Plateau on Asian weather and climate (Boos and Kuang, 2010), and for its correct representation in atmospheric flow models (Seibert et al., 2000). It is known that the ABL over the Tibetan Plateau influences the eastern China (Li and Gao, 2007; Stensrud, 1993; Xu and Cui, 2009), and that some synoptic systems in eastern China have their origins from the planetary boundary layer over the plateau (Gao et al., 1981; Tao and Ding, 1981). Chen et al. (2012a) showed that the Tibetan ABL can be connected with the exchange of air masses between the stratosphere and the troposphere. Even taking these few studies into consideration, the ABL over the Tibetan Plateau remains not well understood.

6.3 Data and method

As illustrated in Fig. 6.1, nine radiosonde observation sites are included in this thesis, which is supported by the Sino–Japan joint cooperation project (Xu et al., 2008). Four sites (Gerze, Lasha, Nagqu, and Litang) were situated on the Plateau. The height of other stations was lower than 2000 m ASL (above sea level), which will be discussed as low altitude observations.

Three sites Gerze, Litang, and Dali were equipped with Vaisala DigiCORA III GPS radiosonding system. The radiosondes used were Vaisala RS92 calibrated with local meteorological measurements before releasing. The sounding data contained profiles of temperature, pressure, relative humidity, horizontal wind speed and direction. Utilizing differential GPS theory, the receiver can automatically compute the height of the sensor and the wind speed. The radiosonde transmits data to the receiver every 2 seconds. With an average 5 m/s ascending speed, the obtained vertical resolution of the data is about 10 meters. The other six sites employed the Chinese meteorological radiosonde system (detailed information about the radiosonde can be founding from the following website: http://www.cwqx.com/). These systems can observe
temperature, pressure, humidity, wind speed and direction at vertical resolution of 100 meters.

Fig. 6.1 Distribution of radiosonde sites over the Tibetan Plateau (ASL-elevation above sea level).

Table 6.1 Intensive observation dates of the three periods in 2008. X means no observation data.

<table>
<thead>
<tr>
<th></th>
<th>Gerze</th>
<th>Lasha</th>
<th>Nagqu</th>
<th>Litang</th>
<th>Lijiang</th>
<th>Dali</th>
<th>Tengchong</th>
<th>Kunming</th>
<th>Mengzi</th>
</tr>
</thead>
</table>

Three intensive observation periods (IOP) were carried out. Detailed information about the observation dates are shown in Table 6.1. The first observation period (IOP1) was in winter time. The second observation period (IOP2) was conducted in the period of monsoon onset time. The third
observation period (IOP3) was almost in the mature phase of monsoon. Four radiosondes were released every day at 01:00, 07:00, 13:00 and 19:00 local standard time (LT).

To study the diurnal variations in the ABL, results of Gerze station (32.17° N, 84.03° E, 4415 m above sea level (ASL)) was presented here. According to the study of Seibert et al. (2000), this station is suitable for studying the ABL over the plateau because there are no high mountains nearby. Moreover, there are some existing surface, radiosonde and ABL data, obtained during three intensive observation periods (IOP). Near surface measurements were made of wind speed, wind direction, temperature, relative humidity at 0.5 and 10 m above ground, while pressure, precipitation, and the four radiation components (downward shortwave radiation (DSR), upward shortwave radiation (USR), downward longwave radiation (DLR), and upward longwave radiation (ULR)). From the two observations level, turbulent flux of sensible and latent heat flux were estimated by applying the bulk transfer method to the gradients difference (Yang et al., 2003). The universal profile form of Högström (1996) was used. The roughness length is given by the method of Yang et al. (2003).

The height of the top of the CBL was determined using the parcel method, which is known to ensure the reliable results even in unstable conditions (Hennemuth and Lammert, 2006; Holzworth, 1964; Seibert et al., 2000). The simple parcel method takes the top of CBL as the height of intersection of the actual potential temperature (θ) profile with the dry-adiabatic ascent, starting at the surface temperature. We used the Bulk Richardson number ($Ri_B$) to confirm the results of the parcel method and the depth of the convective boundary layer. The depth of the boundary layer is defined as the height at which $Ri_B$ reaches a critical value, typically 0.25. $Ri_B$ is given by the following equation (Stull, 1988):

$$Ri_B = \frac{g z (\theta(z) - \theta(s))}{\theta(s) u(z)^2 + v(z)^2}$$

(6.1)

where $\theta(z)$, and $\theta(s)$ denote $\theta$ at a given height 'z' and the surface, respectively, and $u(z)$ and $v(z)$ are the zonal and meridional velocity component. $g$ is gravitational acceleration, 9.8 m/s². We compute $Ri_B$ based on
surface information from the lowest level of radiosonde data. The first sounding level is fixed at 4415 meters amsl before releasing radiosondes. Here we take the meteorological information at the first level measured by the radiosonde as that of surface.

6.4 **Diurnal and seasonal variation of the atmospheric boundary layer**

We chose the 25th of February as a demo day in IOP1 to analyse the diurnal variations in the structure of a very high CBL (Fig. 6.2). The corresponding surface energy fluxes are shown in Figure 6.3; a surface-based inversion was formed by 07:00, and had disappeared by 13:00 (Fig. 6.2a). In the sounding at 07:00 we observed a 500 m-deep cool and stable layer near the surface, and a residual layer above it with a depth of 1700 m and nearly constant $\theta$ (Fig. 6.2b). After sunrise at approximately 09:00, the net radiation (Rn) increased very sharply, resulting in surface heating. As a consequence, by 13:00 a mixed layer (ML) had grown to 3600 m, eroding both the surface and residual layers, and by 19:00 this growing layer was 4780 m deep (9195 m ASL at the site). During this fast development of the ML, the water vapour content (WV) and wind speed were well mixed by turbulence (Fig. 6.2c and 2d). The growth of the CBL favoured the entrainment of relatively dry air from above, driving a decrease in the average WV content of the ML from 01:00 to 19:00. The WV content in the ML also decreased slightly with increases in height, due to surface evaporation. We therefore conclude that for the studied day the development of the CBL was primarily caused by underlying surface heating.

It was observed that above 9300 m ASL the wind speed increased very rapidly and WV dropped quickly at 19:00. These high-gradient layers could be taken as the top of the CBL (well-mixed layers existed below this height), or as an indication of the upper troposphere and the lowermost stratosphere (UTLS). As shown in the picture, the top of the CBL at 19:00 was at a height of 9195 m ASL, and the tropopause was at 9455 m ASL.
Fig. 6.2 Profiles for the temperature (a), potential temperature (b), wind speed (c) and water vapour mixing ratio (d) at 01:00 (dark line), 07:00 (red line), 13:00 (blue line), and 19:00 (magenta line) on 25 Feb. Horizontal solid lines show the heights of the tropopause, and the horizontal dashed lines show the tops of the convective ABL. The stable layer is denoted as SL, the residual layer as RL, and the mixed layer as ML.
The profiles of $\theta_v$ in successive days clearly showed diurnal variations in the ABL height during IOP1. The ABL was fully developed in generally fair-weather conditions during IOP1. By mid-morning, the ML growth had eroded the nocturnal boundary layer, overtaking the remains of the previous day’s ML. The MLs at 13:00 and 19:00 could clearly be seen in the near-uniform and well mixed layers of $\theta_v$ and WV. In daytime, a strong super-adiabatic zone underlying the ML and above the ground was captured by the profiles of $\theta_v$ at 13:00 and 19:00, which suggested that the ABL was experiencing convective conditions. The top of the CBL could be differentiated by an abrupt increase in $\theta_v$, which capped a deep ML (Stull, 1988). The $\theta_v$ in the ML was uniform at 19:00. During IOP1, large gradients of water vapour density occurred at the top of the ABL on 25 February, 27 February, 29 February, 1 March, 2 March, 4 March, and 5 March, which made the gradients suitable for defining the height of the ABL on these days.
The soundings taken at 01:00 and 07:00 (night) during IOP1 often exhibited a cool layer near the surface and a very deep residual layer. The profile at 19:00 on 4 March showed a 5 km CBL, followed by a 4.5 km residual layer in the morning of 5 March.
During IOP2 and IOP3, the $\theta_v$ in the ML was also ‘well mixed’ due to turbulent mixing, but the depth of the CBL was smaller than that in IOP1. A significant seasonal difference in the CBL height was reported in other regions (Cuesta et al., 2008; Seidel et al., 2010; Seidel et al., 2012; Zhang et al., 2011). We also show the variations in the height of the CBL in Figure 6.4. The CBL heights computed using the two methods showed a consistent pattern of variation. The results from the parcel method are used below. Typical late afternoon mixed layer depths ranged from approximately 1-5 km in IOP1, 0.5-4.7 km in IOP2, and 1-3 km in IOP3. The statistical results for the CBL height are listed in Table 6.1. The average CBL height at 13:00 and 19:00 during IOP1 was higher than 2500 m. The mean value for IOP3 at 13:00 and 19:00 was not more than 1500 m.

Using large eddy simulations, Yang et al. (2004) simulated the daytime evolution of the high ABL. The development of the ABL is driven by surface heating. To explain the significant differences in the mixing depth, the surface heating was analysed. The sensible heat flux dominated the surface heating in IOP1. The sensible heat flux for IOP1 was the highest among the three IOPs (Fig. 6.5). The energy budget in the ABL indicated that the sensible heating was the dominant energy for sustaining ABL growth during IOP1. The intense dry thermal convection originating from the heated surface was responsible for the deep ML. After the onset of the monsoon, the latent heat flux became comparable with the sensible heat flux. Most of the net radiation ($R_n$) was used to evaporate moisture rather than to heat the surface, and $R_n$ therefore contributed little to the buoyant forcing. The $w$ in the ABL during IOP1 was not more than 1 g/kg, far lower than the 7.5 g/kg during IOP3. The atmosphere over the plateau became wetter after the onset of monsoon, and wet convection gradually became prevalent. The stronger dry convective activity developed by the sensible heat flux during IOP1 explains why the ABL was higher than in the other periods.

| Table 6.2 Statistical analysis of the CBL height at 01:00, 07:00, 13:00, and 19:00. |
|---------------------------------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| IOP1  | IOP2  | IOP3  |
| Hour  | 01:00 | 07:00 | 13:00 | 19:00 | 01:00 | 07:00 | 13:00 | 19:00 | 01:00 | 07:00 | 13:00 | 19:00 |
| Mean  | 14 m  | 0 m   | 2572 m| 3040 m| 158 m | 15 m  | 1643 m| 2203 m| 20 m  | 24 m  | 1472 m| 1168 m|
| Skewness | 3     | 4     | -0.5  | -0.6  | 5     | 3     | -0.3  | -0.1  | 1.6   | 1.8   | 0.7   | 0.9   |
The relationships between ABL depth and ground surface status were depicted in figure 6.6 and 6.7. All the IOPs show that the high ABL corresponded to high surface solar radiation. Higher CBL has a corresponding higher sensible heating (Fig. 6.6b), and a lower latent heat. IOP1 has a potential possibility to have a higher ABL than IOP2 and IOP3, when forced by a same sensible heating. The same rule does exist for IOP2, which has a higher potential than IOP3. The relationships between CBL height and surface shear stress have different performance among the three IOPs. IOP1 has a positive responding between CBL height and ustar, while IOP3 has a negative relationship between the two values. The scatter of IOP2 has no clear performance. We therefore conclude that the development of the CBL was caused by underlying surface heating and the momentum contribution to the high CBL development is also important.

The same land surface temperature for IOP1 can support much higher CBL height than IOP2 and IOP3 (Fig. 6.7a). The soil moisture (SM) in IOP1 is not
higher than 5%. A high soil moisture suppressed upward development of CBL due to its role on the surface energy departing between sensible heat and latent heat. Although IOP2 and IOP3 have a higher land surface temperature (LST) value, the difference between LST and air temperature in IOP1 has a comparable value with other two periods. The specific humidity shows a negative relation with CBL height in all the IOPs. More wet air suppresses the up-development of CBL. This also explains why same amount of sensible heating can cause a different CBL depth during the three periods (Fig. 6.7b). The dry condition of both soil and atmosphere in IOP1 does support a high CBL development.

Fig. 6.6 Relation between mean daily (09:00 to 13:00 and 19:00 when radiosonde were released) surface downward shortwave radiation (a), sensible heat flux (b), latent heat flux (c), ustar (d) and CBL height observed at 13:00 and 19:00 in the three IOPs
It is known that the development of the boundary layer is related not only with variations in the surface heating, but also with the atmospheric stability of the whole troposphere. Atmospheric stability is proposed in some cases to be the most influential variable controlling the development of the ABL (Santanello et al., 2005). In our analysis of the observational data, we found that the deep development of ABL is connected to the initial thermal structure of the troposphere. The stability is used to represent the thermal structure. The variations in the stability of the troposphere during IOP1, IOP2 and IOP3 are shown by the lapse rate in Figure 6.8. During both day and night, the lapse rates in the air below the top of the daytime CBL were nearly equal to—or higher than—the dry adiabatic lapse rate during IOP1 (Fig. 6.8a). In this case, a parcel of air can gain buoyancy and rise up when it is heated by the surface. The air in the CBL is unstable. While the lapse rate during IOP3 lay between the moist adiabatic lapse rates (a typical value was approximately 5 K/km) and the dry adiabatic lapse rates, the air was conditionally unstable. The stability of the air indicated that the atmospheric stratification of the plateau during IOP1 was more favourable for ABL growth, and that the ABL had the potential to grow much higher than it could during IOP2 or IOP3. Turbulence caused the daytime CBL to be well mixed and to have an adiabatic lapse rate. The establishment of a dry adiabatic lapse rate in the daytime CBL profiles showed that it was in fact a true CBL.

It is important to not only show the profiles of static (lapse rate) stability but also profiles of the dynamic stability. Generally a Richardson number $<0.25$ is taken to be a condition for turbulence associated with the dynamical instability of the atmosphere (Nath et al., 2010). The smaller the value of Richardson number, the less the stability of the flow and the greater the likelihood of turbulence. If the Richardson number is negative, then the region becomes convectively unstable. Here we use the bulk Richardson number to analyse the specific characteristics of the atmosphere’s vertical structure. Figure 6.9 depicts the variations of $Ri_B$ in the three IOPs. All the values of $Ri_B$ in layers below the tops of convective boundary layer (green and blue solid horizontal line) are negative values. It indicates the air was dynamic unstable and once the turbulent generated from surface heating, it will develop upward and makes the ABL convectively unstable. $Ri_B$ in the upper troposphere exhibits a significant discrepancy during the three IOPs. Most values of Rib in the upper troposphere during IOP1 is not higher than 5.
It is pointed out that the atmosphere over the TP is heated not only by diabatic heating from the surface, but also by the adiabatic subsidence in the upper troposphere (Tamura et al. 2010). From the different characteristics of upper troposphere during the three IOPs, especially evidenced by static and dynamic stability, heat transferred by the subsidence may also play a role. The intrusion air from stratosphere caused by tropopause folds can be one contribution of subsidence.
Fig. 6.8 Lapse rate of (a) IOP1, (b) IOP2 and (c) IOP3, where the coloured horizontal solid line shows the top of the mixed layer, the horizontal dashed line gives the height of the tropopause, the dashed vertical lines show the dry adiabatic lapse rate of 9.76 K/km. 01:00 (dark line), 07:00 (red line), 13:00 (blue line), 19:00 (green line). Y-xias is the elevation above sea level (unit km).
Fig. 6.9. Profiles of Richardson number in (a) IOP1, (b) IOP2 and (c) IOP3, where the coloured horizontal solid line shows the top of the mixed layer, the horizontal dashed line gives the height of the tropopause, the dashed vertical lines show $R_b=0.25$ and 1, respectively. 01:00 (dark line), 07:00 (red line), 13:00 (blue line), 19:00 (green line)
6.5 Discussion

Based on high resolution radiosonde observations, we discovered a boundary layer that was higher than any revealed by previous research (at least to our knowledge). The maximum ABL depth during IOP1 was as large as 5 km above ground. Yang et al. (2004) reported a 3 km ABL in June in the Naqu region, Li et al. (2006) reported that the highest ABL in the Qomolangma region could be as high as 3.8 km in April and May, Zuo et al. (2005) observed a 3.5 km height for the ABL in dry season in the Anduo region, and Zhang et al. (2003) found a 2.25 km ABL at Damxung. The Gerze region is in the western plateau. This could show that the ABL in the west is the highest in the plateau region. A high CBL was reported in other dry lowland regions (Cuesta et al., 2008; Gamo, 1996; Han et al., 2011); for example, Takemi et al. (2006) demonstrated a 4 km CBL over the Gobi desert in China, and Raman et al. (1990) reported a 4.7 km ABL over New Delhi in India on one day during the pre-monsoon period. Cuesta et al. (2008) analysed a 6 km Saharan atmospheric boundary layer. Results of previous studies showed that mixed layers develop to high altitudes because the daytime mixed layer becomes linked with a weakly stratified, near-neutral layer above it (Gamo, 1996; Han et al., 2011). Our observations show that a weakly stratified near-neutral layer often exists from 5 to 9 km ASL on the Plateau in the winter season, when surface cooling helps to isolate the upper ML from the surface and maintain a thick residual layer. The deep residual layer is eroded by the development of the CBL, becoming part of the deep boundary layer. The mixed layer over the Plateau can also develop to high altitudes to some extent because of the thick residual layer. Because of the low air density and the intense solar radiation, the CBL has the potential to grow much higher over the Plateau than it does over lowland areas (Yang et al., 2004).

The seasonal variation in CBL height has been studied in other regions. Holzworth (1967) pointed out that the highest mixing depth of the ABL in New York occurs in the period between April and August, and the lowest mixing depth occurred during winter. Cuesta et al. (2008) reported the greatest CBL height during the summer over the Sahara desert, which is different with the phenomena over the Tibetan Plateau. Thermodynamic sounding profiles suggest that the large sensible heat flux from the surface of the Tibetan Plateau during winter daytime was responsible for rising of the turbulence in the deep mixing layer to the upper troposphere. The sensible heat flux plays the main role in winter and the latent heat flux plays the main role in summer and autumn (Ma et
The ABL undergoes seasonal development in response to the surface heating. The sensible heat fluxes from the ground are the major contributors to the heating of the ABL. Due to convection and vigorous vertical mixing caused by dry heat at the surface of the plateau, a dry-adiabatic lapse rate becomes established in the high CBL. A deep residual layer is apparent during winter nights. With the development of the plateau monsoon, the night-time deep residual layer disappears, and the height of the mixed layer falls until it is similar to those observed in lowland areas.

A clearly well mixed potential temperature and water vapour layer can be identified when westerly winds dominate over the plateau. The mechanical turbulence caused by the wind shear of the dominating westerly winds over the plateau also contributes to the mixing of $\theta_v$, and clearly well mixed $\theta_v$ and WV layers can indeed be seen in our observations. All these factors greatly increase the height of the boundary layer.

A deep CBL may imply larger-scale thermals, and the existence of eddies within the CBL (Stull, 1988). Unusually high CBLs of 5 km have been reported in desert and arid regions, but the elevation of the land surface in these regions is typically not more than 1 km ASL. Due to the high elevation of the land in the Tibetan Plateau, the transport process in the UTLS can be connected with the ABL more easily (Chen et al., 2013a) than over lowland areas. A close surface-troposphere-stratospheric coupling system might exist over the Plateau. For example, it has been pointed out that the stratosphere-to-troposphere ozone flux into the ABL in spring is greatest over the Tibetan Plateau and the Rocky Mountains. Skerlak et al. (2012) and Chen et al. (2012a) concluded that the Tibetan Plateau is one of the three key source regions for transport from the boundary layer to the tropopause in the Asian monsoon region. Also, the Tibetan Plateau is a region of maximum multiple tropopause events and tropopause foldings (Chen et al., 2011a; Randel et al., 2007a), which can cause stratospheric air to be transported downward to the ABL via different processes (Johnson and Viezee, 1981). According to the results presented here, the Tibetan Plateau might favour an ABL that is deeper than usual. In the future, it would be useful to use ABL growth models and simulations and compare them with the observed data, to quantify the contribution of the different processes to these kinds of phenomena.
Deep atmospheric boundary layer over the Tibetan Plateau
Chapter 7 The behaviour of the tropopause folding events over the Tibetan Plateau

7.1 Abstract

This chapter analyses the structure of upper troposphere and lower stratosphere (UTLS), and provides observational evidence of stratosphere and troposphere exchange (STE) over the TP. Due to sparseness of high resolution radiosonde data, many previous studies assumed that there was only one thermal tropopause over the TP. Actually, the radiosonde temperature profiles in winter time over the TP often exhibit a multiple tropopause (MT). The MT occurs in winter time with high frequency over the Plateau. MT events during this time are found to be associated with tropopause folds near the subtropical westerly jet. The MT consistently varied with the movement of the jet. The MT becomes a single tropopause with the mature of the monsoon.

Earlier analyses of global MT events resulted in a climatic frequency of MT occurrences in winter season over the Plateau is not more than 40%. Based on high resolution data of intensive radiosonde observations, our estimations of MT occurrence over the Plateau can be as high as 80% during some winter time. This reminds us more attentions should be paid to the MT events above the Plateau. The stratospheric intruding episodes are generally associated with the presence of subtropical jet stream over the Plateau. The complex structure of dynamic tropopause folds over the Plateau have been reflected by the thermal MT events observed by radiosondes. The intrusion of air masses from the stratosphere may contribute to a higher upper tropospheric ozone concentration in winter than in summer above the plateau.

* This Chapter is based on:

The behaviour of the tropopause folding events over the Tibetan Plateau

7.2 Introduction

Due to the low air density and the strong solar radiation, the TP allows more energy from the surface directly heating the middle troposphere. This heat can influences the upper troposphere and warms the tropopause region (Fu et al., 2006). The exchange between upper troposphere and lower stratosphere (UTLS) was found over the Tibetan Plateau (Cong et al., 2002; Fu et al., 2006; Sprenger et al., 2003; Steinwagner et al., 2007; Zhan and Li, 2008). As a huge elevated heating source, a rigorous deep convection area (Yang et al., 2004), as a dynamical pumping and sucking unit (Duan and Wu, 2005) and westerly jet above the Plateau, the TP generates an active stratosphere and troposphere exchange (STE) region, forming a short-circuit and pathway for the STE (Fu et al., 2006). An improved understanding of the STE depends on our ability to quantify the UTLS structure and its variability (Stohl et al., 2003). Due to a lack of high resolution observational data, the structure of the UTLS above the Tibetan Plateau is still insufficiently understood (CCMVal, 2010; Gettelman et al., 2010; Hegglin et al., 2010).

The occurrence of multiple tropopauses (MT) plays a central role in the structure and features of the tropopause layer which plays a crucial role in the exchange between the troposphere and stratosphere above the TP (Añel et al., 2008). Both Añel et al. (2008) (employing Integrated Global Radiosonde Archive database) and Randel et al. (2007) (based on GPS radio occultation measurements and ERA-40) have derived global statistics of MT. The present paper presents statistics of MT and STE events above the Tibetan Plateau based on high resolution sonde data for the first time. Therefore, it will contribute to MT statistics of Añel et al. (2008) and Randel et al. (2007) and helps understanding the exchanges between the stratosphere and troposphere over the Tibetan Plateau.

Tropopause folds are the key feature and favorable structures for cross-tropopause exchange in the subtropics (Shapiro, 1980; Sprenger et al., 2003); The frequency of tropopause folds is highest in the subtropics related to the prevailing subtropical upper level jet stream (Reed, 1955; Schmidt et al., 2005). The Tibetan Plateau is situated at 28°N-38°N, a region which is significantly influenced by the subtropical jet stream as it moves northward from winter to summer. Sprenger et al. (2003) have analyzed frequency and global distribution of tropopause folds. Their results demonstrate that the global frequency of
shallow folds above the Plateau is highest during winter (see figure 3 in Sprenger et al., 2003).

Meanwhile, satellite observations of ozone have shown ‘Ozone Mini-Hole’ events and ozone valley phenomena over the Plateau (Bian, 2009; Tobo et al., 2008). The mechanisms responsible for the low total ozone have been discussed. Previous studies suggest that transport of tropospheric air with low ozone concentration across the tropopause influence the low ozone concentration in the lower stratosphere in summer (Zhou et al., 1995). Tian et al. (2008) found that the variability of the total column ozone over the TP is closely related to uplift and descent of isentropic surfaces. Tobo et al. (2008) observed ozone anomalies near the tropopause (150-70 hPa) having a large contribution to the low total ozone. Intrusions of stratospheric air masses with high ozone concentration into the troposphere were closely associated with tropopause folds (Beekmann et al., 1997; Reed, 1955; Sprenger et al., 2003). Due to high frequency of tropopause folds (Sprenger et al., 2003), stratosphere intrusions should frequently happen over the Plateau. These downward transports of rich ozone air strongly influence the vertical ozone distribution especially at UTLS area over the Plateau. However, the variation of the stratospheric intrusions and their influence on the vertical ozone distribution need further analyses. There is a lack of high resolution information of the tropopause above the Plateau. This study will contribute to characteristics of MT events using the most advanced radiosonde data in this area.

7.3 Observation data and Tropopause definition

The radiosonde data used in this Chapter was given in section 6.3 in Chapter 6. It will not be repeated in this chapter.

For computing the tropopause height, we employ the WMO lapse rate tropopause (LRT) definition (WMO, 1957):

a) The first tropopause is defined as the lowest level at which the lapse rate decreases to 2 K/km or less, provided also the average lapse rate between this level and all higher levels within 2 km does not exceed 2 K/km.

b) If above the first tropopause the average lapse rate between any level and all higher levels within 1 km exceeds 3 K/km, then a second tropopause is defined by the same criterion as (a).
The behaviour of the tropopause folding events over the Tibetan Plateau

The first tropopause is denoted by LRT1, and if a further tropopause is present above LRT1, it is named LRT2 and then LRT3. Double tropopause (DT) is denoted by detection of LRT1 and LRT2. Triple tropopause (TT) is identified as profile detected with LRT1, LRT2 and LRT3.

7.4 Multi-tropopause

In previous studies little attention was paid to MT events over the Plateau. During the intensive observation period, a frequent strong thermal inversion layer presents around 10 km above Gerze station (32.09° N, 84.25° E) situated at western Plateau. MT is calculated for the IOP1 period using the MT definition of WMO (1957) and it is proved that MT often happened during IOP1. Examples of the DT temperature profiles from Gerze station are shown in Fig. 7.1. Continuous radiosonde data for 25 Feb. 2008, 01:00 LT to 26 Feb. 2008, 13:00 LT are exhibited. The profiles show a tropopause identified around 300 hPa, and a second tropopause detected near 100 hPa. Figure 7.2 reveals profiles from 29 Feb. 2008, 01:00 to 1 Mar. 2008, 13:00 LT with only one tropopause above 100 hPa. We will explain what causes the difference in UTLS structure in these two time periods in Sect 7.5.

Fig. 7.1 Temperature profiles from 25 Feb. 2008, 01:00 to 26 Feb. 2008, 13:00 LT at Gerze site.
Employing the WMO LRT definition to all the radiosonde stations (shown in Fig. 6.1), we computed MT frequencies of the three IOPs, and listed them in Table 7.1. The frequency of MT events is given as a percentage with respect to the number of LRT1 events. The MT occurrences of the three Plateau stations (Gerze, Nagqu, and Litang) for IOP1 are as high as from 72.5% to 84%. Our analyses show that the frequency and seasonal variation of MT events are high during winter. The frequency of MT (discussed with Randel) derived from the GPS occultation data and the ERA-40 data during DJF (December, January and February) in Randel et al. (2007) on the Plateau was not more than 40% in their Fig. 9a and Fig. A1b. Añel et al. (2008) analyzed global MT events performed on the Integrated Global Radiosonde Archive database (IGRA). As shown in their Fig. 2, the percentage of DT occurrence had less seasonal variation and the value of DJF is not more than 20% above the Plateau. Thus more attentions should be paid to the MT events above the Plateau and the influence of the low vertical resolution of GPS and ERA-40 data on the estimation of MT events.

At the Dali site, which has an elevation of 1960 m and locates out of the Plateau, the MT occurrence of IOP1 is about 12.9%. This value is much lower than that
of other stations. We also picked out the radiosondes of the same date in each IOP. The statistics suggest that when moving to southern and low areas, the frequency of MT becomes lower. The MT frequencies of all the Plateau stations during IOP3 are the lowest comparing to other two periods. This result supports the conclusion of MT occurs in winter time with high frequency over the Plateau, which is similar to previous climatological studies.

Table 7.1 Frequency of double tropopause (DT) and triple tropopause (TT) during the three observation periods (%). X means no observation data.

<table>
<thead>
<tr>
<th></th>
<th>Gerze</th>
<th>Lasha</th>
<th>Nagqu</th>
<th>Litang</th>
<th>Lijiang</th>
<th>Dali</th>
<th>Tengchong</th>
<th>Kunming</th>
<th>Mengzi</th>
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</thead>
<tbody>
<tr>
<td>IOP1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DT</td>
<td>44</td>
<td>40</td>
<td>X</td>
<td>X</td>
<td>51</td>
<td>27.6</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>TT</td>
<td>51.5</td>
<td>27.6</td>
<td>X</td>
<td>X</td>
<td>12.9</td>
<td>0</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>IOP2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>DT</td>
<td>11</td>
<td>1</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>2.5</td>
<td>0</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>TT</td>
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<td>0</td>
<td>X</td>
<td>X</td>
<td>7.5</td>
<td>X</td>
<td>0</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>IOP3</td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>DT</td>
<td>0</td>
<td>0</td>
<td>2.5</td>
<td>0</td>
<td>X</td>
<td>X</td>
<td>2.6</td>
<td>0</td>
<td>2.7</td>
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<tr>
<td>TT</td>
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<td>0</td>
<td>4.2</td>
<td>0</td>
<td>X</td>
<td>X</td>
<td>2.5</td>
<td>0</td>
<td>2.7</td>
</tr>
</tbody>
</table>

Fig. 7.3 Distribution of multi-tropopause heights at different phase of the monsoon. Table 7.1 lists which sites and data are analyzed in these three IOP.

Figure 7.3 shows the statistical distributions of MT height during the three IOPs. The statistics for IOP1 demonstrate an overall bimodal distribution with maxima near 10 km (primarily associated with LRT1) and 17 km (firstly contributed by LRT2). According to the significant difference in the two
Chapter 7

tropopause height, we mark them as ‘low’ and ‘high’ tropopause separately. LRT1 is more inclined to low tropopause during this time. The majority of LRT2 height in IOP1 is around 17 km, which is more close to high tropopause of equatorial area. The statistics of IOP2 and IOP3 in Figure 7.3 show a single maximum (primarily LRT1) centered at 17 km, which means MT is rarely observed during the monsoon season.

The average heights of DT during the three periods are listed in Table 7.2. LRT1 of the three Plateau stations (Gerze, Nagqu and Litang) have heights around 11-13 km during IOP1. These tropopauses can be treated as low tropopauses. For all observation periods and sites LRT2 is characterized by a height around 17 km. After the monsoon onset, the LRT1 height over the Plateau was elevated with 4-5 km, indicating the disappearance of the low tropopause. This can be explained as a result of both the Plateau’s thermal forcing, which cause large-scale ascent flow and vertical convection, and the poleward movement of the jet stream. Tian et al. (2008) suggested the thermal sink (source), by forcing a descent (ascent) flow, lowers (lifts) the tropopause. The thermal dynamic effects of the Plateau can be one reason of seasonal variation of the tropopause, but does not seem probable to explain variations from 10-17 km in the LRT1 height during winter time. Our explanation will be presented in the next part.

Table 7.2 Average height (unit of km) of LRT during the three observation periods. X means no observation data.

<table>
<thead>
<tr>
<th></th>
<th>Gerze</th>
<th>Lasha</th>
<th>Nagqu</th>
<th>Litang</th>
<th>Lijiang</th>
<th>Dali</th>
<th>Tengchong</th>
<th>Kunming</th>
<th>Mengzi</th>
</tr>
</thead>
<tbody>
<tr>
<td>LRT1</td>
<td>1.2</td>
<td>1.6</td>
<td>X</td>
<td>X</td>
<td>1.2</td>
<td>1.6</td>
<td>X</td>
<td>X</td>
<td>X X X X X X</td>
</tr>
<tr>
<td>LRT2</td>
<td>1.6</td>
<td>1.7</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>1.7</td>
<td>X</td>
<td>X</td>
<td>X X X X X X</td>
</tr>
<tr>
<td>IOP1</td>
<td>1.7</td>
<td>X</td>
<td>1.6</td>
<td>1.7</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.8</td>
<td>1.7 1.8 1.7 1.6 1.6</td>
</tr>
<tr>
<td>IOP2</td>
<td>1.6</td>
<td>1.7</td>
<td>X</td>
<td>X</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.8</td>
<td>1.7 1.8 1.7 1.6 1.6</td>
</tr>
</tbody>
</table>

7.5 The tropopause folds over the Tibetan Plateau

It is well known that altitude variation of tropopause in the extratropics has a close relationship with the local synoptic situation. In order to analyze the meteorological situation associated with the single and double tropopauses, we use the ERA interim reanalysis data provided by ECMWF, the European Centre for Medium–Range Weather Forecasts (Uppala et al., 2005). We take 25 Feb. 12:00 LT (Fig. 7.4a) and 29 Feb. 2008, 12:00 LT (Fig. 7.4b) as an example in first. During these two times, DT and single tropopause were observed at Gerze. Figure 7.4 includes potential vorticity (PV) isolines (units: 1 PVU), zonal wind
and potential temperature derived from ERA interim data. Inverted triangles show simultaneous tropopause heights observed at the two stations of Gerze and Nagqu. In order to exactly locate their positions of thermal tropopauses, we interpolated the ERA interim 1.5 latitudinal resolution to 0.1°, and pressure levels lower than 400 hPa to 10 hPa, higher than 400 hPa to 25 hPa vertical resolution with spline method. Due to the westerly wind, the soundings were usually moved to east with relative small north-south variation. The positions of MT observed by simultaneous radiosondes were marked by the same latitude as its station point in Fig. 7.4.

A method to identify the dynamical tropopause is based on potential vorticity (PV). A series of PV isolines (PV = 1–5 PVU) were plotted to indicate the approximate location of the dynamical tropopause (blue lines in Fig. 7.4). A principal indicator of intrusion of stratospheric air into troposphere is the occurrence of anomalous high PV values reaching down towards middle troposphere. It is because PV is generally greater in the stratosphere than in the troposphere (Holton, 2004). Fig. 7.4a shows that the westerly jet stream is at a latitude of 30°N above the TP. The observed maximum wind speed is higher than 60 m/s at 160 hPa level. The PV isolines were distorted in the vicinity of the northern edge of the subtropical jet stream. The 2 PVU isosurface reaches down to 380 hPa over the Plateau. A folds around 30°N is identified on isentropic surfaces penetrating deeply into the troposphere in Fig. 7.4a. The dynamical tropopause (identified by 1–5 PVU) exhibit a medium folded structure (Sprenger et al., 2003) over the Plateau which indicates a strong stratospheric intrusion. The entrainment of stratospheric air within the folds is marked both by high PV values and high ozone concentration (supported by Sciamachy ozone data). The ozone profile around 30°N shows the intrusion caused by the tropopause fold. The ERA ozone distributions around folds were found coincident with that of PV (see Fig. 7.5). Figure 7.5 shows the dynamical effects of wind on 25 February 2008. The tongue of high ozone air were transported from polar stratosphere to middle troposphere above the Plateau by southern downward meridional wind on the polarward side of the jet streams as shown in Fig. 7.4a. The time series of the same pictures as Fig. 7.5 from 25 February 2008, 01:00 LT to 26 February 2008, 13:00 LT show the development of the folds into deep and shallow events (supplementary figure), propagating from north to south. The deepest folds were classified as medium folds during this time (Sprenger et al., 2003). Because of the high elevation of the Plateau, the intrusions can easily transport stratospheric air mass to its ground.
Fig. 7.4 (a) Meridional cross-section at 84.25° E (over Gerze site) on 25 Feb. 2008, 12:00 LT between 10° and 50° N in latitude and between 1000 hPa and 20 hPa in the vertical derived from ERA interim data, including zonal winds (black contours, m/s), potential vorticity (PV) (blue lines, contours of 1–5 PVU units), and potential temperature (red contours, k). The red and black triangles were tropopauses observed at Gerze and Nagqu station.
The behaviour of the tropopause folding events over the Tibetan Plateau

Fig. 7.4 (b) same as Fig. 7.4 (a) but for 29 Feb. 2008, 12:00 LT.

A similar meteorological situation on 29 Feb. 2008, 12:00 LT (Fig. 7.4b) was further used to study single tropopauses. The second tropopause at 70 hPa observed by radiosondes in Fig. 7.4b should be a stratospheric layer, not real
tropopause. Thus, the profile at this time is taken as a single tropopause event. The jet core (defined as the centre of wind speed higher than 50 m/s) moved northward to 33°N, on the north of Gerze and Nagqu site. The shallow folded structure (Sprenger et al., 2003) at this time retreated to north of the Plateau. The equatorial tropopause extends over the South Plateau. The dynamical tropopause indicates no folds above the two stations. The radiosonde data show high thermal tropopauses at both the stations.

![Diagram of pressure-latitude cross section](image)

**Fig. 7.5** Pressure–latitude cross section of meridional wind vector (black arrow, m/s), ozone mass mixing ratio (red contours, 10^-6 kg/kg), and potential vorticity (PV) (blue lines, contours of 1–5 PVU units) at 84.25°E (over Gerze site) on 25 Feb 2008, 12:00 LT between 10°N and 50°N in latitude and between 1000 hPa and 20 hPa in the vertical derived from ERA interim data. The wind vector shows meridional divergent wind (m/s) with vertical velocity (Pa/s) exaggerated by 60 times.
The behaviour of the tropopause folding events over the Tibetan Plateau

The jet core positions, in both periods in Fig. 7.1 and Fig. 7.2, were compared. From 25 February 2008, 0:00 to 26 February 2008, 13:00 LT, the jet core propagates from around 32° N to 26° N. The tropopause folds also have coordinated displacement when the radiosondes at Gerze site observed MT events (Fig. 7.1).

During the single tropopause periods from 29 Feb. 2008, 01:00 to 1 Mar. 2008, 13:00 LT, the jet core locates at 32° N-33° N. The tropopause folds also move northward and weaken compared to folds observed during the period of double tropopause. The folds have less influence upon the UTLS above Gerze station. The tropical tropopause has entered the station, which reflects a high tropopause in this period. According to the analysis of these two time series, the tropopause folds follow the displacement of the subtropical jet over the Plateau. This process significantly influences the structure of UTLS above the Plateau.

Stratospheric intrusions of air usually occur with tropopause folds. A frequent descent air circulation above the Plateau has been observed during winter (e.g. Yanai et al. (1992)). This flow more strongly favours stratospheric intrusions into troposphere during winter. Seasonal changes in PV in a pressure-longitude cross section do not reveal any folds. This suggests that meridional folds rather than zonally aligned folds are observed in the UTLS above the Plateau. The southern downward intrusions of high ozone from the stratosphere to the Plateau troposphere in winter are caused by the meridional folds rather than zonal folds. The intrusions in winter caused by meridional folds are more frequent than in summer. This effect can be one reason for total column ozone value above the Plateau in winter higher than that of summer. Recently, a middle tropospheric ozone minimum in June (Liu et al., 2009), and low ozone concentration in the UTLS are observed (Tobo et al., 2008). Liu et al. (2009) explained middle tropospheric ozone minimum with the effect of Asian summer monsoon. Together with the effect of Asian summer monsoon anticyclone, less intrusions of stratospheric air in summer may contribute to the above mentioned phenomenon of low ozone concentration in middle and upper troposphere. The high values of tropopause heights above TP in summer contribute also to low total column ozone in summer.

The Dali station (25.71° N, 100.18° E) has lower frequencies of MT throughout the year (in Table 7.1), which can be related to the scarce propagation of the westerly jet and folds passing the latitude of the station. The movement of the
jet in north-south direction is accompanied by the latitudinal movements of tropopause folds, which influence the UTLS structure above the Plateau. The variation of UTLS dynamical structures was reflected by the single or multiple tropopauses observed at the Plateau stations. The observed two distinctive peak values in LRT1 height distribution in winter could be related with latitudinal movement of folds and jets. Both the heating of the Plateau and poleward extending of tropical tropopause (Castanheira et al., 2009; Pan et al., 2009) can contribute to the observed single high tropopause during IOP2 and IOP3, when the folds retreat northward.

Due to the close relationship between the jets and tropopause folds (Barnes and Fiore, 2013; Shapiro, 1980), we display the seasonal variations of the jets and tropopause folds (highlighted by 1 and 2 PVU contour lines) to investigate seasonal variation in tropopause structure (Fig. 7.6). Subtropical jet is strengthened by the cooling effect of the Plateau (Ye and Gao, 1979), and locates at the south of the Plateau during winter season. The folds in February and March are deepest. With the development of the monsoon, subtropical jet weakens and retreats to the north of the plateau, as can be seen by poleward movement of the jet during July. In accordance with seasonal movement of the jet in north-south direction, the tropopause folds also move from south in winter to north in summer. The variation of tropopause folds over the Plateau has also been shown by Sprenger et al. (2003) and the influence of the northward movement of the jet stream by Ding and Wang (2006). Their study attributed summertime maximum of ozone concentration at Waliguan (36.28° N, 100.90° E, North East of the Plateau) to stratospheric intrusions, which were generally associated with prevailing upper-level jet streams above the station.
Fig. 7.6 Pressure–latitude cross section of monthly average zonal wind (black contours, m/s) and potential vorticity (PV) (red lines, contours of 1–2 PVU units) at 90°E between 15°N and 45°N in latitude and between 1000 hPa and 20 hPa in the vertical derived from ERA interim data of 2008.
7.6 Discussion and Conclusion

LRT1 (first lapse rate tropopause) over the Plateau has low and high tropopause characters during the winter, and has a consistent high tropopause character in summer. The characteristics of LRT1 observed in our paper was also captured by Tian et al (2008) using ERA data. Schmidt et al. (2005) also observed the bimodal distribution of LRT1 in the midlatitudes between 30 and 50° N employing GPS radio occultation data. The satellite and reanalysis data have furthered our knowledge about global and climatic characteristics of multiple tropopause (MT). The temporal coverage of our radiosonde dataset is limited and the analysis of Randel et al. (2007b) and Añel et al. (2008) represent a climatology on a larger time period including a larger set of data. The vertical and horizontal resolution of our data and dataset used in Randel et al. (2007b) and Añel et al. (2008) is different. All make a sounding comparison between us and others impossible. The MT statistics over the Tibetan Plateau using ERA-40 and GPS data maybe have shortcomings to capture more tropopause. More attentions should be paid to the MT events above the Plateau and the influence of resolution on the statistics of MT. The phenomena of recent rising trend of MT events (Castanheira et al., 2009) can cause our estimations of MT occurrence in 2008 higher than that of past tenth years and partially explains our results of MT events higher than the climatic value of Randel et al. (2007b) and Añel et al. (2008). Whether are there any other significant difference in MT frequencies still needs to be testified with more high resolution data of radiosonde.

The dynamical tropopause above the Plateau in pressure-longitude cross section and pressure-latitude cross section has been discussed. No folds revealed in pressure-longitude cross section above the Plateau, which is different from the pressure-latitude cross section. The dynamical tropopause exhibits a meridional folded structure around the subtropical jet. This folded structure has been demonstrated by radiosonde observations over the Plateau. This paper further shows the close relationship between the folds in pressure-latitude cross section (meridional folds) and the subtropical jet stream above the Plateau. The meridional folds related with the westerly jet dominate UTLS above the plateau. Zhang et al. (2010b) pointed out that the Plateau terrain has no significant effect on the morphology of folds. The westerly jet maybe the dominate influence on the morphology of the folds above the Plateau.
A persistent maximum of chemical constituents in the UTLS (Park et al., 2009) may also reduce the ozone content in a column air above the Plateau. Tian et al. (2008) pointed out that the low column ozone over the TP is rather related to transport than to chemical reactions. We found intrusions of stratospheric air frequently happen in spring time but rarely occurs in summer time over the Plateau. Our results support that the reduced southern downward transport of stratospheric air may partially contributes to the phenomena of low ozone content in the middle troposphere and low column ozone in summer time above the Plateau. The height variation of the tropopause discussed in our paper can also be used to explain the low value of total column ozone in summer above the TP.

The lack of high resolution data in the atmosphere has hampered our knowledge about tropopause characteristics so far and limiting our ability to evaluate model performance in UTLS (Hegglin et al., 2010). In addition to our data set, more radiosonde data with high vertical resolution are indispensible to further quantify tropopause structure and variability above the Tibetan Plateau. Further data are needed for a better understanding of the high MT frequency during winter above the Tibetan Plateau. Discovering MT frequency are really high in periods of winter time, this work could further deepen our understanding in UTLS structure over the Plateau.
Chapter 8 The connection of the deep atmospheric boundary layer to the upper troposphere and lower stratosphere over the Tibetan Plateau

8.1 Abstract

In the upper reaches of the Earth's atmosphere, ozone is highly beneficial, since it filters the ultraviolet light from the Sun, making the Earth more pleasant to live on. At ground level, however, ozone is bad element, because it is highly corrosive, and it can damage human’s respiratory organs and injure the vascular systems of plants. When it is high enough, an ozone pollution event happens. Due to the human’s activity in this region is limited and pollution is not heavy. The dynamic transport of ozone from upper air to plateau surface may dominate the surface ozone content. Hereby, we study the interactions occurs between deep ABLs and the low tropopause during winter over the Tibetan Plateau by trajectory analysis. The deep ABL and its coupling to the UTLS are significant to the vertical distribution of ozone over the Tibetan Plateau. This study may indicates a major role for stratospheric intrusions in contributing to high-ozone events over the Plateau.

* This Chapter is based on:
8.2 Introduction

Over the past thirty years, some characteristics of the plateau’s ABL structure have been revealed (Li et al., 2006; Ma et al., 2009b; Sun et al., 2007; Xu et al., 2002; Yanai et al., 1992; Yanai and Li, 1994; Yang et al., 2004). From Chapter 6, it is now known that the ABL over the plateau is deeper than that over the lowlands (Fan et al., 2011; Li et al., 2006; Yang et al., 2004; Zuo et al., 2005). However, few studies have investigated the connection of the high ABL with the upper troposphere and lower stratosphere (UTLS), even though the Tibetan Plateau is regarded to be a pathway of mass exchange between the troposphere and stratosphere (Zhou et al., 2006).

The elevation of the Tibetan Plateau varies between 3000 and 8848 m above sea level (ASL). The top of the ABL may be as high as 9000 m ASL, and near to the location of the tropopause. This is a result of both the plateau’s elevation and the deep ABL. This can result in a stronger interaction between the UTLS and the ABL than in lowland areas (Kalabokas et al., 2013). Chen et al. (2012a) concluded that the Tibetan Plateau is one of three key source regions for transport from the boundary layer to the tropopause in the Asian monsoon region. Studies have shown that multi-tropopause events, which are closely related to tropopause folds, frequently occur over the Tibetan Plateau (Añel et al., 2008; Chen et al., 2011a; Randel et al., 2007a). Tropopause folds can cause stratospheric air to be transported downwards to the ABL through a number of different processes (Johnson and Viezee, 1981). A previous study indicated that dynamic transport was the main factor that influences the vertical distribution of ozone over the Tibetan Plateau (Chen et al., 2012b). Surface ozone over the Tibetan Plateau is sensitive to ozone perturbation in the upper layers (Yang et al., 1999), which may suggest that ozone above the planetary boundary layer may strongly influence surface ozone. It is clear that the interaction between the plateau’s ABL and the UTLS is also important for troposphere-stratosphere exchanges.

In order to check possible exchanges between the high CBL and the stratosphere, a trajectory model forced with ERA-Interim data was used to simulate whether the air mass can be transported from troposphere to stratosphere, and vice versa. The FLEXPART v8.1 model (Stohl and James, 2004; Stohl and James, 2005) was used to perform the trajectory analysis.
8.3 The correspondence between variations of tropopause folds and the CBL

The results in Fig. 8.1 indicate that the tropopause was lowest and the CBL was deepest during IOP1, while during IOP2 and IOP3 the tropopause was higher and the CBL was shallower. Chen et al. (2011a) demonstrated that variations in the height of the tropopause were related to tropopause folds. In the next section, the displacement of tropopause folds were analyzed with synchronous CBL variations. For winter days with a low tropopause and a deep CBL, we use trajectory analysis to test the extent of interactions between the CBL and the UTLS over the Plateau.

The UTLS structure around the plateau was analyzed using ERA-Interim reanalysis data, which are plotted as meridional cross sections of the atmosphere at 12:00 GMT for 25-28 February during IOP1 (Fig. 8.2). The ABL is usually fully developed in the afternoon, so the ERA-Interim data at 12:00 GMT was used here. The observed heights of lapse-rate tropopause (LRT) and top of CBL,
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derived from radiosonde measurements at 19:00 BST, are labeled with red triangles and circles in the two figures. Potential vorticity (PV) isolines, zonal wind, and potential temperature derived from the reanalysis data are also shown on these graphs. The ozone content acts as a marker to indicate the tropopause fold events.

Fig. 8.2 Meridional cross-section at 84.25° E (over Gerze site) at 20:00 BST, for the period 25 Feb. to 28 Feb. 2008, derived from ERA interim data, including zonal winds (cyan contours, m/s), potential vorticity (yellow lines, contours of 1, 2, 4 PV units), ozone (solid color, ×10⁶ kg/kg) and potential temperature (red contours, K). The color bar is the scale of ozone concentration. The area in black shows the cross section of the Tibetan Plateau terrain. The red triangles and circles show the position of the LRT and the top of CBL.

During IOP1 the westerly jet stream runs along the 28°N parallel, directly above the southern Tibetan Plateau (Fig. 8.2). The dynamical tropopause, identified by the isolines of 1 and 2 potential vorticity unit (PVU), exhibits a folded structure over the plateau. Several studies have revealed that wintertime tropopause folding beneath the subtropical jet stream can lead to a downward transport of stratospheric ozone into the middle troposphere, the lower troposphere or even the ABL (Betts et al., 2002; Cooper et al., 2005; Johnson and Viezee, 1981; Langford and Reid, 1998). A southern downward motion on the poleward side of the jet transports stratospheric air downwards isentropically. On 25 February,
the layer between 2 PVU and 4 PVU, which can be considered to represent the UTLS, reached down to 425 hPa over the Plateau and gradually decayed between 26 and 28 February. Figure 8.2a and 8.2b indicated that the top of the CBL on 25, and 26 February was inside the tongue of the stratospheric intrusions.

Figure 8.3 GOME-2 ozone results. Top: location of GOME-2 orbits over the Tibetan Plateau, with the altitude scale in km. Bottom left: GOME-2 ozone profiles in units of 10^{12} molecules/cm^3, with ERA-Interim number density contours superimposed. Bottom right: GOME-2 ozone profiles with ERA-Interim horizontal wind speed contours superimposed. The arrows below the x-axis in the bottom two plots show the extent of the Tibetan Plateau as illustrated in the top plot. The dashed line is the position of the thermal tropopause, according to the WMO definition. This Figure is made by Jacob van Peet in Chen et al. 2013.

To validate the modeled ozone distribution derived from ERA-Interim, it was compared with data from the GOME-2 orbit closest to the Gerze station (see Fig. 8.3) on 25 February 2008. The observed ozone concentrations agree very well with the ERA-Interim data, and the tropopause folding event is clearly linked to the location of the jet stream over the Tibetan Plateau.
During IOP3, the jet core moved north of the plateau to 42° N, as did the slightly folded UTLS structure (Fig. 8.4). The quasi-vertical lines of potential temperature over the Tibetan Plateau moved to the north, outside of the plateau domain. The potential temperature was more horizontally stratified over the plateau, and the troposphere was more stable during the monsoon season than during the winter time. Griffiths et al. (2000) showed that the tropopause folds increase the generation of potential instability in its vicinity. When the ABL developed sufficiently to reach the height of the unstable intrusion area, it grew even higher. During the deep intrusions of 25 and 26 February the ABL was deeper than during the shallow intrusion of 28 February. During IOP3, when there were no tropopause folds, the ABL was much shallower than during IOP1.

Figure 8.4 Meridional cross-section at 84.25° E (over Gerze site) at 20:00 BST, for period 7 Jul. 2008 to 10 Jul. 2008 BST. The red triangles and circles show the position of the LRT and the top of CBL.

Figure 8.5 gives a simple illustration of tropopause folds, westerly jet displacement and convective boundary layer variations during winter and summer time. The strongest westerly jet is situated above the plateau, with a tropopause fold below it. The frequency of stratospheric intrusions caused by tropopause folds is high during winter, which makes the upper tropospheric air unstable and facilitates ABL development. When the jet moves to the north of the Plateau and becomes weaker in summer time, the stratospheric intrusions
rarely reach the upper tropospheric air above the plateau. However, whether these variations generally happen still needs more climatological study.

Figure 8.5 A schematic illustration of the correspondence between tropopause folds, westerly jet displacement and convective boundary layer variations during winter and summer time.
Analysis shows that summertime variability of surface ozone depends strongly on the jet stream position over eastern North America, and surface ozone variability follows the robust poleward shift of the jet on decadal time scales (Barnes and Fiore, 2013). Jet position can thus serve as a dynamical predictor of surface ozone variability over eastern North America (Barnes and Fiore, 2013). This conclusion is suggested to be tested with surface ozone measurement over the Tibetan Plateau in future.

8.4 Lagrangian analysis

To gain a better insight into the relationship between a deep ABL and stratosphere and troposphere exchange (STE), trajectories of air masses for the IOP1 were computed, using the Lagrangian particle dispersion model FLEXPART v8.1 forced with ERA-Interim reanalysis data. This approach is useful for studying the origin of air masses in the UTLS (Añel et al., 2012). The analysis was focused on IOP1, because the deep ABL and tropopause folds create a higher potential for STE.

The spatial domain used for FLEXPART was 0-180°E × 0-60°N, with a vertical extent of up to 20 km, well above the top of the ABL and the tropopause. To perform the cluster analysis only the particles within a 10° × 10° box around the station of Gerze at 12:00 UTC on 25 February 2008 were used.

The analysis was split by height within the reference box for the central date into three layers: 5-9 km, 9-10 km and 10-14 km ASL, which roughly represent the ABL-upper troposphere, tropopause, and lowermost stratosphere layers. The results are also clustered according to the main patterns of movement of particles determined by the model. The light grey part represents later simulated times and the dark part earlier times (see Fig. 8.6). The small square shows the central point (12:00 UTC on 25 Feb 2008). Each point/diamond represents one time step (± 3 hours).
Fig. 8.6 Clusters of trajectories for particles within a 10° x10° box around the station of Gerze and within the 10-14 km layer at 12:00 UTC on 25 February 2008. The smallest square shows the central date (12:00 on 25 Feb 2008 GMT). Each point/diamond represents one time step (3 hours). The upper plots show the height-longitude representation: the solid black line represents the height of the tropopause for each time step as calculated by the FLEXPART model, and the diamond at the lowest level corresponds to the earliest time simulated. Colors correspond to the mean PV of the air mass: orange represents 2.5 - 3.0 PVU, yellow 2 - 2.5 PVU, green 1.5 - 2 PVU. The lower plots show the latitude-longitude representation: the dark grey part represents earlier times in the simulation, while the light grey part represents later times. This figure is made by Laura de la Torre.

The positions of most of the clusters in the three vertical layers show that the air masses stayed within the same layer for the analyzed period. We do not include the corresponding figures here. However, it is noteworthy that two clusters, accounting for a higher percentage of particles for the 10-14 km layer, show a
substantial air mass from lower levels (around 8 km) and that the PV for these air masses decreased after the central date. This result indicates the possibility that air at UTLS levels is irreversibly mixed with air from the upper ABL, which can contribute to the development of a high ABL.

8.5 Discussion and Conclusions

The structure of the troposphere over the high-altitude terrain of the Tibetan Plateau is still poorly understood, despite its impacts on regional synoptic and atmospheric circulation. Based on high-resolution radiosonde observations, we have shown measurements of a deeper boundary layer than in any previous research. The radiosonde dataset of three different periods in one year demonstrates a significant seasonal contrast in ABL height. Following the suggestion of Santanello et al. (2005) that atmospheric stability is the most influential variable controlling ABL development, we compared the atmospheric stability in each observation period and found that in winter time the stratification of the troposphere was related to tropopause folding events. Due to the high frequency of tropopause folds co-existing with the westerly jet situated above the plateau in wintertime (Sprenger et al., 2003), the isentropic surfaces in the middle and lower troposphere intersect with the Tibetan Plateau. This distribution of potential temperature lines demonstrates that the troposphere over the plateau is fairly stable during winter time, which makes it easier for a dry and warm eddy to be transported upwards. The instability associated with tropopause folds also provides a potential interpretation of the high ABL. Due to convection and vigorous vertical mixing caused by dry heating at the plateau’s surface, a dry-adiabatic lapse rate is established in the high CBL. By the afternoon, the dry thermal convection originating from the heated surface can reach the upper layers of the troposphere. A well-mixed potential temperature and water vapor layer can clearly be identified, when westerly wind dominates over the plateau. The mechanical turbulence caused by shear of the westerly wind can also contribute to the mixing of potential temperature. All these factors make the ABL much deeper, allowing the formation of larger thermals and eddies (Stull, 1988). It should be pointed out that radiosonde observations are single points in time and space. We have taken radiosonde data from one station to be representative of the desired area. The mixing height is also influenced by advection, radiation and ABL cloud, which have been ignored here. Indeed, due to the low air density and intense solar radiation, the ABL of the plateau has the potential to grow much deeper than the ABL over lowlands (Yang et al., 2004). Thermodynamic sounding profiles
suggest that the direct heating of the Tibetan Plateau during winter daytime can rise to as high as 5 km above ground. These results call for a reinterpretation of the response of the ABL over the plateau to uplifted surface heating and the regional meteorological situation. By comparison with deserts and other arid regions that also have a high ABL (Cuesta et al., 2008; Gamo, 1996; Zhang et al., 2011), the elevation of the Tibetan Plateau makes it easy to demonstrate how transport processes in the UTLS over the plateau affect the ABL. A close surface-troposphere-stratospheric coupled system may exist over the plateau. It has been already pointed out that the ozone flux from stratosphere to troposphere to ABL in spring is greatest over the Tibetan Plateau (Skerlak et al., 2012; Skerlak et al., 2013). Johnson and Viezee (1981) identified four mechanisms governing the fate of stratospheric air injected into the lower troposphere. Their mechanism 2 is a stratospheric intrusion down to the ABL, and the lower portion of the intrusion is mixed down to the ground by turbulent eddies and convection at the top of the boundary layer. Our observations show that the top of the CBL on 25 and 26 February was connected with the stratospheric intrusions, which is a typical illustration of mechanism 2. In this case, air from the lower troposphere can be mixed by turbulence and transported upwards to the tongues of the intrusions. The intrusions can also transport ozone downwards into the ABL, even down to the ground level. It has been pointed out that the strong vertical mixing in the daytime CBL can bring upper-level ozone downward to augment surface ozone production (Rao et al., 2003; Zhang and Rao). Therefore, the connection between a deep CBL and the UTLS is important to the plateau's surface ozone pollution, which is already considered as a dominant factor in diurnal variation of surface ozone in America (Hu et al., 2012; Lin et al., 2008). Wang et al. (2006) suggested that the surface high-ozone events in the northeastern Tibetan Plateau were mostly caused by the downward transport of upper tropospheric air. The deep ABL and its coupling to the UTLS are also significant to the vertical distribution of ozone over the Tibetan Plateau. Barnes and Fiore (2013) discussed that jet position may also modulate ozone variability in northern midlatitude regions.
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Summary

This research aims to improve our understanding of the land-atmosphere interactions over the Tibetan Plateau. First, the model for calculating the land surface fluxes needs to be improved. Afterwards the model result was used to improve our understanding of the plateau surface heating and its interaction with the typical atmospheric layers.

Surface Energy Balance System (SEBS) model was used to quantitatively estimate the energy and water exchange between the plateau land surface and the atmosphere above it. The surface energy balance system has been updated in three ways: 1) to model regional and continental land surface energy balance using a canopy-height-based parameterization method, 2) to improve accuracy of sensible heat simulation using a developed turbulent parameterization in the model and 3) to derive land surface energy fluxes over a meso-scale mountain area.

1. The improvement in the canopy-height-based turbulent parameterization to the model has been accomplished through measurements of ground based flux towers and up-scaling from station points to meso-scale and continental area. A dynamic map of canopy height at different scales was constructed to calculate the canopy-related roughness length, which was then used to calculate the land surface energy fluxes.

2. A comprehensive sensitivity analysis of all the parameters in SEBS has also been conducted. It was found that kB_1 over the study area showed the most problems. Problems with the kB_1 parameterization method in the original version of SEBS have been resolved by this study.

3. A Topographically Enhanced SEBS (TESEBS) version was developed for deriving land surface energy fluxes over a meso-scale mountain area.

The new model versions have been used to generate spatial-temporal dynamics of land surface energy and water fluxes at different scales. The performances of the new versions at different scales were shown by comparisons with ‘ground truth’ measurement. The model results were validated by using the measurement at different flux tower stations which includes land covers of bare soil, alpine meadow, forest, cropland, orchard, grassland, wetland, and snow.
Summary

Then we analyzed the plateau Atmospheric Boundary Layer (ABL) structure (which is directly related to ground surface heating) with the results of the improved SEBS model. Three intensive ABL observation periods were adopted to represent late winter, monsoon onset and monsoon time. The ABL height during the three phases is clearly different. The highest ABL was observed during late winter, with the lowest ABL in monsoon time and intermediate ABL in the phase of monsoon onset. As the ABL height during daytime can be as high as 5 km above the ground, the absolute height of the ABL over the Tibetan Plateau (with an averaged elevation of 4 km) is around 9 km. As this height is very close to the tropopause layer at mid-latitude, this causes stronger connections between the plateau ABL and Upper troposphere and Lower Stratosphere layers (UTLS) than other area. The seasonal variations in tropopause (the boundary between troposphere and stratosphere) height were diagnosed from the upper air synoptic situation, especially the tropopause folds. The tropopause folds are largely controlled by the westerly jet above the plateau. From observations of the ABL and model-simulated stratosphere intrusions, the connections between the high ABL and downward folds intrusion have a higher chance in winter than other times of a year. Our estimation of multi-tropopause occurrence over the plateau can sometimes much higher than other studies. The stratospheric intruding episodes are generally associated with the presence of a subtropical jet stream over the plateau. The complex structure of dynamic tropopause folds over the Plateau have been reflected by the thermal multi-tropopause events observed by radiosondes. The intrusion of air masses from the stratosphere may contribute to a higher upper tropospheric ozone concentration in winter than in summer above the plateau.

As a final note we would like to highlight that the dynamic transport of ozone from the upper air to plateau surface may dominate the surface ozone content (as human activity in the Tibetan Plateau region is limited and pollution is not heavy). The deep ABL and its coupling to the UTLS are significant factors in the vertical distribution of ozone over the Tibetan Plateau. This study may indicate a major role for stratospheric intrusions in contributing to high-ozone events over the plateau. In future, more observational ozone dataset should be included to verify the conclusions.
Samenvatting

Het doel van dit onderzoek is het verbeteren van onze begrip van land-atmosfeer interacties op het Tibetaans plateau. Om tot dit doel te komen moet eerst het model welke de fluxen boven het landoppervlakte berekent verbeterd worden. Hierna kunnen de model resultaten gebruikt worden om ons begrip van de verwarming van oppervlakte en de interactie met de atmosfeer te verbeteren.

Het SEBS (Surface Energy Balance System) energie balans model is gebruikt om de interactie van energie en water tussen het landoppervlak van het plateau en de atmosfeer daarboven kwantitatief te schatten. Het model is in dit onderzoek op 3 punten verbeterd: 1) om regionale en continentale fluxen te schatten met een gewas-hoogte parameterisatie, 2) om accuratesse te verbeteren met een verbeterde turbulente parameterisatie en 3) om land oppervlakte fluxen in een bergachtig gebied te verbeteren.

1. Het model is verbeterd voor schatting van regionale en continentale energie balansen door het implementeren van gewas-hoogte-parameterisatie. De verbetering in SEBS is gedaan op basis van evaluaties van grond stations en de schaling van deze punt metingen naar meso- en continentale schaal. Voor het berekenen van de fluxen is een dynamische kaart van gewas hoogte op verschillende schalen geconstrueerd om de ruwheid van het aardoppervlak te berekenen.

2. Ook is er een gedetailleerde sensitiviteits analyse gedaan voor alle parameters van SEBS. Hieruit bleek dat de kB_1 parameterisatie problemen veroorzaakte. Deze problemen in de kB_1 parameterisatie (een van de belangrijkste berekeningen in SEBS) zijn opgelost in deze studie.

3. Een Topografische Verbeterde versie van SEBS (TESEBS) is ontwikkeld voor het juist schatten van meso-schaal land-atmosfeer fluxen boven een bergachtig terrein.

De nieuwe versies van het model zijn gebruikt om de spatiale-temporale dynamische land-atmosfeer fluxen te genereren. De model resultaten zijn gevalideerd met grondmetingen op verschillende flux torens. De landtypes waar op deze torens stonden zijn: kale bodem, alpenweide, bos, akkerland, boomgaard, grasland, moeras en besneeuwde gletsjers. De resultaten van het model lieten goede overeenkomsten zien met de verschillende grond metingen.
Samenvatting

Hierna hebben we de resultaten van het verbeterde SEBS model gebruikt om de structuur van Atmosferische Grens laag (ABL) (welke direct gerelateerd is met de verwarming van oppervlakte) te analyseren. Drie observatie periodes zijn geadopteerd om de volgende ‘seizoenen’ te representeren: ‘laat winter’, ‘opkomende moesson’, en ‘moesson’. De ABL hoogte gedurende deze drie fases is zeer verschillend. Gedurende de late winter is de ABL hoogte gevolgd door de ‘opkomende moesson’ en gedurende de ‘moesson’ fase is de ABL het laagst. Omdat de maximale relatieve hoogte van de ABL ongeveer 5 km boven de oppervlakte is, betekent dit dat op het Tibetaans Plateau (met een hoogte van 4km) de absolute hoogte 9km is. Omdat deze hoogte dicht bij de Tropopause laag ligt is er een sterke connectie tussen de verwarming van het Tibetaans Plateau en de Boven Troposfeer en beneden Stratosfeer (UTLS) lagen. De verandering per seizoen in de Tropopause (de grenslaag tussen de Troposfeer en de Stratosfeer) zijn gediagnostiseerd met de lucht-synoptische scenario’s waarbij speciale aandacht ging aan de Tropopause plooiien. De Tropopause plooiien worden voor het merendeel gecontroleerd door de westelijke luchtstroom gedurende de winter. Uit ABL observaties en model simulaties van de stratosfeer intrusies is gebleken er een hogere kans bestaat voor connecties tussen de hogere lagen van de ABL en de intrusies van de plooiien in de winter dan in andere tijden van het jaar. Onze berekeningen van Multi-Tropopause (MT) gebeurtenissen blijken soms zo hoog als 80% te zijn gedurende de winter. Dit betekent dat er meer aandacht besteedt moet worden aan MT gebeurtenissen boven het Tibetaans Plateau. De stratosferische intrusies zijn over het algemeen geassocieerd met de aanwezigheid van een subtropische luchtstroom over het plateau. De complexe structuur van de dynamische plooiingen in Tropopause zijn geobserveerd door radiosonde observaties van de thermische MT gebeurtenissen. De intrusies van stratosferische lucht in de Troposfeer kan er toe lijden dat er een hogere ozon concentratie is boven het Tibetaans Plateau gedurende de winter.

Als outlook willen we ook nog belichten dat het dynamisch transport van ozon van de bovenkant van de stratosfeer naar beneden de concentraties op de oppervlakte van het Tibetaans Plateau zal domineren (omdat de populatie en gerelateerde luchtvervuiling laag zijn al hier). De ABL en zijn koppeling met de UTLS zijn daardoor significante factoren in de verticale distributie van ozon boven het Tibetaans Plateau. Het onderzoek in deze dissertatie wijst de sterke rol van intrusies van hoge concentraties. In de toekomst meer ozon observaties zullen echter deze claim moeten verifiëren.
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