The Effects of Anomalous Snow Cover of the Tibetan Plateau on the Surface Heating[1]

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ABSTRACT

On the basis of snow data and AWS (Automatic Weather Station) data obtained from the Tibetan Plateau in recent years (1993 to 1999), the features of sensible heat, latent heat and net long-wave radiations are estimated, and their variations in more-snow year (1997 / 1998) and less-snow year (1996 / 1997) are analyzed comparatively. The relationships between snow cover of the Tibetan Plateau and plateau's surface heating to the atmospheric heating are also discussed. The difference between more-snow and less-snow year in spring is remarkably larger than that in winter. Therefore, the effect of anomalous snow cover of the Tibetan Plateau in winter on the plateau heating appears more clearly in the following spring of anomalous snow cover.

Key words: Tibetan Plateau, Snow cover, Effects, Surface heat fluxes

1. Introduction

It is known that the Asian and global climates are greatly affected by the Tibetan Plateau, and there is a good relationship between the winter snow cover over the Tibetan Plateau (Qinghai-Xizang Plateau) and the summer rainfall in China, especially in the Yangtze River (Changjiang River) valley (Li, 1996). The anomalous snow cover on the Tibetan Plateau during the winter season of 1997 to 1998 is regarded as one of the important cases that have some remarkable effects on the heavy rainfall in the Yangtze River valley in 1998 (Li, 1999; Song, et al., 1999).

Qinghai-Xizang Plateau Meteorological Science Experiment (QXPMEX) was carried out by China from May to August in 1979, some important results on the plateau dynamics and heating influence on the general circulation and weather systems were obtained. However, the surface processes and their climatic feedback effects, especially those relating to the variation of plateau heating and the processes in air-land energy exchange are still little known.

In this study, the influence of winter snow cover on the Tibetan Plateau on the plateau heating is emphasized. The bulk transfer coefficients are determined by the profile-flux

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method and the surface fluxes of momentum, sensible heat and latent heat are computed by the bulk transfer formulations. The gradient observational data of the atmospheric surface layer were used from July 1993 to March 1999 obtained from AWS (Automatic Weather Station) installed in Lhasa, Nagqu, Lingzi and Rikeze on the Eastern Tibetan Plateau under the P.R. China–Japan Cooperation Research Project on Asian Monsoon Mechanism (1993—1999), the AWS observation in these stations is also a part of IOP (Intensive observational period) of TIPEX (The Second Tibetan Plateau Atmospheric Science Experiment) from May to August in 1998. The anomalous variations of surface heat fluxes in winter and the following spring between a less–snow winter (1996 / 1997) and a more–snow winter (1997 / 1998) are analyzed comparatively.

2. Data and preprocessing

The AWS observation data were obtained from four plateau stations at Lhasa, Rikeze, Nagqu and Lingzi (see Table 1). Observational elements include

(a) wind speed at 10 m, 5 m (this level was changed to 1.5 m after June 1995) and 2.5 m height,

(b) wind direction at 10 m height,

(c) gusty wind speed at 10 m height,

(d) relative humidity at 10 m height,

(e) air temperature at 10 m, 5 m (this level was changed to 1.5 m after June 1995) and 2.5 m height,

(f) precipitation,

(g) surface atmospheric pressure,

(h) radiation: global solar radiation, downward total radiation, reflective solar radiation and upward total radiation,

(i) soil temperature (0, 4, 8, 16, 32 and 80 cm at Lhasa, Rikeze and Nagqu or 0, 1.25, 2.5, 5, 10, 20, 40 and 80 cm at Lingzi),

(j) soil moisture (0–15 cm and 15–30 cm).

Table 1. The geographic locations and features of four AWS sites

Site	Lhasa	Rikeze	Nagqu	Lingzi
Longitude (E)	91°08′	88°53′	92°04′	94°28′
Latitude (N)	29°40′	29°15′	31°29′	29°34′
Altitude (m)	3649	3836	4507	2991
Topography	Valley	Basin	Plateau	Valley
Background condition	City	Suburbs	Countryside	Countryside

(a)—(h) are obtained from a part above ground of AWS named AANDERAA, the sampling interval of data is 10 minutes before July 1994 and 20 minutes since July 1994; (i)—(j) are obtained from a part under ground of AWS named LAND SCALE before July 1995 and named KADEC since July 1995, the time interval of data is half an hour before July 1994 and one hour since July 1995. If the data of wind speed and temperature at 2.5 m are not collected correctly by AWS when bulk Richardson number is estimated, the same observed elements at 5 m or 1.5 m are chosen to replace, and the related heights are replaced meanwhile.

The precision and reliability test of AWS data is the guarantee of the research work in this study. A comparison of the AWS data during the period of July to December 1993, with the routine observational data, has been done by the researchers in Chengdu Institute of Meteorology. The results showed that the values and variation tendency of the AWS data used in this study are all in agreement with each other. Therefore they are comparatively precise and reliable. Moreover, the data sequence indicates that the AWS data are also stable, i.e. there is no obvious difference in the AWS data each year (Li et al., 1996).

The AWS data used in this paper are preprocessed as follow: when the instrument is not in good condition according to observational reports, the data cannot be used. Some considerably anomalous data are filtered with the possible values of meteorological elements from the routine observation stations on the Tibetan Plateau. The data are averaged on daily basis according to GMT before the calculation except for the determination of the surface roughness length.

3. Computational schemes

3.1 *Bulk transfer coefficients*

According to the profile–flux method, the bulk transfer coefficients can be expressed as

$$C_D = \frac{k^2}{[\ln(Z/Z_0) - \psi_M(Z/L, Z_0/L)]^2}, \tag{1}$$

$$C_H = \frac{k^2}{P_{ro}[\ln(Z/Z_0) - \psi_M(Z/L, Z_0/L)][\ln(Z/Z_0) - \psi_H(Z/L, Z_{0H}/L)]}, \tag{2}$$

$$C_E = \frac{k^2}{P_{ro}[\ln(Z/Z_0) - \psi_M(Z/L, Z_0/L)][\ln(Z/Z_0) - \psi_E(Z/L, Z_{0E}/L)]}. \tag{3}$$

where C_D, C_H, and C_E are the bulk transfer coefficients for momentum, heat and moisture, respectively. k is von Karman's constant, it is taken as 0.4. P_{ro} (=0.74) is the turbulent Prandtl number for neutral stability. ψ_M, ψ_H and ψ_E are the integration results involved profile functions with respective heights from Z_0 to Z. The surface roughness lengths Z_0, Z_{0H} and Z_{0E} correspond to wind speed, temperature and humidity, respectively. For convenience, the above three lengths are roughly assumed equal in value. L is Monin–Obukhov stability parameter, its applied expression is $\zeta = Z/L$. Because $\psi_E = \psi_H$ therefore it is usually assumed that, $C_E = C_H$. For stable conditions, ψ_M and ψ_H are taken from the relations suggested by Businger et al. (1971); for unstable conditions, the expressions proposed by Paulson (1970) are used. The stability parameter ζ can be expressed as a function of Richardson number R_{ib}. A set of analytical explicit solutions derived by Byun (1990) is used in this paper. These solutions showed a much better agreement on the estimation provided by a numerical iteration method than the usual approximation, and the procedure in computing these solutions is simpler. For extreme stable conditions ($Z/L > 1$, or $R_{ib} > 0.125$), the stability parameter ζ should be determined by using an iterative procedure to obtain the reasonable values of C_D and C_H.

The surface roughness length Z_0 is one of the important parameters to calculate the bulk transfer coefficients and surface fluxes. As the surface similarity theory is not appropriate for weak wind, the data under strong wind (e.g. above 5 m/s at 10 m height) and for near–neutral condition (i.e. for neutral condition, the absolute value of bulk Richardson number is not larger than a criterion which is taken as 0.0002 in this paper), then the surface roughness length Z_0 is calculated by the above–chosen data of wind speed at 10 m, 5 m (or 1.5 m) and 2.5 m heights from a logarithmic wind law using the least square method (Li et al., 1999).

3.2 *Estimation of the surface fluxes*

According to the bulk transfer theory, the surface flux of momentum, sensible heat and latent heat (evaporation) can be respectively expressed as

$$F_M = \rho C_D U^2, \tag{4}$$

$$F_H = \rho C_P C_H U(T_g - T), \tag{5}$$

$$F_L = \rho \beta L_E C_E U(q_{gs} - q) \tag{6}$$

where ρ is air density and is expressed as an exponential descending function of the sea level elevation in this paper. T_g is temperature on the ground (0 cm), q_{gs} is saturation specific humidity on the ground at T_g, wind speed and temperature of air are also taken as the corresponding values at the highest height (10 m) for a good representation. β is moisture availability factor or evaporation constant, it is an ascending function of soil moisture. This factor is estimated by the observational data of soil moisture, which is better than the method based upon relative air humidity indirectly. Then we have the following parameterization method (Li et al., 2000)

$$\beta = \frac{\overline{W} - W_{ec}}{W_{sat} - W_{ec}} \tag{7}$$

where $\overline{W}$ is the mean of relative soil moisture in two layers of soil, W_{sat} is the mean of saturated soil moisture, W_{ec} is a criterion for meeting the demands of soil evaporation and is determined by the observational data of soil moisture when soil is close to freezing or unfreezing. W_{sat} is the mean of saturated soil moisture in the upper soil (0–30 cm) and is taken from the measurement by Dr. Shigenori Haginoya in Meteorological Research Institute, Meteorological Agency of Japan (through the private mail).

4. Analysis and discussion

4.1 *Analysis method and choice of anomalous snow cover year*

The time series of snow cover on the Tibetan Plateau appears a wave–type variation since the 1980's. For convenience to analyze the effect of snow cover of the Tibetan Plateau on the surface processes, two typical anomalous snow cover years (or winters) were chosen according to the snow data and related climate data from the Tibetan Plateau. The year 1997 (winter from 1996 to 1997) is an abnormally less–snow year, whereas 1998 (winter from 1997 to 1998) is an abnormal more–snow year (as shown in Fig.1, it should be pointed out that Fig.1 at Nagqu is only a representative of the whole Tibetan Plateau to a certain extent).

4.2 *The difference of elements in the surface layer between less–snow year and more–snow year*

4.2.1 *Winter*

Table 2 shows that surface albedo increases remarkably at Nagqu and Lingzi in more–snow year, but not at Lhasa and Rikeze, this indicates that distribution of snowfall on the Tibetan Plateau is highly erratic, and the anomaly of more snow is remarkably at Nagqu and Lingzi. The temperature difference between ground and surface air drops in the more–snow winter except Lhasa and Nagqu. The bulk transfer coefficients for momentum and heat increase in varying degrees except Lhasa which is mainly caused by a rise of unstable conditions. Surface momentum flux not only depends on bulk transfer coefficient of momentum but also surface wind speed, therefore the difference of this flux between less–snow year and more–snow year is not the same at each station. In Fig. 2a, the sensible heat flux decreases to a great extent in more–snow year (especially at Rikeze), which mainly depends on the drop of temperature difference between ground and surface air. The latent heat flux of evaporation in the more–snow year is almost the same as the less–snow year because the soil tends to freeze in winter.

The direct ways in which the surface heats the atmosphere are sensible heat and net long–wave radiation, and the latter has the greatest contribution in this regard. It decreases

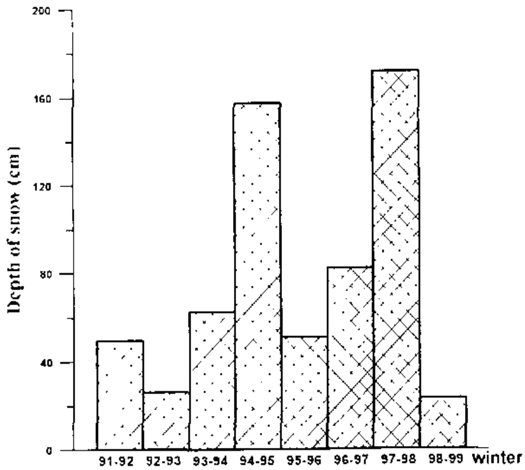

Fig. 1. The yearly variation of snow depth in winter (Nov.–Feb.) at Nagqu.

obviously in more–snow winter at station where the snow cover is remarkably anomalous, for example at Nagqu and Lingzi. In summary, the surface heating in more–snow year decreases remarkably at these obviously anomalous stations, but increases at other stations (such as Lhasa and Lingzi). Then the variations of the surface heating in more–snow year comparing to the less–snow year are not identical at each station in winter (as shown in Fig. 2d).

Table 2. The winter comparison of elements in the surface layer between less and more–snow year

		A_g	Δ_T (°C)	C_D ($\times 10^{-3}$)	C_H ($\times 10^{-3}$)	F_M ($\times 10^{-2}$N / m^2)*
Lhasa	More–snow year	0.18	1.82	3.03	4.03	0.97
	Less –snow year	0.21	0.97	3.06	4.07	0.96
Rikeze	More–snow year	0.20	−3.64	3.51	4.66	1.22
	Less –snow year	0.20	−1.64	2.83	3.70	1.07
Nagqu	More–snow year	0.38	3.10	4.70	6.40	2.74
	Less –snow year	0.30	2.81	4.50	6.08	3.05
Lingzi	More–snow year	0.28	2.42	4.04	5.46	2.57
	Less–snow year	0.22	2.81	3.32	4.45	1.79

A_g is surface albedo, Δ_T is temperature difference between ground and surface air.

4.2.2 *Spring*

The anomalous snow cover of the Tibetan Plateau in winter influences not only the surface processes in winter but also the features of these processes and the general circulation in spring due to the persistence and time–lag effect of snow cover. In spring, snow cover is beginning to thaw with rising air temperature. In Table 3, the surface albedo of more–snow year decreases and is even lower than that of less–snow year, because the soil moisture increases due to melting snow. The temperature difference of more–snow year in spring still deceases,

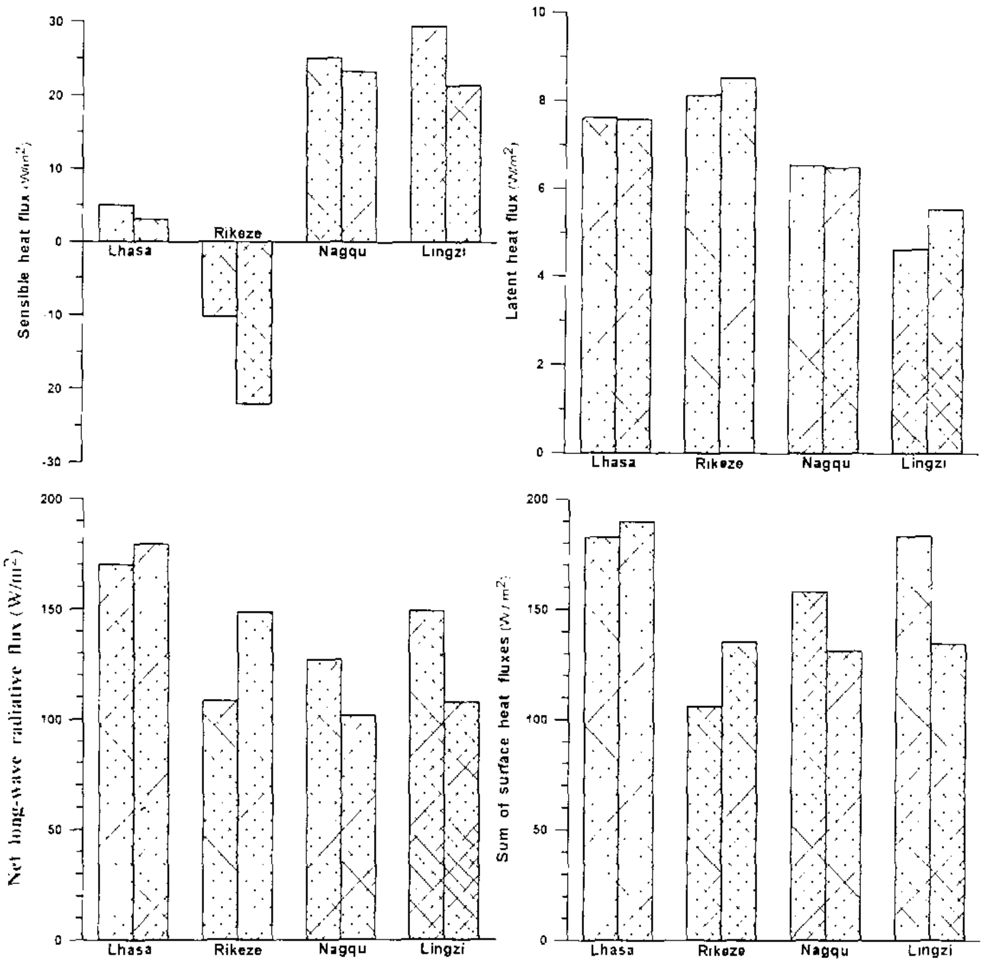

Fig. 2. The surface heat fluxes in less–snow year (left column) and more–snow year (right column) at four stations in winter (a) sensible heat (b) latent heat (c) net long–wave radiation (d) sum of surface heat fluxes to the atmosphere.

Table 3. The spring comparison of elements in the surface layer between less and more–snow year

		A_g	Δ_T (°C)	C_D ($\times 10^{-3}$)	C_H ($\times 10^{-3}$)	F_M ($\times 10^{-2}$N · m^{-2})
Lhasa	More–snow year	0.16	5.71	5.12	6.86	1.68
	Less–snow year	0.18	8.07	5.47	7.41	2.04
Rikeze	More–snow year	0.19	2.07	5.93	8.01	2.28
	Less–snow year	0.19	5.26	5.64	7.57	2.23
Nagqu	More–snow year	0.22	4.64	4.81	6.51	4.86
	Less –snow year	0.26	5.87	4.77	6.45	2.63
Lingzi	More–snow year	0.19	5.71	4.53	6.12	3.67
	Less –snow year	0.19	6.64	4.46	6.04	2.52

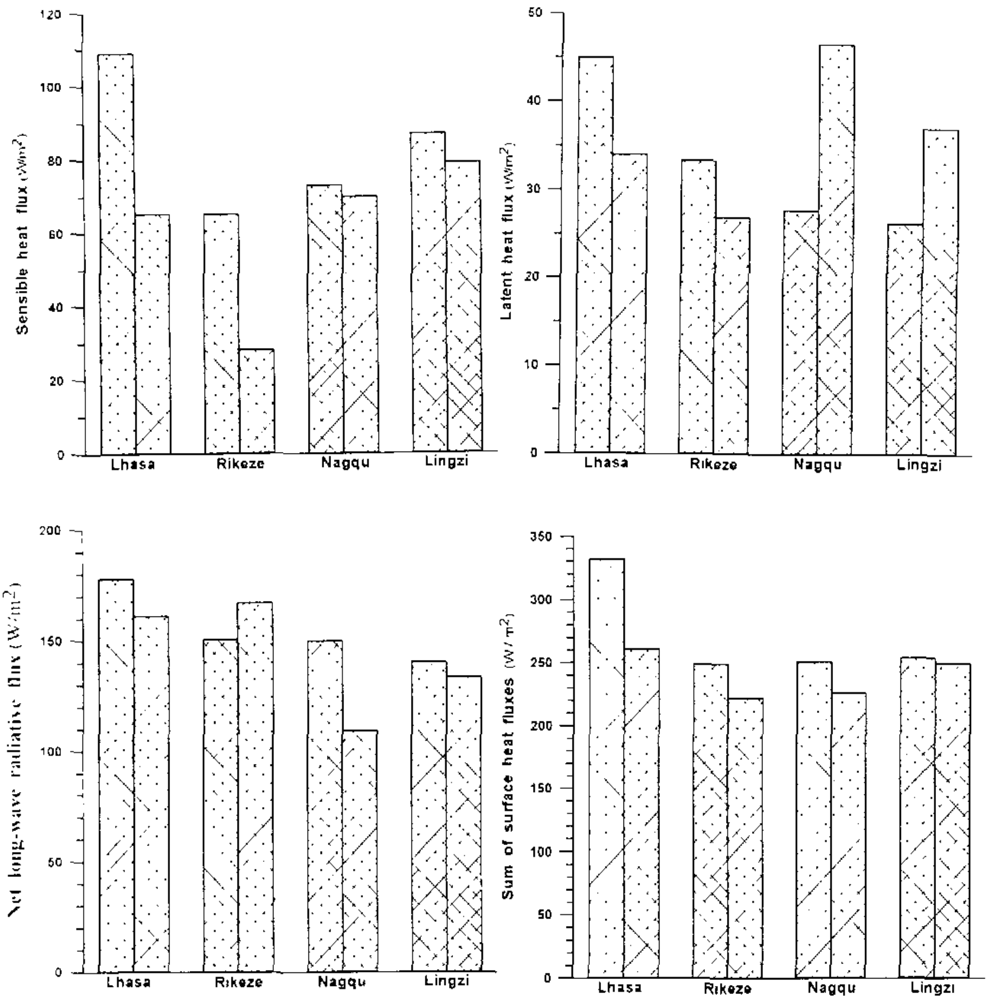

Fig. 3. The surface heat fluxes in less–snow year (left column) and more–snow year (right column) at four stations in spring (a) sensible heat (b) latent heat (c) net long–wave radiation (d) sum of surface heat fluxes to the atmosphere.

its uniformity and variability is larger than that in winter. The variation of the bulk transfer coefficient for momentum and heat in spring between less–snow and more–snow year is the same as in winter. In Fig.3, the difference of sensible heat flux between less–snow and more–snow year is very prominant in spring and is larger than that in winter, especially at Lhasa and Rikeze. The tendency of all the four stations is also the same. The latent heat flux of more–snow year in spring increases with a rise of soil moisture and soil evaporation due to melting snow especially at stations where the snow cover is remarkably anomalous (such as Nagqu and Lingzi) . In spring, the net long–wave radiation obviously decreases in more–snow year except Rikeze. Then the sum of heat fluxes to the atmosphere obviously decreases at four stations (see Fig.3d), and the difference between more–snow and less–snow year in spring is remarkably larger than that in winter. Therefore, the effects of anomalous snow cover of the Tibetan Plateau in winter on the plateau heating appear more clearly in the following spring of anomalous snow cover.

5. Concluding remarks

Comparing more–snow year to less–snow year of the Tibetan Plateau, some elements of surface layer are calculated and analyzed. The characteristics of these elements, especially surface heating to the atmosphere when snow cover is anomalous, are shown and discussed in detail. The main objective of this study is conceptual, the results obtained would be helpful to study deeply the surface processes on the Tibetan Plateau and effects of their variations on the fluctuations in weather and climate of China and surrounding regions. However, it is noted that some AWS data should be checked further, such as soil moisture and long–wave radiation. The influence of anomalous snow cover on the plateau's surface heating to the atmosphere is also to be investigated in quantitative terms, and the climate effect of anomalous snow cover may be closely related with the general circulation. Therefore, extensive research studies are necessary and may be concentrated on the problems described above.

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青藏高原积雪异常对高原地面加热的影响

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摘　要

利用青藏高原积雪资料和中日亚洲季风机制合作研究在1993～1999年获得的自动气象站观测资料，计算了地面感热、潜热和净长波辐射，并对比分析了这些物理量在一个多雪年(1997／1998)和一个少雪年(1996／1997)的变化。关于青藏高原积雪与高原地面加热关系的讨论表明：多雪年和少雪年在接下来的春季里的差别明显大于冬季，因而在冬季发生的青藏高原积雪异常的效应在接下来的春季会表现得更加强烈。

关键词：青藏高原，积雪，影响，地面热通量