Decreased surface albedo driven by denser vegetation on the Tibetan Plateau

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Abstract

The Tibetan Plateau (TP) has fundamental ecological and environmental significance to China and Asia through its influence on regional and continental climates. In recent years, climate warming has caused unprecedented changes to land surface processes on the TP, which would unavoidably undermine the ecological and environmental functions of the TP. Among the numerous land surface processes potentially impacted by climate warming, the effect of vegetation greenness on surface energy balance is one of the most critical, but has been long ignored. In this study, we investigated the spatial and temporal patterns of land surface albedo (LSA) on the TP and evaluated the vegetation greenness in relation to patterns of LSA. We found that LSA has been decreasing in most of the vegetated grasslands on the TP from 2000 to 2013, as compared to a flat trend for desert area. The regions where LSA has been decreasing were spatially correlated to areas of increased vegetation greenness. Along rising altitude, LSA decreasing rate exhibited an overall decreasing trend. Across the TP, elevated vegetation greenness in grasslands acted as a primary factor pulling down LSA. The driving effects of vegetation greenness on LSA vary with grassland types, as revealed by a more significant relationship between vegetation greenness and LSA for the sparsely vegetated zone (i.e. steppe) than the more densely vegetated zone (i.e. meadow). Furthermore, the driving effect of vegetation greenness on LSA exhibited an obvious dependence on altitude as effects with rising altitude were relatively strong up to 3000 m, then weakened from 3500 m to 5000 m, and then the effects again increased from 5000 to 6000 m. The growing season LSA trend revealed in this study emphasizes the need to give greater attention to the growing season LSA flux in future surface energy balance studies.

Keywords: albedo, alpine, altitude, energy balance, global warming, land surface process

1. Introduction

The Tibetan Plateau (TP) has fundamental ecological significance to China and Asia. In addition to serving as a headwater region for several of the major rivers in China and southern Asia, the TP is the ‘heat pump’ for Asian monsoon. It exerts a large thermal forcing on the mid-troposphere over the middle-latitudes of the Northern Hemisphere during spring and summer to regulate regional energy cycles (Ye and Gao 1979). The thermal forcing effectively enhances the Asian summer monsoon and modulates its variability (Wu et al 2012). The surface heating also triggers vigorous deep convections over the TP (Yang et al 2004) that greatly enhance troposphere-stratosphere exchanges of water vapor (Fu et al 2006). Due to its existence, the thermal difference between the Pacific Ocean and Asian continent is amplified, and consequently, the eastern Asian monsoon is pushed deep into northern China.

Meteorological observations (Liu and Chen 2000) and climate models (Chen et al 2003) have indicated that climate
warming on the TP has been more pronounced than throughout the rest of the northern hemisphere in recent decades. The dramatic warming by approximately 0.4 °C per decade (Tao et al. 2013) has extensive ecological and environmental consequences (Yang et al. 2014). Special research attention has been paid to changes in patterns of vegetation and plant growth as well as changes in glacier and snow cover (Yao et al. 2012, Tao et al. 2014). For example, enhanced vegetation productivity (Piao et al. 2006), advanced green-up date (Zhang et al. 2013), and glacier shrinkage have now been widely reported (Yao et al. 2012).

The shifted vegetation growth and glacier coverage trends have resulted in consequences for a range of terrestrial land surface processes that regulate exchanges of mass and energy between biosphere and atmosphere, which can greatly affect the ‘heat pump’ effects of the TP on eastern Asian monsoon. Among the series of land surface processes, the one most tightly related to energy exchange is the land surface albedo (LSA).

The LSA is the ratio of the upwelling radiant energy relative to the downwelling irradiance incident upon a surface (Stroeve et al. 2006). In the physical climate system, albedo determines the surface radiation balance and affects the surface temperature (Sellers et al. 1996). In ecological systems, albedo can affect physical and physiological processes of ecosystems, such as energy balance and evapotranspiration by regulating the microclimate conditions of plant canopies and their absorption of solar radiation (Wang et al. 2001, 2002). Albedo can modulate regional climate events such as drought, moisture flux convergences and rainfall (Knorr et al. 2001). Since the physical climate system is sensitive to surface albedo, ecosystems may have significant feedbacks to projected climate scenarios through albedo changes. As such, impacts of climate change on surface albedo and ecosystem feedbacks are of high significance for climate studies, especially in those systems which are highly sensitive to climate change, such as the TP.

A variety of land surface processes can affect LSA. Previous studies analyzing the driving mechanisms behind albedo dynamics on the TP have primarily focused on winter snow effects (Wu et al. 2012). In addition to snow effects, the LSA is closely related to vegetation growth, ecosystem dynamics, and landscape conversion (Betts 2000, Bounoua et al. 2002). For example, local surface temperature would fall in response to decreased surface radiation loading associated with deforestation if radiation were the only energy transfer process involved (Lee et al. 2011). Increased grassland albedo could also decrease radiative forcing (Lenton and Vaughan 2009, Ridgwell et al. 2009, Doughty et al. 2011). Therefore, increased vegetation coverage in response to climate warming results in higher absorption of shortwave solar radiation, thereby decreasing LSA which may lead to further climatic warming (Wookey et al. 2009).

Vegetation driven albedo changes have been well-studied in the tundra of the Northern Hemisphere (McFadden et al. 1998, Chapin et al. 2000, Eugster et al. 2000, Thompson et al. 2004), where widespread shrub species have been progressively moving into tundra (Jia et al. 2003). A tundra-to-shrubland transition or a tundra-to-forest transition then results in an increase in surface heating (Sturm and Douglas 2005). Compared to the centuries taken by the tundra-to-forest transition and the multi-decades needed for the tundra-to-shrub transition, changes in vegetation greenness in the TP driven by climate flux may take place in a matter of years. For example, vegetation on the TP has experienced dramatic changes in the past few decades alone (Zhang et al. 2013).

In addition, the TP receives much higher solar irradiance than in lower elevation zones and the growing season solar irradiance is stronger than in non-growing season (Wang et al. 2004). Therefore, the energy balance on the TP during summer has special significance. To date, the critical driver of the energy balance, the growing season LSA (GSAlbedo), and its spatial and temporal patterns on the TP are not well understood and the mechanism driving changes in GSAlbedo have yet to be fully investigated.

The objectives of this study were set to (1) explore the spatial and temporal patterns of GSAlbedo on the TP; (2) test the hypothesis that changes in patterns of vegetation are driving contemporary changes in GSAlbedo. An accurate LSA estimation on the TP is necessary for predicting its ‘boost’ function on East Asian monsoon. A comprehensive understanding of these processes and their consequences is critical to quantify the impacts of land process feedbacks to climate.

2. Method

2.1. Study area

The TP extends from sub-tropical to mid latitude regions, spanning over 25 degrees of longitude. Known as ‘the roof of the world’, it has an average elevation of more than 4000 m (Zheng 1996). The spatial domain of the study area ranges within 26.5°–39.5° N, and 78.3°–103° E, and covers an area of 2.57 × 106 km2 (figure 1(A)). The mean temperature for the coldest month and the warmest month is approximately −10 °C and 10 °C, respectively (Zhang et al. 1982). The temperature and precipitation follow distinct decreasing gradients from southeastern to northwestern TP (Zheng 1996). Due to its high elevation, the TP surface receives higher solar radiation than other inland areas of China, and the growing season solar radiation is much higher than in winter season (Ye and Gao 1979).

From south-east to north-west, three main types of grasslands (i.e., meadow, steppe, and desert steppe) are distributed (Tao et al. 2013). Sparse vegetation grow in the arid north-west high mountains with very low values of growing season NDVI (GSNDVI) (<0.15) (figure 1(B)). A relatively small area of evergreen forest grows in the warm and wet southeast, where the highest GSNDVI is observed (figure 1(C)). In this study, we excluded forest and only considered grassland vegetated areas due to forests’ complex relationship with surface albedo (figure 1(B)). In addition, grasslands cover a majority (about 70%) of the TP (Piao et al. 2011). The GSNDVI value for grassland ranges between
0.1 and 0.8 on the TP (figure 1(C)), and exhibits a decreasing gradient from southeast to northwest (figure 1(C)).

2.2. Data source

2.2.1. Vegetation type data. The vegetation type data was based on the China vegetation map (Liu et al. 2003) obtained from Data Center of Resources and Environmental Sciences, Chinese Academy of Sciences. Six grassland types were extracted: alpine meadow (occupying 47.05%), alpine steppe (30.98%), alpine desert steppe (7.41%), alpine meadow steppe (4.21%), low land meadow (6.74%), and temperate steppe (3.61%) (figure 1(B)). Based on their occupied areas and distribution patterns, alpine meadow, alpine steppe and alpine desert steppe were treated as the dominate grassland types in this study.

2.2.2. MODIS data. Growing season is composed of May, June, July, August, and September (Zhang et al. 2013). Both GSNDVI and GSA data used in this study were products of the Moderate Resolution Imaging Spectroradiometer (MODIS) (Justice et al. 1998, Schaaf et al. 2002). The MODIS NDVI and albedo products have been previously validated on the TP (Qin et al. 2011). We used the quality-filtered function of the subssetting tool (http://daac.ornl.gov/MODIS/modis.shtml) to exclude snow and cloud covered pixels. The MODIS Terra NDVI data were 16-days composite products (MOD13Q1) (Blok et al. 2011), with a 250 m spatial resolution. To avoid the effects of the Terra MODIS sensor degradation on our analysis (Wang et al. 2012), we repeated the analysis on the relationship between the mean GSAbedo and the mean GSNDVI using the MODIS Aqua data. During the growing season, the set of data includes a total of 10 NDVI images produced between day of year (DOY) 129 and 273 (8th May–30th September). We extracted the maximum monthly values from the adjacent two NDVI images in each month. For example, the data for the month of May was extracted as the maximum value between DOY 129 and DOY 145 for each pixel.

For albedo data, we used the MODIS shortwave white sky albedo product of the collection5 (MOD43A3). The MODIS tool provides black-sky for direct and white-sky albedo for isotropic diffuse radiation at local solar noontime. Between the two products, the white sky can more closely reflect the true condition of the surface land cover and is more stable (Schaaf et al. 2002). Since this study was aimed to explore the albedo pattern driven by vegetation, we relied on white sky albedo product. It has a spatial resolution of 500 m and an eight-day temporal resolution. Similar to handling NDVI data, we used the quality-filtered function of the subssetting tool to restrict the analysis to full bidirectional reflectance distribution inversion data (highest quality) (Blok et al. 2011). Then using the 20 albedo images produced between day of year (DOY) 121 and 273 (1st May–30th September), we compiled the monthly minimum albedo for each month. For example, the minimum May albedo was extracted as the minimum values among values of DOY121, 129,137 and 145 for each pixel.

Figure 1. (A) The grassland vegetation zones. (1): Alpine meadow; (2): Alpine steppe; (3): Alpine desert steppe; (4): Alpine meadow steppe; (5): lowland meadow; (6): temperate steppe. (B) The GSNDVI (May–September) during 2000–2013. (C) The average GSAbedo (May–September) on the Tibetan Plateau during 2000–2013.
2.3. Quantitative metrics and analysis

Theoretically, the annual minimum GSAlbedo should occur simultaneously with the annual maximum GSNDVI. In using the linear regression model to explore the driving effects of vegetation greenness on LSA, we matched the minimum GSAlbedo and average GSAlbedo with the annual maximum GSNDVI and annual average GSNDVI, respectively. The annual GSAlbedo\_min and the GSAlbedo\_\text{mean} were calculated using the formula (1) and (2), respectively:

\[
\text{GSAlbedo}_{\text{min}} = \text{Min} \left( \text{Albedo}_{\text{min}, i} \right) \quad (1)
\]

\[
\text{GSAlbedo}_{\text{mean}} = \frac{1}{5} \times \sum_{i=1}^{5} \text{Albedo}_{\text{min}, i} \quad (2)
\]

where Albedo\_min was the minimum albedo value of each month, \( i \) denotes the month of May, June, July, August, and September.

Similarly, the annual GSNDVI\_max and GSNDVI\_\text{mean} were calculated by the formulas (3) and (4).

\[
\text{GSNDVI}_{\text{max}} = \text{Max} \left( \text{NDVI}_{\text{max}, i} \right) \quad (3)
\]

\[
\text{GSNDVI}_{\text{mean}} = \frac{1}{5} \times \sum_{i=1}^{5} \text{NDVI}_{\text{max}, i} \quad (4)
\]

where NDVI\_max was the maximum NDVI of each month in a year, \( i \) represents the months of May, June, July, August, and September.

To confirm the driving effects of vegetation greenness on LSA, we also analyzed the LSA trend for desert on the TP in the same period.

We further explored the altitude dependent pattern of the relationship between GSNDVI and GSAlbedo. We calculated the average GSNDVI and GSAlbedo for each 500 m band and then evaluated their correlation within each 500 m elevation band. We defined the trend slopes of GSNDVI and GSAlbedo as their change rates, which can be increasing or decreasing rate.

3. Results

3.1. The spatial and temporal patterns of GSAlbedo and GSNDVI

In exploring the spatial patterns of GSAlbedo and GSNDVI, we used the 14-year average (2000–2013) of the monthly maximum NDVI and the monthly minimum albedo. With alpine meadow distributed in the southeastern plateau and alpine steppe in the northwestern plateau (figure 1(A)), vegetation greenness decreased along the gradient from east to west (figure 1(B)). The highest GSNDVI value for alpine meadow reached approximately 0.8 and the lowest GSNDVI value for alpine steppe was around 0.2 (figure 1(B)).

In accordance with the greenness gradient, the GSAlbedo increased from approximately 0.1–0.4 along the gradient from southeast to northwest. The GSAlbedo of alpine meadow and alpine steppe fell in the range of 0.1–0.2 and 0.2–0.4, respectively (figure 1(C)).

The change trends of the GSAlbedo and the GSNDVI during 2000–2013 had strong spatial heterogeneities (figure 2). The regions with significantly increasing GSNDVI were located mostly in the north (figure 2(C)). Accordingly, the GSAlbedo declined mostly in areas where the GSNDVI increased (figures 2(A), (B)). A large proportion of the grassland vegetated areas experienced decreased GSAlbedo. The regions with significant GSAlbedo change matched spatially well with those experiencing significant GSNDVI change and were mostly distributed in the north (figures 2(C), (D)).

We divided the entire TP grasslands into 500 m elevation bins, among which, the one with elevation of 5000–5500 m occupies the largest area, followed by the bin with elevation of 5500–6000 m (figure 3(A)). The temporal patterns of GSNDVI varied with elevation and exhibited an increasing trend in areas above 2000 m. Among the elevation bins, the increases in GSNDVI reached statistically significant levels in bins between 2500 m and 4000 m. Along increasing elevation, the temporal change rate of the GSNDVI (i.e. the increasing magnitude) exhibited a declining trend (figure 3(B)). The GSAlbedo has been decreasing across the grassland vegetated TP since 2000. The decreasing trends were statistically significant in areas from 1000 m to 6000 m, and failed to reach a statistically significant level above 6500 m (figure 3(C)). Along rising altitude from 1000 m to 4000 m, the decreasing rate followed an overall decreasing trend.

3.2. The coordinated inter-annual change of the GSAlbedo and GSNDVI

For the entire grassland area with the TP, along with an increasing annual mean GSNDVI as shown by both MODIS Terra and Aqua data, the mean GSAlbedo followed a decreasing trend (figure 4(A)) and the trends were both statistically significant at 95% confidence levels. From 2000 to 2013, the mean GSNDVI increased from 0.361 in 2000 to the highest level of 0.388 in 2012 and then saw a minor dip in 2013. Accordingly, the mean GSAlbedo decreased from 0.181 in 2000 to the lowest level of 0.161 in 2012 and then picked up a sudden spike value of 0.173 in 2013. The maximum GSNDVI followed a flat trend from 2000 to 2008 and then exhibited a quick increasing from 2008 to 2010 (figure 4(B)). While the minimum GSAlbedo followed a consistent decreasing trend from 2000 to 2012, as the value decreased from 0.170 in 2000 to 0.155 in 2012.

We further quantified the driving effects of vegetation greenness on the LSA by exploring their potential linear correlation. As shown in the insets of figures 4(A) and (B), vegetation greenness exhibited a statistically significant and negative relationship with the LSA. The decreasing slope of the maximum GSNDVI against the minimum GSAlbedo (-0.617) was steeper than that of the mean GSNDVI against the mean GSAlbedo (-0.498).
The magnitude of the impact of changes in vegetation greenness on LSA varied with grassland type. The annual mean GSNDVI of alpine steppe and temperate steppe witnessed a significant increasing trend (figure 5(2) (6)). For the other four grassland types, the regression slopes indicated their overall increasing trends, but the increments failed to reach statistically significant levels (figure 5). Among the six grassland types, the mean rate of increase in GSNDVI was greatest in the temperate steppe, with a slope of 0.002 (figure 5(6)).

The annual maximum GSNDVI followed increasing trends for all the grassland types, but alpine meadow steppe (figure 6). The increments reached significant levels for alpine steppe, alpine desert steppe, and temperate steppe (figure 6 (2) (3) (6)). The increasing magnitude was greatest for temperate steppe, with a slope of 0.002 (figure 6(6)). The minimum GSAIbedo exhibited decreasing trends for the six types of grasslands (figure 6). For alpine steppe, alpine meadow steppe, low land meadow, and temperate steppe, the trends reached statistically significant levels. For desert areas, the minimum GSAIbedo followed a flat trend from 2000 to 2013 (P value = 0.529) (figure 7).

For the three dominant grasslands (i.e. alpine meadow, Alpine steppe, and alpine desert steppe), the relationships between the mean GSAIbedo and the mean GSNDVI reached significant levels at 90% confidence levels. The relationship was strongest for alpine steppe ($R^2 = 0.543$, $P$ value = 0.003), followed by alpine desert steppe ($R^2 = 0.526$, $P$ value = 0.003), then lastly alpine meadow ($R^2 = 0.216; P$ value = 0.094) (table 1). The same correlation significance order was applicable to the relationships between the minimum GSAIbedo and maximum GSNDVI.

The driving effects of GSNDVI on GSAIbedo, as indicated by their correlations, exhibited an obvious altitudinal gradient. Along a rising elevation from foothills to 3500 m, the driving effect of vegetation greenness on albedo change followed an enhanced trend, then it decreased along rising altitude from 3500 to 5000 m, and then shifted to an enhanced trend again from 5000 to 6000 (figure 8).

4. Discussions

4.1. The spatial and temporal patterns of GSAIbedo

Spatially, across the grassland vegetated areas on the TP, the GSAIbedo exhibited an increasing pattern from southeast to
northwest. Along a temporal dimension, the GSAIbedo followed an obvious decreasing trend during our study period. Vegetation greenness exerted a tight control on the GSAIbedo as demonstrated by its regulation on the spatial and temporal patterns of the GSAIbedo. Furthermore, the GSAIbedo for the desert area exhibited a flat trend in the same period, which lends further support on the driving effects of vegetation greenness on the GSAIbedo.

Furthermore, the temporal trend of the GSAIbedo expressed a clear altitude dependent pattern, which can be explained by the altitude dependent climate change (Tao et al. 2013). Both day and night temperatures have seen greater increasing rates in lower elevation zones in the past decades. The alpine vegetation system is strongly constrained by low temperature. Therefore, elevated temperature can release the temperature constraint, and a higher increasing rate means a greater magnitude of facilitation on vegetation growth. Consequently, the altitude dependent vegetation growth dynamics resulted in the altitude dependent albedo change trend.

The regulation of vegetation greenness on albedo varies with grassland types. Among the three typical grassland types, alpine steppe possesses a stronger relationship between vegetation greenness and surface albedo than alpine meadow does. The reason is because the vegetation coverage of alpine meadow is higher than that of alpine steppe. This finding lends further support to that the relationship between vegetation greenness and surface albedo is stronger for sparse vegetation community than for the dense one (Lee et al. 2011). One noteworthy point is that in sparse areas, an equal amount of vegetation greenness increase will result in a larger spectral change compared to more densely vegetation areas. Furthermore, the stronger correlation between the minimum GSAIbedo and the maximum GSNDVI than that between the mean GSNDVI and the mean GSAIbedo illustrates a closer time match of the previous pair than the later one.

The regulation of vegetation greenness on albedo also varies with altitude. Along rising elevation from foothill to 3500 m in elevation, vegetation height gradually decreases and vegetation composition and structure become increasingly simple. For those tall and complex structured vegetation, their effects on surface albedo will be more likely to be diluted due to canopy reflection. As a result, the contribution of vegetation greenness to albedo dynamics is enhanced from foothills to 3500 m in elevation. When elevation further climbs from 3500 to 5000 m, the contribution of
vegetation greenness to albedo change weakens as vegetation becomes sparser and other factors, such as soil moisture, would co-act with vegetation to influence albedo dynamics. Furthermore, above 3500 m, the effect of year-around snow coverage begins to emerge. When elevation reaches above 5500 m, vegetation becomes scarce and patchily distributed and NDVI can reach only about 0.15, which is similar to bare ground and its effects on surface albedo are almost negligible.

4.2. Implications of GSAlbedo dynamics on the Tibetan PLATEAU

Regarding impacts of environmental changes on surface albedo on the TP, previous research has mostly focused on snow or ice induced consequences, which were more related to the non-growing season period. In this study, we found that the GSAlbedo displays high inter-annual variation, which can be critical in driving surface energy balance. Over the grassland vegetated areas of the TP, the mean GSAlbedo decreased from 0.18 to 0.16, equal to a decrease of 11%, during the time period from 2000 to 2012. The minimum GSAlbedo followed a more consistent decreasing trend from 0.17 in 2000 to 0.154 in 2012 and decreased 9%. The over 10% decrease in albedo on the TP can significantly change the TP energy balance by modifying its thermal forcing effect.

The effect of albedo change on energy balance can be amplified on the TP due to its greater solar irradiance than in the Arctic zone and other low elevation inland areas. In addition, growing season solar irradiance is higher than that in non-growing season. In one typical semi-desert region on the TP, the solar irradiance exhibits obvious seasonal dynamics, ranging from the lowest value of approximately 520 W m\(^{-2}\) in winter to the highest value of approximately 1000 W m\(^{-2}\) in

![Figure 5. The trends of the mean GSNDVI and mean GSalbedo for each grassland type.](image-url)
summer (Wang et al. 2004). In the region around Northern 50 °N, the lowest solar radiation in winter is approximately 200 w m$^{-2}$, though the highest values can reach almost 900 w m$^{-2}$ due to its long duration of irradiance. In the Northern 75 °N zone, the lowest radiation value is zero and

**Table 1.** Regression coefficients of mean GSAIbedo against GSNDVI, and of the minimum GSAIbedo against GSNDVI for the three typical grasslands.

| Grassland Type          | Mean GSAIbedo and GSNDVI | Min GSAIbedo max GSNDVI |
|-------------------------|--------------------------|--------------------------|
|                         | Slope        | R$^2$  | P   | Slope       | R$^2$  | P   |
| Alpine meadow           | $-0.182$     | 0.216  | 0.094 | $-0.086$    | 0.035  | 0.519 |
| Alpine steppe           | $-0.521$     | 0.543  | 0.003 | $-0.434$    | 0.576  | 0.002 |
| Alpine desert steppe    | $-0.710$     | 0.526  | 0.003 | $-0.193$    | 0.506  | 0.004 |

**Figure 6.** The trends of the annual minimum GSAIbedo and annual maximum GSNDVI for each grassland type from 2000 to 2013.

**Figure 7.** The minimum GSAIbedo trend for desert area on the Tibetan Plateau from 2000 to 2013.

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the highest value can reach about 600 W m\(^{-2}\) (Sturm and Douglas 2005). The average solar irradiance over the entire China is about 170 W m\(^{-2}\) (Yang et al 2014). It was predicted that the vegetation coverage change accompanying the tundra-to-shrubland transition would increase July mean air temperature by +1.5 to +3° in the Arctic (Chapin et al 2000, Eugster et al 2000). Due to the greater value of solar irradiance, reduced LSA means more irradiance absorbed, which would further exacerbate climate warming on the TP.

Enhanced vegetation coverage leads to a lower surface albedo and more energy absorbed (Bonan et al 1992, Betts 2000, Barnes and Roy 2008), which improves the growing conditions for alpine grasslands on the TP where temperature is the main constraining factor. Consequently, vegetation will grow better, further lowering the albedo, and so on. Glaciers have been reported to be shrinking on the TP at an unprecedented rate which further lowers albedo (Yao et al 2012). Added together, the terrestrial amplification effects under increased vegetation greenness to the ice-albedo-feedback mechanism, we have strong reason to expect that this amplification effect will accelerate the ongoing warming on the TP.

Other processes can affect surface albedo dynamics. For example, during our study period, the lake areas have expanded significantly, which can increase surface albedo to some extent (Song et al 2013). Conversely, black carbon in snow reduces albedo and thereby change radiative forcing (Wang et al 2011, Hadley and Kirchstetter 2012, Xu et al 2012, Lee et al 2013). Other snow impurity contents, for example pollution related, can also change albedo and radiative forcing (Warren 2013). Surface whitening caused by urbanization has also been estimated to result in a potential radiative forcing of \(-0.17\ W\ m^{-2}\) (Hamwey 2007). In addition, the surface sensible heat has decreased significantly in the past decades due to a significant decline in radiative convergence (Yang et al 2014).

Given the many interacting factors, predicting whether the self-amplifying feedback will occur involves large uncertainty. More solar heating absorbed can drain out soil moisture, thereby altering soil microbial communities and moisture levels (Sturm et al 2005), consequently adding to the water stress on plant growth. Other related ecosystem processes, such as nutrient cycling and nutrient availability, would also change with the soil heating condition, thereby possibly limiting plant growth and lowering NDVI.

The results of our study highlight the importance of monitoring ongoing enhanced GSNDVI caused GSAlbedo change closely. As an entirety, the thermal forcing of the TP has been enhanced, which would bring about corresponding changes to the Eastern Asian monsoon. It has been reported that weakened surface heating and atmospheric heating under climate changes (Zhu et al 2007, Yang et al 2011) can affect summer precipitation downstream in China (Ding et al 2009, Duan et al 2013). A clear understanding of the GSAlbedo dynamics would improve our projections of the ‘heat pump’ effects of the TP on Eastern Asian monsoon.

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**Figure 8.** The altitude dependent correlations between the maximum GSNDVI and the minimum GSAlbedo.
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