Amplified water vapour feedback at high altitudes during winter

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ABSTRACT: During the last five decades, the Tibetan Plateau has experienced a warming trend of 0.4°C/decade in winter, which is at least twice that of any other season. Some studies have suggested that this anomalous winter warming is caused, in part, by the observed increases in near-surface water vapour and its amplifying effect on the surface longwave downward radiation (LDR). This study uses observations of surface-specific humidity ($q$) and temperature as input to a one-dimensional radiative transfer model to assess the influence of lower atmospheric increases in water vapour on surface LDR, and the sensitivity of this process to different elevations and seasons on the Tibetan Plateau. The results from three idealized experiments are examined based on realistic atmospheric column profiles of temperature and moisture. They show that when an equal mass of water vapour is added into the atmospheric boundary layer during winter, a substantially greater increase ($8\times$) in LDR is found at the high-elevation site relative to the low-elevation site. During summer, the LDR increases are much smaller as are the differences between the two sites. Experiments, where both $q$ and temperature are increased, suggest that the influence of temperature changes on LDR is much greater than those caused by changes in $q$ in all cases, except for the high-elevation-winter case when the opposite is true. These results provide further evidence for the possibility of a strong modulation of surface LDR caused by increases in atmospheric water vapour in high altitude regions (>3000 m) during the cold season. Copyright © 2012 Royal Meteorological Society

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1. Introduction

As a greenhouse gas in the Earth’s atmosphere, water vapour causes the largest absorption of planetary thermal infrared radiation, also called the longwave radiation. However, unlike other important greenhouse gases, tropospheric water vapour produces a rapid climate feedback because of its short residence time (weeks), which is controlled by the climate of the troposphere. Therefore, as the troposphere warms in response to increases in anthropogenic greenhouse gases, it is expected that water vapour will increase in the atmospheric column and produce a positive feedback on the initial warming. This water vapour feedback mechanism is predicted to substantially amplify the global scale warming initiated by increases in the greenhouse gas forcing caused by anthropogenic emissions (Hall and Manabe, 1999; Held and Soden, 2000).

Surface observations (Wang et al., 2001; Dai, 2006) and radiosonde measurements (Zhai and Eskridge, 1997; Ross and Elliott, 2001) during the last few decades have indicated increased moistening of the lower troposphere. Since the 1980s, satellite measurements have also shown increases in water vapour mass in the whole atmospheric column (Wentz and Schabel, 2000; Trenberth et al., 2005) and the upper troposphere (Soden et al., 2005). These increases in surface and column water vapour have been attributed to the anthropogenic greenhouse gas warming (Santer et al., 2007; Willett et al., 2007). The atmospheric concentration of water vapour, which is highly variable in space and time, generally decreases with height because the equilibrium water vapour concentration is proportional to air temperature. For future increases in the greenhouse gas forcing, global climate models suggest that the fractional increase in the water vapour mass will be greater in the upper atmosphere (Soden et al., 2005). Since the water vapour absorptivity is proportional to the logarithm of its concentration, the strength of the positive water vapour feedback mechanism should be substantially greater in the upper atmosphere because it is the fractional change, and not the absolute change, in the water vapour mass that governs the strength of this feedback mechanism (Held and Soden, 2000).

Several studies have also suggested that water vapour absorptivity has strong sensitivity to changes in water vapour concentrations during the cold season in certain regions of the Earth’s surface, which include the polar (Chen et al., 2006b, 2011) and the high mountain (Rueckstuhl et al., 2007; Rangwala et al., 2010) regions. One such region is the Tibetan Plateau, which is occasionally referred to as the ‘third pole’ because of the large
During the latter half of the 20th century, this region has warmed by approximately 0.2 °C/decade (Liu and Chen, 2000; Chen et al., 2006a; Rangwala et al. 2009). However, there is a significant seasonal variability in the warming rates on the Plateau. Liu and Chen (2000) reported that the warming rate in winter was about twice as large as the annual mean warming rate, a result which has been supported by subsequent studies (Du et al., 2004; Chen et al., 2006a; Liu et al., 2006; You et al., 2007; Rangwala et al., 2009). During winter, there are also greater increases in the daily minimum temperature relative to the daily maximum temperature (Liu et al., 2009; Rangwala et al., 2009).

Rangwala et al. (2009, 2010) hypothesized that the large winter warming trend in the Tibetan Plateau could, in part, be explained by observed increases in surface humidity (q) and its influence on the longwave downward radiation (LDR) at the surface. This hypothesis is supported by LDR–q relationship found in observations (Ruckstuhl et al., 2007) and climate models (Rangwala et al., 2010) that show LDR is more sensitive to changes in moisture when q is below 5 g kg\(^{-1}\); a condition that occurs most frequently during cold seasons and at higher elevations. These LDR increases should enhance the surface warming, particularly the daily minimum temperature. This study is motivated by the desire to examine this hypothesis further under a controlled environment where the influence of q on LDR could be more selectively evaluated.

This article presents results from idealized experiments conducted using a one-dimensional (1-D) radiative transfer model to assess the influence of lower atmospheric increases in water vapour on surface LDR, and the sensitivity of this process to differences in surface elevation and seasons. These experiments consider ‘clear sky’ atmospheric conditions that are associated with the Tibetan Plateau and its vicinity, and they are informed by observed increases in q and surface air temperature in these regions. The next section briefly describes the radiative transfer model and discusses the motivation and design of the different experiments. The results from these experiments confirm a strong influence of q on LDR during winter at high surface elevations. These are presented and discussed in Section 2. Conclusions are presented in Section 3.

2. Methods

The longwave standalone version of the Rapid Radiative Transfer Model (RRTM; https://rtweb.aer.com) is used to conduct the experiments. This model utilizes the correlated-k approach to calculate instantaneous radiative fluxes and cooling rates at different levels within an atmospheric column for user-specified atmospheric profiles of water vapour mixing ratios and temperatures. The k-distributions used for radiative flux calculations are obtained directly from a line-by-line radiative transfer code, which has been extensively validated against observations. These calculations could be made in 16 contiguous spectral bands in the long wave (10–3250 cm\(^{-1}\)). Further detail about this model can be found in Mlawer et al. (1997) and Iacono et al. (2000).

The motivation for this study is to determine whether there are differences in the LDR–q sensitivity between high- and low-elevation sites at the same latitude for both winter and summer. To facilitate this, mean atmospheric column profiles of q and temperature for the months of June and December, 2000, are extracted from the ERA-40 Reanalysis Data available at the National Center for Atmospheric Research (https://dss.ucar.edu/pub/era40), for a high (5138 m) and a low (114 m) elevation sites in the vicinity of the Tibetan Plateau (Figure 1). The main reason for using the ERA-40 Reanalysis is that it has assimilated some of the available surface climate observations (e.g. temperature, moisture and wind) from the Tibetan Plateau region, whereas NCEP Reanalysis has not (You et al., 2010). Therefore, the ERA-40 Reanalysis should provide a better representation of the atmospheric profile for this region.

Figure 1. The Tibetan Plateau and its vicinity. The contours show surface elevation (m). The ‘+’ sign shows the location of the experimental regions: high (left; 30°N, 87°E; 5138 m) and low (right; 30°N, 105°E; 114 m).
profiles for the four cases: (1) high-elevation-December, (2) low-elevation-December, (3) high-elevation-June and (4) low-elevation-June are extracted for 29 selected sigma levels in the vertical and are shown in Figure 2. These profiles are used as the control conditions on which the experimental perturbations in $q$ and temperature are introduced into specific sections of the atmospheric column. For a majority of the experiments, the perturbations are only made in the atmospheric boundary layer, which for the purpose of these experiments is assumed to contain all layers with sigma $>0.9$ and, therefore, included the bottom six layers in our experimental data. The thickness of this layer is approximately 700 m. The experiments are performed for clear sky conditions only and for the ‘mid-latitude’ setup of the model code. An example input file for the model is provided in the Supporting Information. All results presented here, related to changes in surface LDR, are differences between the surface LDR simulated for the experimental and control conditions.

Finally, three specific sets of experiments (Table I) are performed to test three specific questions:

(1) What is the change in surface LDR for equivalent increases in water vapour amounts in the boundary layer? This experiment tests the sensitivity of water vapour absorptivity to similar increases in water vapour mass in the boundary layer for the four cases. In this experiment, $q$ is increased by 0.4 and 0.2 g kg$^{-1}$, in each layer within the boundary layer, at the high- and low-elevation sites, respectively, based on the fact that the boundary layer (as defined) at the low-elevation site has twice the amount of air mass relative to the high-elevation site, even though the geometric thickness is about the same. In other words, for each cubic meter of air at these two elevations we add approximately 0.25 g of water vapour in the experimental conditions.

(2) How do observed changes in $q$ and temperature affect the surface LDR when the changes are first introduced separately and then at the same time? Increases in $q$ of 0.4 g kg$^{-1}$ and temperature of 2°C have been observed in the Tibetan Plateau, between 1961 and 2000, and are introduced within the boundary layer for the four cases.

(3) How does the atmospheric level at which $q$ increases are made affect surface LDR? This question is examined by changing $q$ in different sections of the atmospheric column including (1) the boundary layer, (2) the whole column, (3) the whole column minus the boundary layer and (4) the upper atmosphere (<200 mb). For the Tibetan Plateau, ERA-40 Reanalysis does not show any increases in $q$ between 1961 and 2000, which is inconsistent with observations. Therefore, to facilitate physically based changes in $q$ in the atmospheric column, this experiment uses the
ensemble mean of fractional changes in $q$ simulated during the 21st century by two global climate models, GFDL2.0 and GFDL2.1 (http://data1.gfdl.noaa.gov; SRES A2 only), for these two sites. These fractional changes are shown in Figure 3.

3. Analysis of LDR–$q$ relationship

3.1. Experiment 1: Equivalent total water vapour mass increase in the boundary layer

This idealized experiment tests whether similar masses of water vapour added into the boundary layer produce significantly different responses in LDR at the high- and low-elevation sites. The analysis was done for both December and June. To approximate equal increases in water vapour at these two sites, $q$ is increased by 0.4 and 0.2 g kg$^{-1}$ at the high- and low-elevation sites, respectively, based on the assumption that the boundary layer at the low-elevation site has twice the amount of air mass relative to the high-elevation site.

Figure 4 shows the changes in the LDR response for the four cases. For December, the LDR increase at the high-elevation site (5.64 W m$^{-2}$) is eight times greater than that at the low-elevation site (0.71 W m$^{-2}$). For June, the increase in LDR at the high-elevation site is only three times greater than at the low-elevation site (1.20 W m$^{-2}$ vs 0.47 W m$^{-2}$). These results support earlier findings by Ruckstuhl et al. (2007) and Rangwala et al. (2010), which suggest that LDR is more sensitive to changes in $q$ at high elevations during winter and, therefore, predict a greater increase in LDR in high-elevation regions, relative to low lying regions for similar increases in lower atmospheric water vapour.

Comparable increases in $q$ have occurred at vastly different altitudes in the Tibetan Plateau. Rangwala et al. (2009) found observed increases in $q$ of 0.4 g kg$^{-1}$ between 1961 and 2000, during winter at different altitudes within the Tibetan Plateau. They also reported a winter warming of 2°C for that time period. For the next experiment, we introduce these increases in $q$ and temperature, both separately as well as simultaneously, and analyse their impact on LDR.

3.2. Experiment 2: Similar ‘$q$’ and ‘temperature’ increases in the boundary layer

Changes in temperature have a significant effect on longwave emission. For this experiment, we analyse both the separate and combined effects of increases in $q$ and temperature in the boundary layer on surface LDR. The experimental setup here is identical to that in Experiment 1 – with identical atmospheric profiles of $q$ and temperature for December and June – except that identical increases in $q$ of 0.4 g kg$^{-1}$ and temperature of 2°C, based on historical observations in the Tibetan Plateau, are introduced within the boundary layer at both the high- and low-elevation sites for December and June.

Figure 5 shows that when only $q$ is changed, the results are qualitatively similar to Figure 4 with the largest and most significant LDR increases at the high-elevation site during winter. However, the LDR increases at the low-elevation site for both December and June is twice that of Experiment 1 because of doubling of total water vapour mass (and thus a doubling of the number of absorbing molecules) in the boundary layer at the low-elevation site.

When only temperature is increased by 2°C in the boundary layer, the largest LDR increases (8.5 W m$^{-2}$) are found at the low-elevation site in June, and the smallest increases (3.0 W m$^{-2}$) at the high-elevation site in December. The sensitivity of LDR to changes in temperature is generally greater at the low-elevation site because of higher values of the absolute temperature there – or rather of the effective temperature of the longwave radiation that reaches the ground. Blackbody emission of longwave radiation is proportional to the fourth power of the absolute temperature, based on the Stefan–Boltzmann law, and, therefore, even with
similar increases in temperature at both the high- and low-elevation sites, there will be a greater increase in longwave emission at the low-elevation site. For example, based on this physical relationship, there will be a 25\% greater increase in the blackbody emission at the low-elevation relative to the high-elevation site for a 2° C increase in surface temperature in June.

When both $q$ and temperature in the boundary layer are increased simultaneously, the net increases in LDR are comparable for the high- and low-elevation cases for a winter (December) and a summer (June) month. The imposed increases in $q$ (0.4 g kg$^{-1}$) and T (2 K) in these experiments are based on the reported observed trends in these two variables between 1961 and 2000 for the Tibetan Plateau by Rangwala et al. (2009).

3.3. Experiment 3: ‘$q$’ increases in different sections of the column – ‘High-Elevation-December’ case only

In this experiment, the effect of adding increments of $q$ within the whole column and not just the boundary layer is examined. The motivation for this experiment is to assess the influence of $q$ increases in different sections of the atmospheric column on surface LDR. The $q$ increments at different pressure levels are estimated based on the mean ensemble increases found in two GCM simulations (GFDL 2.0 and GFDL 2.1; SRES A2 only), between the first and last decade of the 21st century, for the location associated with the high- and low-elevation sites. Figure 3 shows the percentage increases that are made to $q$ at different atmospheric levels. Here, only the results for the high-elevation-December case are presented.

Figure 6 shows the effect of adding increments of $q$ into different sections of the column which include (1) the boundary layer, (2) the whole column, (3) the whole column minus the boundary layer, and (4) the upper atmosphere (<200 mb). The largest increases in

LDR are found when $q$ is increased in the whole column (8.64 W m$^{-2}$), which is nearly twice as much as the increases found for the boundary layer only case (4.63 W m$^{-2}$). The LDR increase for ‘whole column minus boundary layer’ is similar to that of the ‘boundary layer’ experiment; and almost a negligible increase is found for the experiment where increases were made only in the upper layer.

The results from the other three cases – low-elevation-December, high-elevation-June, and low-elevation-June – are qualitatively similar to those for the high-elevation-December case. Proportionally, however, LDR increases for the ‘whole column’ cases relative to ‘boundary layer only’ are greater at the low-elevation site because the associated atmospheric column is deeper and optically thicker. Essentially, these experiments are sensitive to the quantity of absorbers in the column. Nonetheless, the ‘boundary layer’ moistening has the largest effect on surface LDR for the high-elevation site in December when these increases are similar to those arising from the moistening of rest of the column. For all other cases, the LDR increases for ‘whole column minus boundary layer’ are 30\% or more than those for the ‘boundary layer only’ cases.

4. Conclusions

Controlled experiments using a 1-D radiative transfer model (RRTM-LW) are conducted to assess the influence of lower atmospheric increases in water vapour on surface LDR, and the sensitivity of this process to differences in surface elevations and seasons. For high altitude regions, these experiments indicate high sensitivities of surface LDR to increases in lower atmospheric $q$ during winter. When equal masses of water vapour are added into the boundary layer during winter, the LDR increases eight times more at the high-elevation site relative to the low-elevation site. For summer, the LDR increases are much
smaller in magnitude at both sites, including the effects of elevation. On the basis of the observed trend for winter, even when only half the numbers of water vapour molecules are added in the boundary layer at the high-elevation site relative to the low-elevation site, there is a much larger increase in LDR at the high-elevation site.

Experiments where both q and temperature are increased suggest that the influence of temperature changes on LDR is greater than those caused by changes in q in all cases, except for changes during winter at high elevations when the opposite is true. Furthermore, when q increases in the whole column, LDR increases at least twice as much as when q is only increased in the boundary layer. Projections of future climate change indicate that temperature and q will increase in the whole atmospheric column which will cause greater increases in LDR than when these increases are just confined to the boundary layer. Nonetheless, the atmospheric moistening effects on surface LDR are proportionally greater for high-elevation-winter conditions than for all other combinations of surface elevations and seasons in the mid-latitudes.

It is generally accepted that changes in the upper tropospheric water vapour have the largest contribution to the positive water vapour feedback mechanism because they affect the top of the atmosphere longwave emission decreases. In these emissions will force the troposphere to warm to balance the energy budget at the top of the atmosphere. However, this warming response occurs on longer time scales of decades to centuries. On the other hand, the warming response of lower tropospheric moistening, particularly in regions where the air is optically under-saturated in water vapour absorption lines, is relatively instantaneous, and becomes important only during certain months of the year.

The results from this study provide further evidence for the possibility of a strong modulation of surface LDR caused by increases in atmospheric water vapour in high altitude regions (>3000 m) during the cold season, when the air is optically under-saturated in the longwave water vapour absorption lines. The changes in LDR are, however, controlled by both the LDR–q sensitivity and changes in q (Rangwala et al., 2010). Therefore, in certain cases, variable changes in q along the altitudinal gradients will be important in influencing LDR. The conditions for high LDR–q sensitivities in the lower atmosphere are expected to be present in high mountain regions and regions near the poles. And many of these regions may continue to exhibit the high range of sensitivities during the 21st century. We can, therefore, expect continued, and possibly enhanced, winter-time warming in these regions caused by a continued moistening of the atmosphere as a result of increases in the anthropogenic greenhouse gas forcing during this century.

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References


