

The physics of orographic elevated heating in radiative-convective equilibrium

Shineng Hu and William R. Boos

Department of Geology and Geophysics, Yale University, New Haven, CT 06511, USA

October 27, 2016

Submitted to *Journal of the Atmospheric Sciences*

Corresponding author:

Shineng Hu
210 Whitney Avenue
Kline Geology Laboratory
New Haven, CT 06511
shineng.hu@yale.edu
Tel: 203-909-3201

Abstract

Elevated heating of the atmosphere by large plateaus has been argued to influence regional climate in Asia and other regions, but the mechanisms that cause the troposphere to equilibrate at warmer temperatures over elevated terrain are not well understood. This paper quantitatively describes the physics that controls temperatures over elevated terrain in radiative-convective equilibrium (RCE). First, a cloud system-resolving model (CSRM) is used to simulate RCE states over surfaces with various elevations. Then a theory for the influence of surface elevation on temperatures in RCE is presented, and off-line radiative transfer calculations are used to show that the magnitude of the elevated heating effect is set by top-of-atmosphere radiative flux changes dominated by the sensitivity of the moist adiabatic lapse rate to surface height. Sensitivity functions obtained through these off-line calculations suggest that elevated heating is weaker in warmer climates, and additional CSRM simulations support this hypothesis. Under certain circumstances, even the sign of the elevated heating effect can change to produce cooler temperatures at a given pressure-level as the surface is lifted in RCE.

1. Introduction

Atmospheric temperature at a given height above sea level is expected to be warmer over an elevated surface than over a non-elevated surface (Molnar and Emanuel 1999), a phenomenon sometimes referred to as “elevated heating” (Wu et al. 2007). Elevated heating by large plateaus has been argued to drive circulations that influence climate in a broad range of locations. The Bolivian Plateau in South America (Rao and Erdogan 1989), the Colorado Plateau in North America (Tao et al. 1999), the Zagros Plateau in Iran (Zaitchik et al. 2007), and the Tibetan Plateau in Asia (Staff Members of Academia Sinica 1958) have all been claimed to provide thermal forcings for regional monsoons. Even the meridional gradient in surface height on Mars is thought to produce an annual-mean, thermally direct meridional circulation by providing elevated heating (Richardson and Wilson 2002).

Elevated heating has been studied most intensely over the Tibetan Plateau, where the great height and horizontal extent of that surface (4 km and 2.5 million km², respectively) were long held to drive the large-scale South Asian summer monsoon (Flohn 1968; Hahn and Manabe 1975; Molnar et al. 1993). The idea that the Tibetan Plateau’s elevated heating is strong enough to make it the dominant regional heat source has been recently challenged from both observational and modeling perspectives (Chakraborty et al. 2006; Boos and Kuang 2010; Nie et al. 2010; Ma et al. 2014), but the more fundamental question of what sets the magnitude of the Tibetan Plateau’s elevated heating still remains. More generally, how does elevated topography thermally interact with the atmosphere? This question is especially important in the tropics and subtropics, where topography can set the direction of the incident prevailing wind, and its answer is crucial to understanding the evolution of regional climate on geological time scales.

Answering this question requires a fundamental understanding of the physics of elevated heating by orography, and attaining such an understanding is the central goal of this study. We build on the work of Molnar and Emanuel (1999), who showed that temperatures at a given pressure level of the upper troposphere are warmer over elevated surfaces than over non-elevated surfaces in the theoretical state of radiative-convective equilibrium (RCE). In an RCE state, radiative cooling of the atmosphere is balanced by the latent heating of precipitating convection, and there are no temperature or moisture tendencies produced by large-scale flow. Although RCE is never achieved in the real atmosphere, it is nevertheless a conceptually useful state when theoretical and numerical models are employed to understand the radiative and convective physics that influence climate (Emanuel et al. 1994; Held et al. 2007; Romps 2011; Popke et al. 2013; Emanuel et al. 2014; Silvers et al. 2016). Indeed, the temperature of the RCE state of a particular column of the atmosphere is sometimes used as a forcing in idealized simulations of planetary-scale circulations (Lindzen and Hou 1988; Satoh 1994; Schneider 2006), providing a clean separation between large-scale dynamics and the radiative-convective physics that can be studied in the context of RCE. Here we use the RCE framework to study the mechanism of elevated heating, viewing this as a necessary first step in understanding the complete interaction between radiation, moist convection, and large-scale dynamics over elevated topography.

A variety of mechanisms have been argued to be responsible for the influence of surface elevation on the temperature of the overlying atmosphere, although these have been discussed by only a few authors. Flohn (1953) argued that radiative heating of the surface by solar radiation

increases as the surface is lifted because less scattering occurs in a thinner atmosphere, while radiative emission from the surface changes comparatively little. He reasoned that this should produce a warmer atmosphere over mountains than over adjacent non-elevated terrain. Molnar and Emanuel (1999, hereafter ME99) argued that this view is overly simplistic because the rate of atmospheric infrared cooling can change with surface elevation. They also noted that the vertically integrated mass of infrared absorbers might be reduced over an elevated surface, leading to cooling as the local greenhouse effect is reduced. This dependence of the column-integrated mass of infrared absorbers might produce a state-dependence of the elevated heating effect, exemplified by two limiting regimes (ME99): i) the nearly transparent (in the infrared) atmosphere of cold and dry climates would produce only a weak decrease of surface temperature as surface elevation rises, so temperatures at a given pressure-level of the upper troposphere warm strongly as the surface rises; ii) in warm and moist climates with optically thick atmospheres, a reduction in atmospheric mass cools the climate and eliminates nearly all of the elevated heating effect. In the latter regime, ME99 argued that elevated heating will be weak but still positive, with the rate at which surface temperature decreases as surface elevation rises limited to the moist adiabatic lapse rate. Although ME99 did find that surface temperature fell as the surface was lifted in their RCE simulations, they did not actually assess the roles played by changes in the vertically integrated mass of infrared absorbers, in clear-sky shortwave scattering (Flohn 1953), or in other processes.

Here we extend the work of ME99 by actually quantifying the influence of various processes on the magnitude of the elevated heating effect. This turns out to be more than a minor incremental effort because we discover that the largest influence on the magnitude of the elevated heating effect is a process not previously discussed: surface temperature drops when the surface is elevated because the moist adiabatic lapse rate decreases as surface elevation rises, leading to stronger longwave cooling of the upper troposphere. This lapse rate effect is similar to the lapse rate feedback discussed in the climate sensitivity literature for greenhouse gas-forced climate change (e.g. Cess 1975; Colman 2003), but is caused by the pressure-dependence of the moist lapse rate rather than by its temperature dependence. We also document the magnitude of several radiative mechanisms by which changes in longwave absorption by greenhouse gases influence the magnitude of elevated heating; these mechanisms have been discussed in the planetary science literature as ways in which the thickness of a planetary atmosphere can affect its equilibrium temperature (e.g. Halevy et al. 2009; Pierrehumbert 2010), but they do not seem to have been discussed in previous studies of elevated heating by orography. Our study also extends the efforts of ME99 in a methodological sense: we simulate RCE states over surfaces of different heights using a cloud-system resolving model (CSRМ) in which moist convection is represented explicitly rather than parameterized (ME99 used a single-column model with parameterized convection), and we use offline radiative transfer calculations to better understand the radiative-convective mechanisms that govern the CSRМ behavior.

This paper is organized as follows. The CSRМ used in this study and the experimental set-up are described in Section 2. We discuss the elevated heating effect simulated by this model in Section 3. We derive an index and present a theory for the elevated heating effect in Section 4. Off-line radiative calculations are conducted in Section 5 to quantitatively assess this theory. We further investigate the elevated heating effect across a wide range of climates in Section 6, where we also discuss the possibility that the elevated heating effect can change sign to become elevated

cooling in certain circumstances. We discuss implications and possible directions for future work in Section 7.

2. Numerical model details

We use a CSRM called the System for Atmospheric Modeling (SAM) version 6.3, which solves the anelastic equations of motion with prognostic liquid water moist static energy, total non-precipitating water, and total precipitating water (Khairoutdinov and Randall 2003). We use a horizontal domain size of $96 \text{ km} \times 96 \text{ km}$ with 3 km horizontal resolution and 64 vertical levels. This domain has approximately the same size as a single column of a typical global climate model (GCM), so our integrations can be viewed as representative of the state that would be achieved by a single column of a GCM in the absence of any grid-scale horizontal flow between GCM grid cells. Rotation is not included in the equations of motion. Insolation is set to the annual- and diurnal-mean value at 30°N on Earth ($\sim 364 \text{ W/m}^2$). We use the radiative transfer scheme from the NCAR Community Climate Model (CCM3; Kiehl et al. 1998).

The lower boundary of the model is an idealized representation of a land surface having a heat capacity equivalent to that of one meter of water. Surface temperature is interactively computed based on a surface energy balance. Surface fluxes are based on bulk aerodynamic formulae with constant surface exchange coefficients and a constant surface wind speed of 7 m/s. To represent the reduced evaporation that typically occurs over land compared to ocean, we multiply the latent heat flux by a fraction $\beta = 0.5$, similar to what was done in ME99. Substantial free-tropospheric cooling occurs when β is reduced below 0.25, as expected for a very dry climate, but there is little sensitivity to β for $\beta \geq 0.25$.

In our CSRM simulations, we vary surface pressure, or equivalently surface elevation, at a prescribed surface albedo, and compare the tropospheric temperatures after the system reaches an RCE state. That is, we conduct one simulation for each combination of surface pressure and surface albedo, presenting horizontal and time averages over the last 300 days when the system is in the equilibrium state. We refer to integrations by both surface height and surface pressure, with approximate heights of 0, 1, 2, 3, 4, and 6 km having surface pressures of, respectively, 1008, 890, 790, 710, 600, and 500 hPa. Further description of the experimental set-up is provided in the relevant sections below.

3. Elevated heating in a CSRM

In the first set of experiments, we set surface albedo to 0.24 (near the observed mean for the Tibetan Plateau), and vary surface pressure from 1008 to 500 hPa by roughly 100 hPa increments, as mentioned above. We find that elevated heating, previously found in the RCE states of a single-column model (ME99), indeed exists in these CSRM simulations that explicitly represent three-dimensional moist convective processes (Fig. 1a). Specifically, surface temperature decreases with surface height, but at a rate of 2.2°C/km that is much slower than typical moist adiabatic lapse rates ($6\text{-}10^\circ\text{C/km}$), consistent with the estimate of ME99. A given pressure-level of the troposphere is thus warmer over an elevated surface than over a non-elevated surface in RCE, by as much as 20°C for a 5 km change in surface height.

In a convecting atmosphere, upper-level temperature covaries with subcloud equivalent potential temperature, θ_{eb} , better than it does with surface temperature; this is a central tenet of the theory of convective quasi-equilibrium (CQE). This can be expressed as

$$\delta\theta_{eb} \propto \delta\theta_e^*, \quad (1)$$

where δ represents a variation in time or space, and θ_e^* is the saturation value of the equivalent potential temperature of the convecting layer (θ_e^* depends only on temperature and thus provides an appropriately rescaled measure of temperature; also, for our purposes, equivalent potential temperature could be replaced by moist static energy or moist entropy). When variations in θ_{eb} and θ_e^* are equal rather than simply proportional, the relationship is referred to as strict quasi-equilibrium (SQE). The CQE framework and the SQE assumption have been used extensively in studies of large-scale tropical and monsoon dynamics (Emanuel et al. 1994; Brown and Bretherton 1997; Neelin and Zeng 2000; Nie et al., 2010). Here we find that SQE holds when surface elevation is varied in simulations of RCE. In particular, Fig. 1b shows θ_{eb} plotted against θ_e^* averaged over an upper-tropospheric layer (200-400 hPa), for the same simulations plotted in Fig. 1a and for two additional sets of simulations where surface elevation is fixed at 0 and then at 4 km while surface albedo is varied from 0.22 to 0.40. The points fall very close to the one-to-one line, consistent with the SQE hypothesis. The elevated heating effect can thus be characterized by the effect of surface elevation on θ_{eb} or on upper-tropospheric temperature – the two serve as nearly equivalent metrics. This provides a more convenient and insightful way to use surface air properties to think about the elevated heating effect: one does not need to compare the rate of surface temperature decrease with the moist adiabatic lapse rate, but only assess whether θ_{eb} increases as surface elevation rises. More generally, one might ask what controls the quantitative sensitivity of θ_{eb} to surface elevation. Is this sensitivity always positive, or might θ_{eb} decrease and the upper-troposphere cool as the surface is lifted?

4. Theory

4.1. Derivation of an elevated heating index (EHI)

To answer these questions, we first define an elevated heating index (EHI), which is simply the sensitivity of θ_{eb} to surface elevation (z_s),

$$EHI \equiv \frac{d\theta_{eb}}{dz_s}. \quad (2)$$

In our model and in an atmosphere that can be described by a CQE framework, this index also represents the sensitivity of upper-tropospheric temperature to z_s , as discussed above. Now we express the EHI in terms of other variables by first approximating the subcloud θ_{eb} as the surface air equivalent potential temperature, θ_{es} ,

$$\theta_{eb} \approx \theta_{es} \approx T_s \left(\frac{p_0}{p_s} \right)^{R_d/c_{pd}} \exp \left(\frac{L_v q_s}{c_{pd} T_s} \right). \quad (3)$$

Here q_s is surface air specific humidity, R_d is the gas constant of dry air, c_{pd} is the specific heat at constant pressure of dry air, L_v is the latent heat of evaporation, p_0 is a reference pressure (here taken to be 1000 hPa), and T_s is surface air temperature. Taking the logarithm of each side of Eq. (3) and then differentiating with respect to surface elevation, we obtain

$$\frac{1}{\theta_{eb}} \frac{d\theta_{eb}}{dz_s} \approx \frac{1}{T_s} \frac{dT_s}{dz_s} - \frac{R_d}{c_{pd} p_s} \frac{dp_s}{dz_s} + \left(\frac{L_v}{c_{pd} T_s} \frac{dq_s}{dz_s} - \frac{L_v q_s}{c_{pd} T_s^2} \frac{dT_s}{dz_s} \right). \quad (4)$$

Specific humidity is related to vapor pressure by

$$q_s = \frac{\varepsilon H_s e_s^*}{p_s}, \quad (5)$$

where ε is the ratio of the gas constants for dry air and water vapor, H_s is surface relative humidity, and e_s^* is the saturation vapor pressure of surface air, which depends only on T_s . Assuming that H_s does not change much with surface elevation (Sherwood et al. 2010), which is consistent with our simulations (not shown), changes in specific humidity can thus be due to surface pressure changes or saturation vapor pressure changes. Combining Eqs. (4) and (5), we find

$$\frac{1}{\theta_{eb}} \frac{d\theta_{eb}}{dz_s} \approx \left[1 + \frac{L_v q_s}{c_{pd}} \left(\frac{1}{e_s^*} \frac{de_s^*}{dT_s} - \frac{1}{T_s} \right) \right] \frac{1}{T_s} \frac{dT_s}{dz_s} - \left(1 + \frac{L_v q_s}{R_d T_s} \right) \frac{R_d}{c_{pd} p_s} \frac{dp_s}{dz_s}, \quad (6)$$

where part of the last term can be expressed as

$$\frac{1}{p_s} \frac{dp_s}{dz_s} = -\frac{g}{R_d T_s}, \quad (7)$$

using hydrostatic balance and the ideal gas law. Substituting Eq. (7) into Eq. (6), we then find

$$\frac{1}{\theta_{eb}} \frac{d\theta_{eb}}{dz_s} \approx \left(1 + \frac{L_v q_s}{c_{pd}} \frac{1}{e_s^*} \frac{de_s^*}{dT_s} \right) \frac{1}{T_s} \frac{dT_s}{dz_s} + \left(1 + \frac{L_v q_s}{R_d T_s} \right) \frac{g}{c_{pd} T_s}, \quad (8)$$

where we dropped the $1/T_s$ term in square brackets in Eq. (6) because it is more than an order of magnitude smaller than the term involving de_s^*/dT_s . Based on Eq. (8), we obtain the elevated heating index (EHI; defined as $d\theta_{eb}/dz_s$),

$$EHI \approx \frac{\theta_{eb}}{T_s} \left[\left(1 + \frac{L_v q_s}{c_{pd}} \frac{1}{e_s^*} \frac{de_s^*}{dT_s} \right) \frac{dT_s}{dz_s} + \left(1 + \frac{L_v q_s}{R_d T_s} \right) \frac{g}{c_{pd}} \right], \quad (9)$$

which can be simplified as

$$EHI \approx A \cdot \left(\frac{dT_s}{dz_s} + \Gamma_{ms} \right), \quad (10)$$

with A a dimensionless number,

$$A = \frac{\theta_{eb}}{T_s} \left(1 + \frac{L_v q_s}{c_{pd}} \frac{1}{e_s^*} \frac{de_s^*}{dT_s} \right). \quad (11)$$

Although we originally defined T_s as the temperature of surface air, dT_s/dz_s is exactly equal to the derivative of surface temperature if the air-ground temperature difference does not change with z_s .

The coefficient A is positive definite and state dependent, being larger over high-elevation surfaces (i.e. larger θ_{eb}/T_s) and larger in warmer and moister climates; for example, A equals 2.1 for $T_s=15^\circ\text{C}$ and $p_s=1000$ hPa, and increases to 4.0 either when T_s increases to 32°C or p_s decreases to 500 hPa. The variable Γ_{ms} is

$$\Gamma_{ms} = \left[\left(1 + \frac{L_v q_s}{R_d T_s} \right) \right] / \left[\left(1 + \frac{L_v q_s}{c_{pd}} \frac{1}{e_s^*} \frac{de_s^*}{dT_s} \right) \right] \frac{g}{c_{pd}}, \quad (12)$$

which has a similar expression to, and here will thus be interpreted as, the moist adiabatic lapse rate evaluated at surface conditions. The only difference with the standard moist lapse rate is the use of specific humidity q_s instead of saturation specific humidity q_s^* , but this has a relatively small effect on Γ_{ms} . More precisely, Γ_{ms} is better interpreted as the moist adiabatic lapse rate at the lifted condensation level (LCL) where the air parcel becomes saturated; in this case the temperature T_s should be replaced with the temperature at the LCL and the EHI represents the sensitivity of θ_e at the LCL to surface elevation.

Eq. (10) is the central result of this subsection, and it intuitively but quantitatively shows the dependence of EHI on the relative magnitudes of the moist adiabatic lapse rate and dT_s/dz_s : upper-tropospheric temperature will be higher over an elevated surface than over a non-elevated surface only if the surface temperature T_s decreases, as z_s rises, more slowly than temperature along a moist adiabat (e.g. Fig. 1a). As is discussed above, the term A in (10) is a proportionality constant that gives the quantitative rate at which θ_{eb} changes with surface height, and it modulates the magnitude of the elevated heating effect simply because here EHI is defined with θ_{eb} ; if we instead define EHI using upper tropospheric temperature, the term A will disappear. The terms that control the sign and magnitude of the elevated heating effect are thus the moist adiabatic lapse rate and dT_s/dz_s . The former is easy to estimate given a background climate state, but what controls the sign and magnitude of dT_s/dz_s ?

4.2. Surface temperature dependence on surface height

We now present a theory for the value of dT_s/dz_s and explain how the effect of surface elevation on top-of-atmosphere (TOA) radiative fluxes sets the magnitude, and even the sign, of the elevated heating effect. We use a traditional climate sensitivity framework (Cess et al. 1990; Held and Soden 2000 and references therein),

$$\frac{dT_s}{dz_s} = - \left(\frac{dR}{dz_s} \right) / \lambda, \quad (13)$$

where R is the net radiative flux at TOA and $\lambda = dR/dT_s$ is the net climate feedback parameter of the radiative-convective system, representing the sensitivity of TOA radiation to surface temperature changes. In a traditional climate sensitivity calculation (e.g. Cess et al. 1990), such as for a doubling of CO_2 , the radiative forcing is imposed and so R can be viewed as the external forcing. Here there is another step in the chain of causation, since R is set by the radiative changes that occur as a consequence of the surface lifting. Although the changes in R and T_s that occur in response to a surface lifting occur simultaneously, we conceptually partition them into two parts: first the surface is lifted without any change in surface temperature and both radiation and moist convection adjust to the new surface elevation, then surface temperature adjusts to the altered net TOA radiation. These two conceptual parts of the response are represented by the numerator and denominator of Eq. (13), respectively, and we now discuss them in sequence.

We decompose the numerator of the right-hand side of Eq. (13) into shortwave and longwave components, and further separate the latter into changes associated with CO_2 , H_2O , and lapse-rate effects,

$$\begin{aligned}
\frac{dR}{dz_s} &\equiv \frac{dISR}{dz_s} - \frac{dOLR}{dz_s} \\
&\approx \frac{dISR}{dz_s} - \left(\frac{\partial OLR}{\partial \tau_{CO_2}} \right)_{T, \tau_{H_2O}} \left(\frac{d\tau_{CO_2}}{dz_s} \right) - \left(\frac{\partial OLR}{\partial \tau_{H_2O}} \right)_{T, \tau_{CO_2}} \left(\frac{\partial \tau_{H_2O}}{\partial z_s} \right)_T - \left(\frac{\partial OLR}{\partial T} \right)_{\tau_{CO_2}} \left(\frac{dT}{dz_s} \right) \quad (14) \\
&= SW - LW_{CO_2} - LW_{H_2O} - LW_{LAPSE}.
\end{aligned}$$

In the second line of Eq. (14), τ represents the optical depth of a given constituent. The four terms on the right-hand side of Eq. (14) represent the effect of surface elevation on, respectively, (i) incoming shortwave radiation (ISR) due to changes in clear-sky scattering and absorption by air (labeled in the third line above as SW), (ii) outgoing longwave radiation (OLR) due to changes in the optical depth of carbon dioxide (LW_{CO_2}), (iii) OLR due to changes in the amount of longwave absorption accomplished by a given mass of water, e.g. the pressure broadening of water vapor's absorption lines (LW_{H_2O}), and (iv) OLR due to changes in the atmospheric lapse rate (LW_{LAPSE}). The last of these includes the effects of changes in tropospheric temperature as well as the changes in water vapor that accompany the lapse rate changes, assuming fixed relative humidity. Cloud radiative effects are highly uncertain and not included in Eq. (14), but will be discussed later. To be clear, these four terms represent the effect of changes in surface height on TOA radiation at fixed surface temperature; the resulting change in surface temperature is then found using Eq. (13) as in a traditional climate sensitivity calculation.

The denominator of Eq. (13) can be determined in multiple ways. First, we double CO_2 concentrations in one of our CSRM simulations that has a surface albedo of 0.24 and a surface pressure of 500 hPa, and find that $\lambda \approx 2.5 \text{ W}/(\text{m}^2 \cdot \text{K})$. In the next section we present an alternate estimate of λ calculated for a change in the TOA shortwave induced by varying surface albedo. Although the value of λ might be expected to depend on the details of how R is perturbed, we find good quantitative agreement between values of λ obtained through different radiative forcings. We emphasize that here λ is an equilibrium climate feedback parameter; the global mean transient climate sensitivity has been shown to vary in time as regions with different surface heat capacities warm at different rates, but even that complication can be explained by assuming climate feedbacks that are linear and constant in time for a particular region (Armour et al. 2013). The value of λ for our simple RCE system additionally seems to have little sensitivity to the nature of R , although we have not sampled very dry states or a large range of basic state temperatures.

5. Demonstrations with off-line radiative calculations

Now we use off-line radiative calculations to estimate the value of each term in Eq. (14), thereby understanding what sets the magnitude of the elevated heating effect. We use a single-column radiative-convective toolkit called CliMT (Climate Modeling and diagnostics Toolkit; Caballero 2012) to quantitatively attribute the net TOA radiative flux change to the processes represented by each term. Like other single-column models, CliMT has no horizontal degrees of freedom; it is vertically discretized into 26 levels. We take the approach of imposing vertical profiles of temperature, humidity, and CO_2 in the model, then using CliMT to conduct off-line radiative transfer calculations to obtain TOA radiative fluxes. To ease comparison with our CSRM results,

we chose the same NCAR CCM3 radiative scheme for CliMT as used in our CSRM (although some details of that scheme’s implementation may differ).

We estimate the various partial derivatives in Eq. (14) based on TOA radiative fluxes from a large ensemble of CliMT simulations with different surface temperatures and surface elevations. We start with a set of simulations having non-elevated surfaces (i.e. surface pressure = $p_{ref} = 1000$ hPa) and vertical temperature profiles that are moist pseudoadiabatic from the specified surface temperature up to the tropopause, above which a constant stratospheric temperature of -100°C is prescribed. We used a moist pseudoadiabat starting from the surface instead of from the LCL for the sake of simplicity and consistency with our derivation of the EHI, but expect similar results if a dry adiabat was used below the LCL, as long as the LCL is not very high. The relative humidity (RH) profile mimics that in the CSRM simulations and is fixed for all climate sensitivity calculations here: it has a surface value of 69%, local maxima of 87% and 90% at altitudes of 1 km and 15 km above the surface, respectively, a local minimum of 60% at 8 km, is zero above 21 km, and is piecewise linear in between. There are no cloud-radiative interactions. Surface albedo is tuned for each surface temperature to obtain an RCE state (i.e. one in which the net TOA radiative flux is zero). We refer to this as the reference state, with one reference state existing for each value of surface temperature examined. We then lift the surface until its pressure drops to p_{target} , then calculate how the TOA radiative flux, R , changes as a result of that lifting while T_s is held constant (this gives dR/dz_s in Eq. [13], and it is the climate feedback parameter λ that determines how T_s responds to this change in R after equilibration). This is done for many combinations of T_s and p_{target} (listed in Section 6); for each combination we perform several versions of this calculation to find the various total and partial derivatives in Eq. (14):

(i) Estimating the first term in Eq. (14), $d(ISR)/dz_s$, is straightforward: it is simply the net ISR change at TOA, relative to the reference state, when the surface is elevated while T_s and the CO_2 mixing ratio are fixed but other quantities are allowed to change. In particular, the lapse rate is allowed to change (Fig. 2a shows the pressure-dependence of the moist lapse rate) and vapor pressure is modified to maintain fixed $RH(z)$. Changes in clear-sky shortwave absorption by water vapor are included in this estimate but are typically about a factor of 4 smaller than the changes due to scattering by dry air.

(ii) The second term, which represents the influence of changes in CO_2 optical depth on OLR, is estimated by comparing OLR in two configurations that differ only in their CO_2 concentrations. Both have the same T_s and $T(z)$ [with z indicating the height above each surface] as the reference state but both use a lifted surface (with surface pressure p_{target}). One of these configurations uses the same CO_2 mixing ratio as the reference state, while the mixing ratio in the other is increased by a factor of $(p_{ref}/p_{target})^2$ to obtain the same CO_2 optical path as the reference state. Note that the column-integrated mass of CO_2 changes by p_{ref}/p_{target} , but the optical path of CO_2 changes quadratically because of the pressure-broadening of absorption lines (Pierrehumbert 2010). Since water vapor absorption features are also broadened by pressure, we multiplied the vapor pressure in both configurations by a scale factor in order to hold the water vapor optical path constant. We determined this scale factor by varying it iteratively until the OLR difference with the reference state was less than 0.001 W/m²; the factor turns out to be slightly greater than unity.

(iii) The third term represents the influence of modified water vapor absorption while CO_2 optical path, T_s , $T(z)$, and $RH(z)$ are all held constant. This is estimated by comparing OLR for two cases, both with lifted surfaces and with the CO_2 scaled by $(p_{ref}/p_{target})^2$ to obtain the same CO_2 optical path as the reference state. In one case the vapor pressure is set to be the same as in the reference case, while in the other it is scaled by the same factor described in (ii) above to compensate for changes in pressure-broadening.

(iv) The fourth term, which represents the effect of changes in lapse rate, is estimated by comparing OLR in two cases, both with lifted surfaces with the same T_s and with CO_2 scaled by $(p_{ref}/p_{target})^2$ to obtain the same CO_2 optical path as the reference state. One case uses the same $T(z)$ and vapor pressure as the reference state, while in the other $T(z)$ is modified so that lapse rates are moist adiabatic at the modified surface pressure; vapor pressure is also changed to maintain constant $RH(z)$.

To directly compare with the conditions used in our main set of CSRM experiments, we now show in Fig. 3 our off-line estimates of each term in Eq. (14) for a change in p_s from 1000 hPa to 500 hPa and a surface albedo of 0.24, which has a corresponding RCE surface temperature of about 28°C for the non-elevated surface. Using the off-line calculations described in items (i) and (ii) above, we find that the increase in absorbed shortwave radiation due to reduced clear-sky scattering is nearly equal to the increase in outgoing longwave radiation due to the reduction in CO_2 optical path (i.e. $\text{SW} \approx \text{LW}_{\text{CO}_2}$ in Eq. [14]; note that CO_2 optical depth increases quadratically with p_s , rather than linearly, because of pressure-broadening of its absorption lines). The water vapor term ($\text{LW}_{\text{H}_2\text{O}}$, the calculation of which is described in item [iii] above) has a cooling effect of similar magnitude to each of the previous two terms, and arises from the fact that a given vertical profile of water vapor pressure has a smaller optical depth when surface pressure is reduced. The largest term, LW_{LAPSE} , has not been considered by previous studies of elevated heating and so will now be discussed in detail.

The LW_{LAPSE} term in Eq. (14) represents enhanced longwave cooling that results from the reduction in moist adiabatic lapse rate that occurs as p_s drops. It can be understood by considering what happens to R as the surface is elevated with surface temperature and relative humidity fixed. Since vapor pressure depends only on temperature and since p_s decreases, the water vapor mixing ratio increases and thus the moist adiabatic lapse rate is reduced (Fig. 2a). That is, as p_s decreases there is less dry air to be heated by condensation of a given mass of water. As a result, air temperature above the elevated surface will be warmer than air temperature the same distance above the non-elevated surface (e.g. Fig. 2b, where height is referenced in km relative to each surface). The vertically integrated effect of this dependence of the moist adiabatic lapse rate on surface pressure produces a much warmer troposphere above the elevated surface (again, at a given altitude relative to each surface). For surface temperatures of $5\text{-}15^\circ\text{C}$, a reduction of surface pressure from 1000 to 500 hPa produces a temperature contrast as large as $10\text{-}30^\circ\text{C}$ at altitudes of 10-15 km (Fig. 2c).

This increase of air temperature at a given altitude above the surface increases OLR, cooling the column. Some of this cooling is offset by the reduction in OLR that occurs as water vapor pressure increases in the warmer upper troposphere, where relative humidity has been argued to stay roughly constant as the lapse rate changes (Sherwood et al. 2010). This lapse rate effect,

including the associated water vapor changes, is analogous to that which occurs as CO_2 is increased over a non-elevated surface (Soden and Held 2006). As in that well-known scenario, our calculations show that the net effect of the change in lapse rate is an increase in OLR with surface elevation (i.e. the temperature effect dominates). For the particular case of lifting the surface from 1000 to 500 hPa, the temperature part of the lapse rate effect accounts for -48.1 W/m^2 while the associated water vapor effect is $+36.7 \text{ W/m}^2$, yielding a total lapse rate effect of -11.4 W/m^2 . This net lapse rate effect is larger than any of the other terms discussed above or mentioned in ME99 and, unlike the other terms, it involves both moist convective and radiative processes rather than radiation alone. One could categorize the terms in Eq. (14) differently and add together all the contributions from greenhouse gas changes (LW_{CO_2} , $\text{LW}_{\text{H}_2\text{O}}$, and the water vapor component of LW_{LAPSE}), yielding a positive value that would cancel much but not all of the temperature component of LW_{LAPSE} .

The sum of all four of these individual effects (i through iv above) is -20.9 W/m^2 for our surface lifting from 1000 to 500 hPa. Nonlinearities are slight: simulation in CliMT of the net change in R produced by the same surface lifting while T_s is fixed has a magnitude that is only about 1 W/m^2 smaller (compare black and gray lines in Fig. 3).

To predict the change in surface temperature caused by the surface lifting, we combine this net change in R with the climate feedback parameter λ . As mentioned in the previous section, a doubled- CO_2 CSRM simulation yielded an estimate of $\lambda = 2.5 \text{ W}/(\text{m}^2 \cdot \text{K})$. Although one might think that the climate feedback on a CO_2 doubling could differ substantially from the feedback on a surface lifting, it is important to remember that a change in surface elevation alters surface temperature partly through a change in CO_2 optical depth. To test if the climate feedback parameter is sensitive to the details of how the TOA radiative fluxes are perturbed, we separately estimated λ in CliMT using two different approaches with the same background state as in the CSRM (surface albedo = 0.24, surface pressure = 500 hPa). In the first approach we doubled CO_2 concentrations just like we did in the CSRM, while in the second approach we examined the surface temperature change caused by a change in surface albedo. The latter provides an estimate of λ based on a shortwave rather than a longwave forcing. We obtained a nearly identical value of λ for these two independent approaches using CliMT: 2.0 W/m^2 . The modest differences in λ obtained from the CSRM and from CliMT are expected, given their totally different treatments of convection and the suppressed cloud-radiation interaction in CliMT.

Given this value of λ (i.e. $2.5 \text{ W}/[\text{m}^2 \cdot \text{K}]$), the net change in R of -20.9 W/m^2 in response to a surface lifting from 1000 to 500 hPa is expected to cool the surface by about 8.4 K. This explains about 65% of the 13 K surface temperature drop found in our CSRM (Fig. 1a, where the black dashed line represents our linearized estimate). The remaining one-third of the temperature drop simulated in the CSRM might come from the full complexity of those simulations, e.g. cloud radiative effects, relative humidity changes, deviation of temperatures from moist adiabatic, and radiative transfer details that differ from those of our off-line calculations. Note that if we use the CliMT estimate of λ , this climate sensitivity framework can exactly predict the surface temperature dependence on surface elevation in the idealized off-line calculations with CliMT (i.e. no cloud radiative effect, fixed relative humidity profile, and temperatures following moist adiabats).

6. Elevated heating across a wide range of climates

We now use our offline radiative-convective calculations together with additional CSRM simulations to characterize the elevated heating effect over a broad range of surface temperatures and surface elevation changes. We begin by repeating the complete set of offline radiative-convective calculations described in the previous section for every T_s between 0 and 40°C, with an interval of 1°C, and for every target surface pressure between 500 and 990 hPa, with an interval of 10 hPa. This allows the magnitude of each term in Eq. (14) to be plotted as a function of T_s and target surface pressure (Fig. 4a-d). For every T_s and p_{target} combination, we also calculate the net change in R that includes all effects and their interactions (i.e. the surface is lifted and the lapse rate is allowed to change while T_s , CO_2 mixing ratio, and $RH(z)$ are fixed). The sum of all individual terms (i – iv above) is very nearly equal to this net change in R (compare panels e and f of Fig. 4; also see Fig. 3), which shows that the various effects are additive with small nonlinearities; this also lends confidence to our methodology.

The results show that the net change in shortwave has an opposite sign but similar magnitude to the change in OLR that is caused by the modified CO_2 optical path. The shortwave term increases with surface temperature because of the larger ISR associated with the smaller surface albedo required to achieve a warmer RCE reference state (Fig. 4a). It is thus important to remember that the surface temperature dependence of each term shown in Fig. 4 may differ if surface temperature is altered by a forcing other than surface albedo; e.g. the shortwave term may not increase in magnitude if the climate state is warmed by an increase in greenhouse gas concentrations.

The change in OLR due only to pressure-induced changes in water vapor absorption has the same sign and roughly the same magnitude as the CO_2 -driven OLR changes. These two terms have little sensitivity to T_s and scale near-linearly with the change in surface pressure (Fig. 4b,c).

The fourth term, which includes all effects of changes in lapse rate, is relatively weak in cold climates but is much stronger than any of the other terms for large changes in surface pressures in warm climates (Fig. 4d). Its sensitivity to T_s is much larger than, and of opposite sign to, that of the shortwave change (Fig. 4a,d). We emphasize again that the lapse rate term includes the direct effect of temperature (e.g. OLR increases as the surface is lifted because the upper-troposphere warms as the moist lapse rate decreases) as well as the compensating effect of moisture (e.g. a reduced lapse rate and fixed RH require an increase in upper-tropospheric vapor pressure and thus a reduction in OLR). Unlike the shortwave term, the temperature-dependence of the lapse rate term is expected to hold regardless of what causes the change in climate state, as long as the vertical temperature profile is strongly constrained by moist convection.

The combined effects of these four terms, dR/dz_s , is negative and increases in magnitude as the climate warms (Fig. 4e,f), causing an increase in the magnitude of dT_s/dz_s given a constant climate feedback λ . However, the actual dT_s/dz_s change will be affected by the state-dependence of λ , the value of which involves large uncertainty due to cloud-radiative feedbacks and which might furthermore be model dependent. This is an important topic in and of itself, but we leave it for future investigation. At the same time, the moist lapse rate is smaller in warmer climates (Fig.

2a), which directly controls the last term on the right-hand side of Eq. (10). Both factors (i.e. a more negative dT_s/dz_s and a less positive Γ_{ms}) work together to decrease the EHI. The exact value of EHI in our definition is also dependent on the coefficient A in Eq. (11), but as we will show later, the state-dependence of the EHI is dominated by changes in dR/dz_s .

Since it is clear that dR/dz_s is dominated by the lapse rate term in the warmest climates (Fig. 4d), we now discuss the physics and implications of this effect in greater detail. The OLR change can be linearized as $4 \sigma T^3 \Delta T$, with σ the Stefan-Boltzmann constant, T the effective emission temperature, and ΔT the change in temperature at the emission level due to the lapse rate change. As climate warms, the lapse rate change (ΔT) that occurs as the surface is lifted becomes larger in the upper troposphere (e.g. Fig. 2c). This increases the OLR change, making dT_s/dz_s more negative in warmer climates and reducing the magnitude of elevated heating.

To confirm the hypothesis that elevated heating weakens in warmer climates, we conduct additional tests using the cloud-system resolving model. Due to computational resource limits, we are not able to explore the wide range of surface temperatures and elevations examined in the off-line radiative calculations. Instead, we prescribe surface albedos that range from 0.22 to 0.40 to control the background climate state, and we elevate the surface from 0 km (i.e. 1000 hPa surface pressure) to 4 km (600 hPa surface pressure). We find that the decrease in T_s that occurs as the surface is lifted is larger at lower albedos (Fig. 5a). Also, we find that a given surface lifting produces a smaller increase in θ_{eb} in warmer climates, which confirms that elevated heating is indeed reduced in warmer climates in our CSRMs (Fig. 5b). As for the first set of simulations described in Section 3, SQE holds for these simulations and thus θ_{eb} is a good indicator of upper tropospheric temperature (Fig. 1b). These additional CSRMs thus confirm our hypothesis, motivated by our theory and off-line calculations, that elevated heating is weaker in warmer climates.

Can the elevated heating effect approach zero or even change sign to become elevated cooling? As shown by Eq. (10), the sign of the EHI is determined by the relative magnitudes of the decrease in surface temperature, dT_s/dz_s , and the moist adiabatic lapse rate Γ_{ms} . As illustrated in Fig. 5a, the former term increases in magnitude and the latter term decreases in warmer climates. Therefore, if the magnitude of dT_s/dz_s exceeds that of Γ_{ms} , elevated cooling will occur.

To achieve a warmer climate state, we set the surface albedo to 0.2 in the CSRMs, which is lower than any of the values shown in Fig. 5. At this surface albedo, the system with a non-elevated surface reaches a runaway greenhouse regime because the net downward solar radiation exceeds the upper limit of OLR the system can emit (Ingersol 1969). A runaway greenhouse state is also achieved for surface elevations of 1 km (i.e. 890 hPa surface pressure) at this albedo. However, when we elevate the surface to 1.5 km (i.e. 850 hPa surface pressure), the system achieves an RCE state. Further increasing the surface elevation to 2 km (i.e. 790 hPa surface pressure) produces an RCE state with a cooler troposphere than that in the case with surface elevation of 1.5 km (Fig. 6a,b). Although the temperature at a given pressure level only drops by less than 1 K when we lift the surface from 1.5 to 2 km, this clearly shows that one cannot universally expect tropospheric warming as surface elevation increases. In very warm and moist climates, the lapse rate effect in Eq. (14) dominates to produce elevated cooling. This elevated cooling

effect (i.e. negative EHI) persists when the CSRM simulations are repeated without radiatively active clouds, showing that it is not dependent on uncertain cloud-radiative effects (Fig. 6c,d).

7. Discussion

Elevated heating produced by the rise of orography has been widely invoked as a forcing for thermally direct circulations that influence regional climate in Earth's tropics and subtropics (e.g. Staff Members of Academia Sinica 1958; Wu et al. 2007). It has also been argued to set some characteristics of the general circulation of the Martian atmosphere (Richardson and Wilson 2002). Yet despite its importance, the physics of elevated heating has not been well understood. For example, no theory existed for the quantitative sensitivities of surface temperature and upper-tropospheric temperature to surface elevation, even in the idealized state of radiative-convective equilibrium.

Here we built on the work of ME99, who used parameterized convection to simulate radiative-convective equilibrium states over surfaces of different elevations. It is intuitive that elevated heating will occur if surface temperature decreases with surface elevation more slowly than the moist adiabatic lapse rate; we used a formalization of this idea together with a climate sensitivity framework to understand the radiative-convective processes that set the magnitude and sign of the elevated heating effect. In particular, we showed that the magnitude and sign of the elevated heating effect are set by top-of-atmosphere radiative changes dominated by the sensitivity of the moist adiabatic lapse rate to surface height. Based on the quantitative characteristics of this lapse rate feedback, we hypothesized that surface temperature should decrease more rapidly with surface elevation in warmer climates, and thus elevated heating should be weaker in warmer climates. This hypothesis is supported by off-line radiative calculations and is further confirmed by cloud-system resolving simulations. In a sufficiently warm climate, elevated heating can even change sign to become elevated cooling.

Much work remains to be done to connect our results with real climate states where large-scale dynamics interact with radiative-convective processes, and where surface albedo and soil moisture may be influenced by surface elevation. For instance, observations show a faster warming trend during the last 50 years over the Tibetan Plateau compared with its lower-elevation surroundings (e.g. Mountain Research Initiative EDW Working Group 2015). This observed trend might be taken to suggest that elevated heating increases in magnitude in warmer climates, which would contradict our RCE findings. However, the enhanced warming trend at high elevations might be caused primarily by changes in the large-scale circulation, land surface drying, surface albedo reductions due to the melting of snow and ice cover, aerosol changes, and more. We will quantitatively explore in separate work how changes in surface albedo can potentially compensate for the elevated heating effect (Hu and Boos 2016, in preparation for *Geophys. Res. Lett.*), and hope that other idealized modeling and observational analyses will help in understanding the more general interactions between large-scale dynamics, land surface hydrology, and the radiative-convective physics of elevated heating.

Acknowledgements

We acknowledge support from National Science Foundation grants AGS-1253222 and AGS-1515960, Office of Naval Research grant N00014-15-1-2531, and the Yale University Faculty of Arts and Sciences High Performance Computing facility. We thank Peter Molnar, Mark Brandon, and Yiqi Zheng for helpful discussions.

References

- Armour, K. C., C. M. Bitz, and G. H. Roe, 2013: Time-Varying Climate Sensitivity from Regional Feedbacks. *J. Climate*, **26**, 4518–4534, doi: 10.1175/JCLI-D-12-00544.1.
- Boos, W. R., and Z. M. Kuang, 2010: Dominant control of the South Asian monsoon by orographic insulation versus plateau heating. *Nature*, **463**, 218–222, doi:10.1038/nature08707.
- Brown, R. G., and C. S. Bretherton, 1997: A test of the strict quasi-equilibrium theory on long time and space scales. *J. Atmos. Sci.*, **54**, 624–638, doi:10.1175/1520-0469(1997)054<0624:ATOTSQ>2.0.CO;2
- Caballero, R., 2012: CliMT: Climate Modeling and Diagnostics Toolkit. [Available online at <https://github.com/rodrigo-caballero/CliMT>.]
- Chakraborty, A., R. S. Nanjundiah, and J. Srinivasan, 2006: Theoretical aspects of the onset of Indian summer monsoon from perturbed orography simulations in a GCM. *Ann. Geophys.*, **24**, 2075–2089, doi:10.5194/angeo-24-2075-2006.
- Cess, R. D., 1975: Global climate change: an investigation of atmospheric feedback mechanisms. *Tellus*, **27**, 193–198, doi:10.1111/j.2153-3490.1975.tb01672.x.
- Cess, R. D., et al., 1990: Intercomparison and interpretation of climate feedback processes in 19 atmospheric general circulation models, *J. Geophys. Res.-Atmos.*, **95**, 16601–16615, doi:10.1029/JD095iD10p16601.
- Colman, R., 2003: A comparison of climate feedbacks in general circulation models. *Climate Dyn.*, **20**, 865–873, doi:10.1007/s00382-003-0310-z.
- Emanuel, K., A. A. Wing, and E. M. Vincent, 2014: Radiative-convective instability. *J. Adv. Model Earth Sy.*, **6**, 75–90, doi:10.1002/2013MS000270.
- Emanuel, K. A., J. D. Neelin, and C. S. Bretherton, 1994: On Large-Scale Circulations in Convecting Atmospheres. *Q. J. Roy. Meteor. Soc.*, **120**, 1111–1143, doi:10.1002/qj.49712051902.
- Flohn, H., 1953: Hochgebirge und allgemeine Zirkulation. II. Die Gebirge als Wärmequellen. *Arch. Meteorol. Geophys. Bioklimatol.*, **A 5**, 265–279.
- Flohn, H., 1968: *Contributions to a meteorology of the Tibetan Highlands*. Department of Atmospheric Science, Colorado State University.
- Hahn, D. G., and S. Manabe, 1975: Role of Mountains in South Asian Monsoon Circulation. *J. Atmos. Sci.*, **32**, 1515–1541, doi:10.1175/1520-0469(1975)032<1515:TROMIT>2.0.CO;2.
- Halevy, I., R. T. Pierrehumbert, and D. P. Schrag, 2009: Radiative transfer in CO₂-rich paleoatmospheres, *J. Geophys. Res.-Atmos.*, **114**, D18112, doi:10.1029/2009JD011915.
- Held, I. M., and B. J. Soden, 2000: Water vapor feedback and global warming. *Annu. Rev. Energy Environ.*, **25**, 441–475, doi:10.1146/annurev.energy.25.1.441.

- Held, I. M., M. Zhao, and B. Wyman, 2007: Dynamic radiative-convective equilibria using GCM column physics. *J. Atmos. Sci.*, **64**, 228-238, doi:10.1175/JAS3825.11.
- Hu, S., and W. R. Boos, 2016: Competing effects of surface albedo and orographic elevated heating on regional climate, *Geophys. Res. Lett.*, in preparation.
- Ingersol, A. P., 1969: Runaway Greenhouse - a History of Water on Venus. *J. Atmos. Sci.*, **26**, 1191-1198, doi:10.1175/1520-0469(1969)026<1191:TRGAHO>2.0.CO;2.
- Khairoutdinov, M. F., and D. A. Randall, 2003: Cloud resolving modeling of the ARM summer 1997 IOP: Model formulation, results, uncertainties, and sensitivities. *J. Atmos. Sci.*, **60**, 607-625, doi:10.1175/1520-0469(2003)060<0607:CRMOTA>2.0.CO;2.
- Kiehl, J. T., J. J. Hack, G. B. Bonan, B. A. Boville, D. L. Williamson, and P. J. Rasch, 1998: The National Center for Atmospheric Research Community Climate Model: CCM3. *J. Climate*, **11**, 1131-1149, doi:10.1175/1520-0442(1998)011<1131:TNCFAR>2.0.CO;2.
- Lindzen, R. S., and A. Y. Hou, 1988: Hadley Circulations for Zonally Averaged Heating Centered Off the Equator. *J. Atmos. Sci.*, **45**, 2416-2427, doi:10.1175/1520-0469(1988)045<2416:HCFZAH>2.0.CO;2.
- Ma, D., W. Boos, and Z. Kuang, 2014: Effects of Orography and Surface Heat Fluxes on the South Asian Summer Monsoon. *J. Climate*, **27**, 6647-6659, doi:10.1175/JCLI-D-14-00138.1.
- Molnar, P., and K. A. Emanuel, 1999: Temperature profiles in radiative-convective equilibrium above surfaces at different heights. *J. Geophys. Res.-Atmos.*, **104**, 24265-24271, doi:10.1029/1999JD900485.
- Molnar, P., P. England, and J. Martinod, 1993: Mantle Dynamics, Uplift of the Tibetan Plateau, and the Indian Monsoon. *Rev. Geophys.*, **31**, 357-396, doi:10.1029/93RG02030.
- Mountain Research Initiative EDW Working Group, 2015: Elevation-dependent warming in mountain regions of the world. *Nature Climate Change*, **5**, 424-430, doi:10.1038/nclimate2563.
- Neelin, J. D., and N. Zeng, 2015: A quasi-equilibrium tropical circulation model—Formulation. *J. Atmos. Sci.*, **57**, 1741-1766, doi:10.1175/1520-0469(2000)057<1741:AQETCM>2.0.CO;2.
- Nie, J., W. R. Boos, and Z. M. Kuang, 2010: Observational Evaluation of a Convective Quasi-Equilibrium View of Monsoons. *J. Climate*, **23**, 4416-4428, doi:10.1175/2010JCLI3505.1.
- Pierrehumbert, R. T., 2010: *Principles of planetary climate*. Cambridge University Press.
- Popke, D., B. Stevens, and A. Voigt, 2013: Climate and climate change in a radiative-convective equilibrium version of ECHAM6. *J. Adv. Model Earth Sy.*, **5**, 1-14, doi:10.1029/2012MS000191.
- Rao, G. V., and S. Erdogan, 1989: The Atmospheric Heat-Source over the Bolivian Plateau for a Mean January. *Bound.-Lay. Meteorol.*, **46**, 13-33, doi:10.1007/BF00118444.
- Richardson, M. I., and R. J. Wilson, 2002: A topographically forced asymmetry in the martian circulation and climate. *Nature*, **416**, 298-301, doi:10.1038/416298a.
- Romps, D. M., 2011: Response of Tropical Precipitation to Global Warming. *J. Atmos. Sci.*, **68**, 123-138, doi:10.1175/2010JAS3542.1.
- Satoh, M., 1994: Hadley Circulations in Radiative-Convective Equilibrium in an Axially-Symmetrical Atmosphere. *J. Atmos. Sci.*, **51**, 1947-1968, doi:10.1175/1520-0469(1994)051<1947:HCIREI>2.0.CO;2.

- Schneider, T., 2006: The general circulation of the atmosphere. *Annu. Rev. Earth Planet. Sci.*, **34**, 655-688, doi:10.1146/annurev.earth.34.031405.125144.
- Sherwood, S. C., W. Ingram, Y. Tsushima, M. Satoh, M. Roberts, P. L. Vidale, and P. A. O'Gorman, 2010: Relative humidity changes in a warmer climate. *J. Geophys. Res.-Atmos.*, **115**, D09104, doi:10.1029/2009JD012585.
- Silvers, L., B. Stevens, T. Mauritsen, and M. A. Giorgetta, 2016: Radiative convective equilibrium as a framework for studying the interaction between convection and its large-scale environment. *J. Adv. Model Earth Sy.*, **8**, doi:10.1002/2016MS000629.
- Soden, B. J., and I. M. Held, 2006: An assessment of climate feedbacks in coupled ocean-atmosphere models. *J. Climate*, **19**, 3354-3360, doi:10.1175/JCLI3799.1.
- Staff Members of Academia Sinica, 1958: On the General Circulation over Eastern Asia III. *Tellus*, **10**, 299-312, doi:10.1111/j.2153-3490.1958.tb02018.x.
- Tao, W. K., J. Simpson, C. H. Sui, C. L. Shie, B. Zhou, K. M. Lau, and M. Moncrieff, 1999: Equilibrium states simulated by cloud-resolving models. *J. Atmos. Sci.*, **56**, 3128-3139, doi:10.1175/1520-0469(1999)056<3128:ESSBCR>2.0.CO;2.
- Wu, G. X., and Coauthors, 2007: The influence of mechanical and thermal forcing by the Tibetan Plateau on Asian climate. *J. Hydrometeorol.*, **8**, 770-789, doi:10.1175/JHM609.1.
- Zaitchik, B. F., J. P. Evans, and R. B. Smith, 2007: Regional impact of an elevated heat source: The Zagros Plateau of Iran. *J. Climate*, **20**, 4133-4146, doi:10.1175/JCLI4248.1.

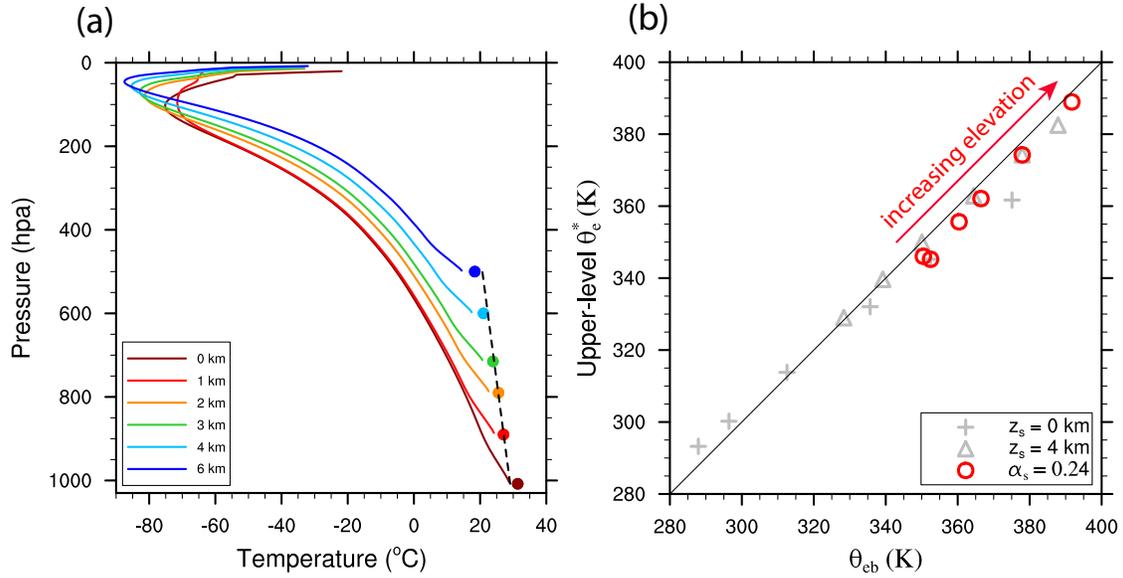


Figure 1: (a) Equilibrium temperature profiles from cloud system-resolving simulations with (a) various surface elevations but the same surface albedo of 0.24. Colored dots indicate the surface temperature for the corresponding case; the black dashed line marks the theoretical estimate of surface temperature obtained from our cloud-free single-column model (see text for details), with its slope based on surface elevations of 0 km and 6 km. (b) Upper tropospheric (200–400 hPa averaged) saturation equivalent potential temperatures versus subcloud (25 hPa above surface) equivalent potential temperatures for the simulations shown in panel a (red circles) and for the simulations with different surface albedos discussed in Section 6 (gray markers; crosses and triangles are for surface elevations of 0 km and 4 km, respectively). The black solid one-to-one line corresponds to the strict quasi-equilibrium assumption.

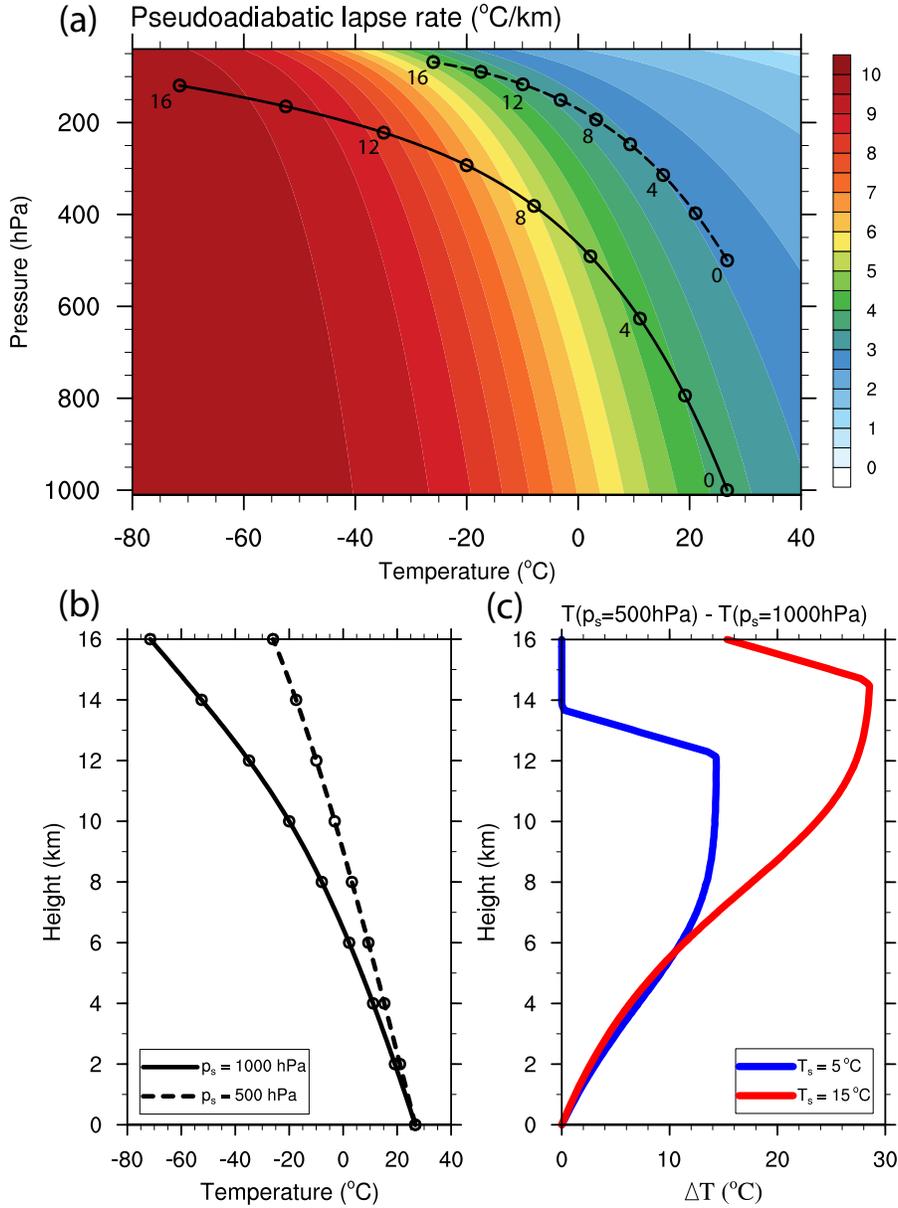


Figure 2: (a) Theoretical moist pseudoadiabats starting from surface elevations of 0 km (solid line) and 5 km (dashed line), together with the moist pseudoadiabatic lapse rate (colors). Numbers on the pseudoadiabats indicate the height in km above each surface, with circles shown every 2 km. (b) The same pseudoadiabats as in panel a but plotted in height coordinates, with the vertical axis indicating the height above each surface. (c) Vertical profile of the temperature difference between moist pseudoadiabats over surfaces with surface pressures of 1000 and 500 hPa, starting at surface temperatures of 5 $^{\circ}\text{C}$ (blue) and 15 $^{\circ}\text{C}$ (red). Note the greater temperature difference in the warmer climate, particularly in the upper troposphere.

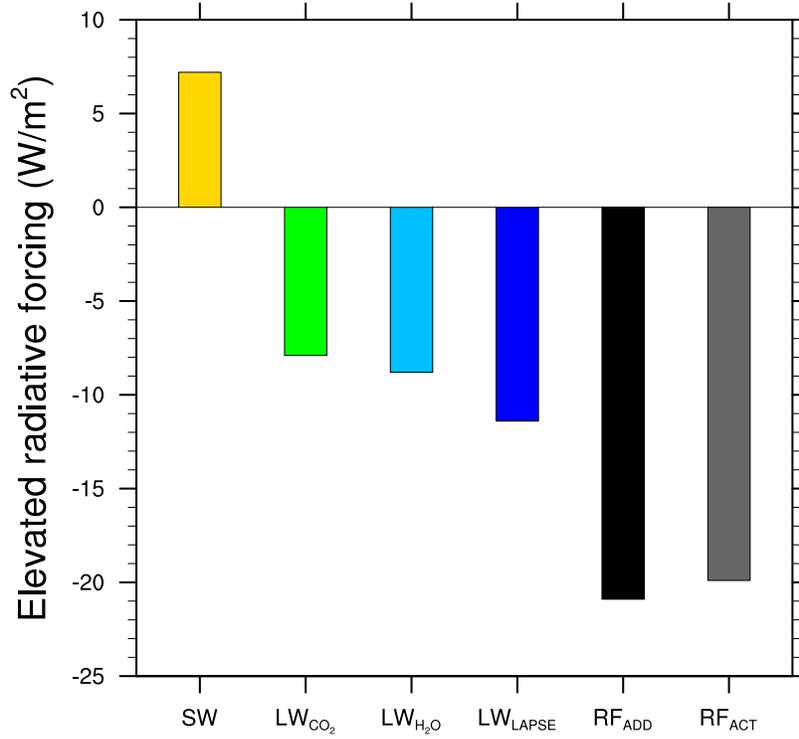


Figure 3: Elevated radiative forcing when the surface is elevated from 1000 hPa to 500 hPa with a fixed albedo of 0.24, for each term in Eq. (14). RF_{ADD} is for the addition of the four terms, and RF_{ACT} is the actual radiative forcing change with all the aforementioned effects and their interactions.

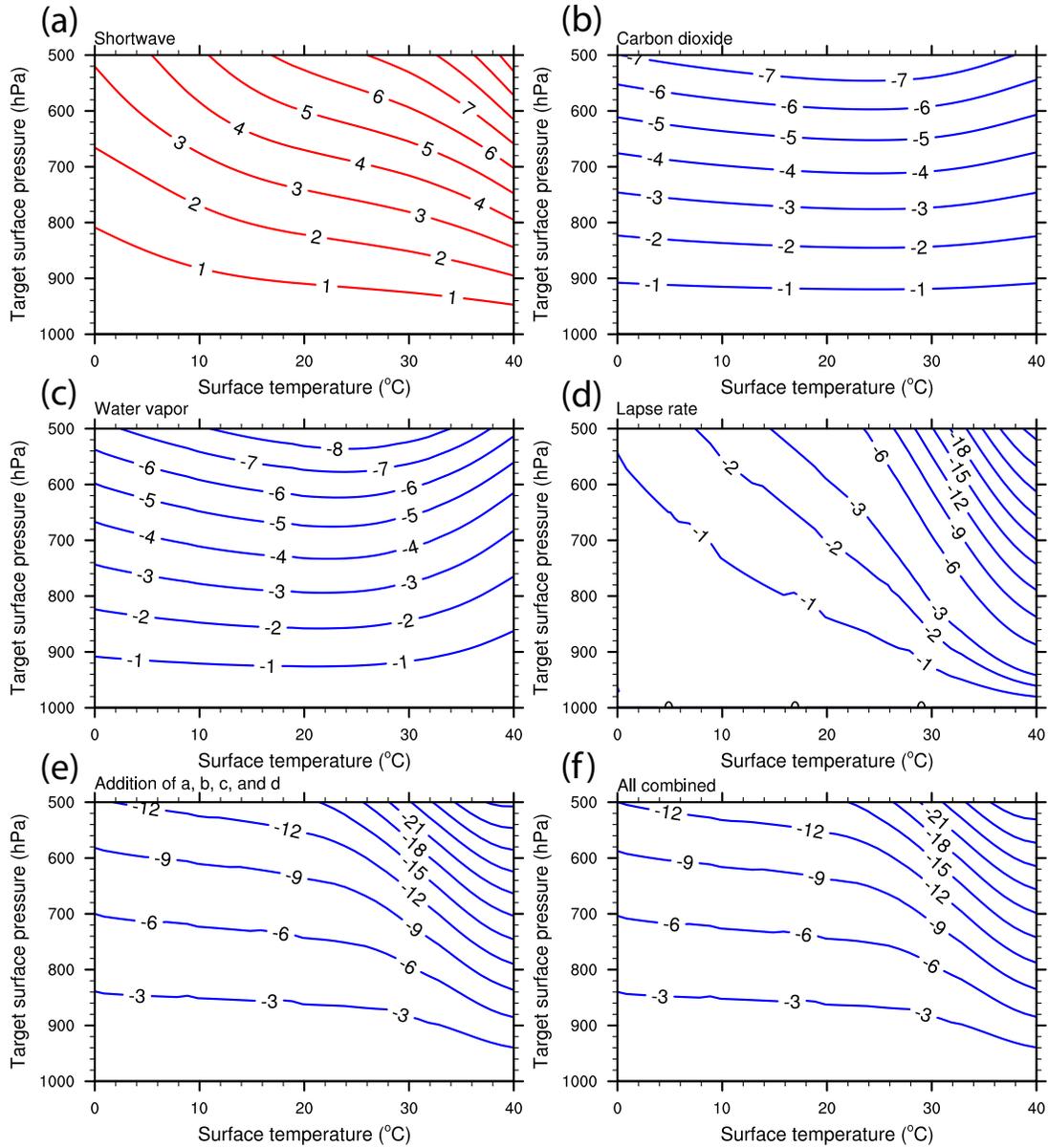


Figure 4: Contributions to the net top-of-atmosphere (TOA) radiative flux change when the surface is elevated from the reference level (surface pressure 1000 hPa) to the target level indicated on the vertical axis while surface temperature is fixed at value indicated on the horizontal axis. Contributions are from (a) shortwave, (b) longwave due to changes in carbon dioxide optical path, (c) longwave due to changes in water vapor optical path at a fixed lapse rate, and (d) longwave due to lapse rate changes. Also shown are (e) the sum of (a-d), and (f) the net TOA radiative flux change including all aforementioned effects and their interactions. Positive values indicate a warming effect and negative a cooling effect. The physical meaning of each term and the method of calculation are described in Section 5.

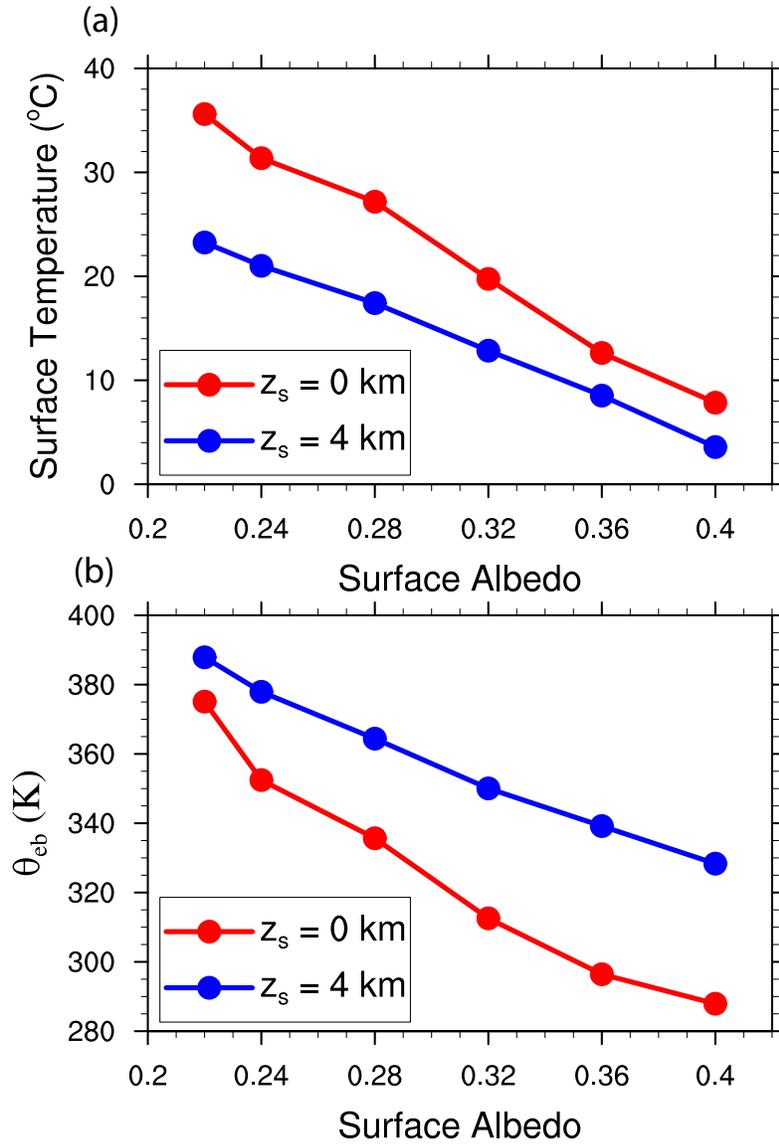


Figure 5: (a) Surface temperatures and (b) low-level equivalent potential temperatures over non-elevated surfaces (red) and over surfaces with 4 km elevation (blue), for a broad range of surface albedos, all from cloud system-resolving simulations of radiative-convective equilibrium. Note the greater surface temperature drop and the weaker elevated heating effect in warmer climates (i.e. those with lower surface albedos).

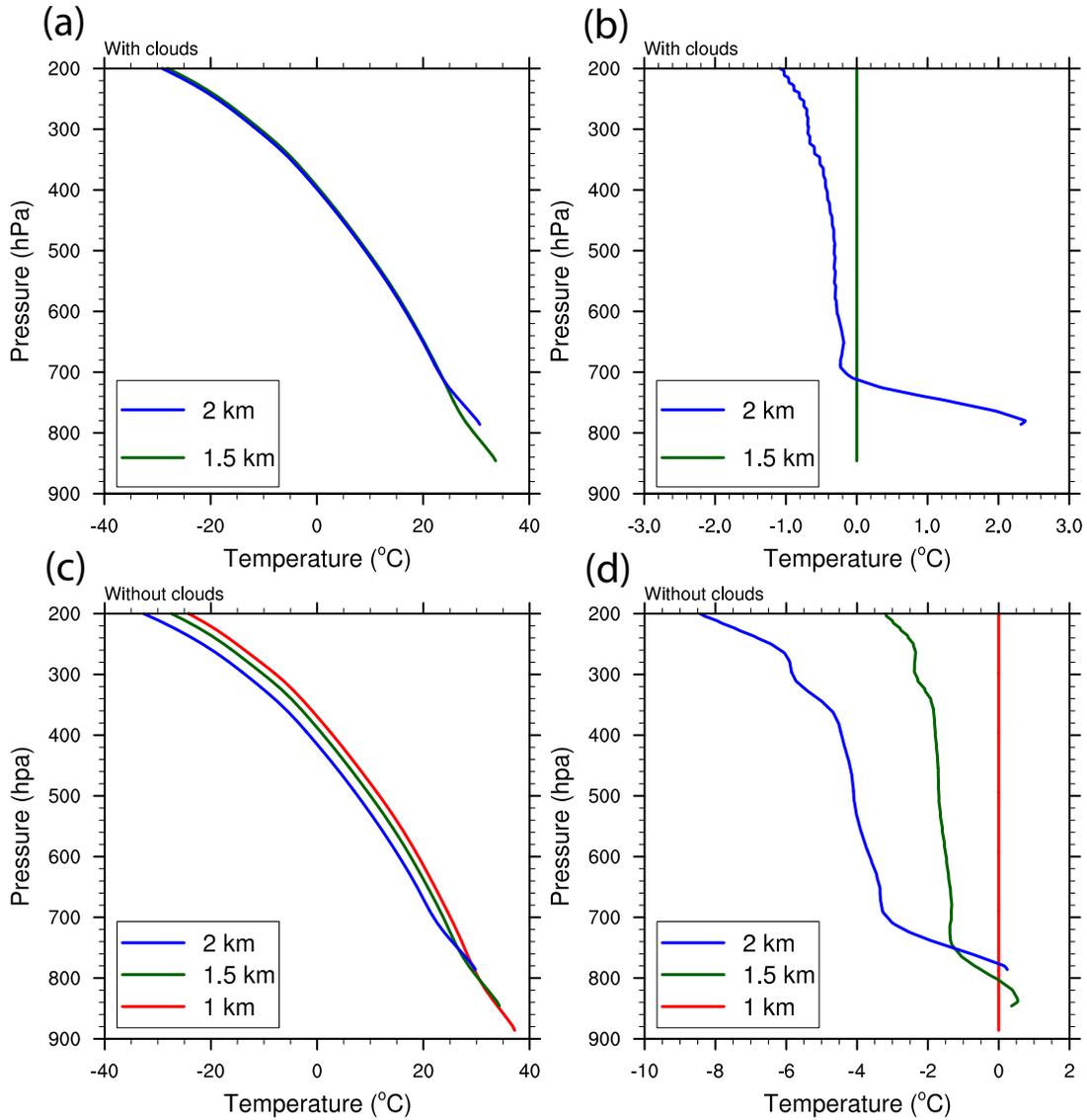


Figure 6: (a,b) Vertical profiles of (a) temperature and (b) temperature anomaly with respect to the 1.5 km case, for the simulations with surface albedo of 0.2; the legend indicates the surface elevation used for each curve. (c,d) Same as (a,b) but without radiatively-active clouds; temperature anomaly is with respect to the additional 1 km case.