JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 104, NO. D20, PAGES 24,265-24,271, OCTOBER 27, 1999

Temperature profiles in radiative-convective equilibrium above surfaces at different heights

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Abstract. Soundings of temperature and specific humidity calculated assuming radiative-convective equilibrium over surfaces at different heights show only small variations in surface temperature and specific humidity with surface height. Calculated surface temperatures decrease at ~2°C km^{-1}, with all other parameters held fixed. Calculated temperatures aloft in the upper troposphere above surfaces at different heights differ by 10°-20°C; those above surfaces at 3000-5000 m are significantly warmer. These differences support the contention that an elevated terrain can force significant atmospheric circulation through its effect on equilibrium temperatures, as occurs in the case of the Tibetan Plateau forcing the south Asian monsoon.

1. Introduction

Conventional wisdom, buttressed by a few measurements, holds that temperatures in the atmosphere over elevated regions should be higher than those in the free atmosphere over surfaces at sea level with the same solar radiation [e.g., *Barry*, 1992, p. 57]. Because of such higher temperatures, high plateaus, like Tibet, are often assigned major roles in large-scale atmospheric circulation, such as for the Asian monsoon [e.g., *Flohn*, 1974; *Hahn and Manabe*, 1975: *Kutzbach et al.*, 1993]. Indeed, nowhere in the world do temperatures at pressures of 200-500 mbar exceed those in summer over Tibet [*Li and Yanai*, 1996], where surface pressures lie between 500 and 600 mbar. Yet what processes combine to make temperatures over high plateaus so high remains an open question.

A thin atmosphere over a high plateau contributes directly to the temperature structure in two opposite ways. Less scattering and reflection of short wavelength radiant energy should heat an elevated surface more than a lower one. At the same time, the reduction in absorbers of long wavelength, infrared energy should decrease the "greenhouse effect," whereby much of the heating of the surface is by reëmission of infrared radiation by the atmosphere. According to *Barry* [1992, p. 57], *Flohn* [1953] first suggested that high areas should be hotter than air at the same pressure but over lower surfaces, in part, because of the more efficient transmission of short wavelength radiation to the higher surface but, more importantly, because the divergence of the infrared radiative flux should be nearly constant with pressure in the troposphere. Heat is transferred from the surface to the atmosphere not just by radiation, however, but also by conductive, sensible heating of the overlying atmosphere and by evaporation from the surface. Because infrared absorption (and emission) depends mostly on the concentration of water vapor in the atmosphere, and specific humidity depends strongly on the temperature of the air, water vapor is concentrated near the bottom of the troposphere. If specific humidity above elevated surfaces were significantly less than that above surfaces at sea level, the greenhouse effect might be substantially reduced over a high terrain. Thus perhaps it is not intuitively obvious that the atmosphere over an elevated surface should be hotter than that of the free atmosphere over a lower surface.

One can imagine limiting conditions that provide some guidance for accounting for contributions by different processes. In the (absurd) limit in which all absorption took place high in the atmosphere, such that convection mixed the underlying atmosphere to maintain a dry adiabatic lapse rate below, the temperature at any height, or pressure, would be independent of the height of the surface. At the other extreme, in which the atmosphere was nearly transparent to all radiation, the temperature at the surface would be close to its value at black-body equilibrium and thus independent of the height of the surface. For this extreme, air above an elevated area would be much warmer than that in the free atmosphere at the same pressure above a lower region. In general, surface temperatures should lie somewhere between these extremes of decreasing with surface height at the dry adiabatic lapse rate and being constant, independent of surface height. In fact, *Tabony* [1985] noted both extremes: surface temperatures in winter across the high plains of Canada near 50°N decrease with surface height at 9°C km^{-1}, which is nearly equal to the dry adiabatic lapse rate of 9.8°C km^{-1}, but winter surface tempera-

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Paper number 1999JD900485
0148-0227/99/1999JD900485$09.00

tures for the central United States at 40°N are nearly constant in the elevation range from 1600 to 270 m.

To understand what might make the atmosphere above a plateau hotter than the free atmosphere above a lower region, we calculated soundings for radiative-convective equilibrium of atmospheres above surfaces with different heights and therefore with different surface pressures. As should be clear below, because of the large number of parameters that must be assumed, we did not attempt to simulate temperature and moisture profiles for specific regions. In fact, we used annual averages of incident radiation and ignored not only diurnal but also seasonal variations. Our purpose has been to examine how a sensible scheme for making such calculations responds to changing surface elevations without changing other, in some cases arbitrarily, assigned parameters. One-dimensional (1-D) radiative-convective equilibrium may never be realized in the Earth's atmosphere, for a circulating atmosphere can transport heat and moisture laterally before equilibrium is reached. The main justification for considering radiative-convective equilibrium is that it quantifies the state toward which radiative and convective processes drive the atmosphere.

2. Radiative-Convective Equilibrium Scheme

We used the formulation of *Emanuel* [1991, 1994], modified by *Emanuel and Zivkovic-Rothman* [1999]. As in most such formulations [e.g., *Rennó et al.*, 1994a,b], solar energy impinging on the Earth is parameterized in terms of a solar constant, a zenith angle, which depends on latitude, and a value for the surface albedo. The surface is heated, and both heat and moisture are transferred to the overlying atmosphere by mechanical turbulence proportional to a specified wind speed and to differences in temperature and in the mixing ratio between the surface and air immediately overlying it. The atmosphere is divided into layers, in our case 46 layers for a surface pressure of 1000 mbar, with intervals of 25 mbar in the troposphere and then decreasing toward the top of the stratosphere. (*Emanuel and Zivkovic-Rothman* [1999] showed that a vertical grid spacing of no more than 50 mbar is necessary to resolve water vapor in the atmosphere.) Heat is also radiated upward by the surface, and at all levels in the atmosphere it is radiated equally upward and downward according to the Stefan-Boltzmann law for the temperature and the emissivity at the level. Both absorption and emission are dictated by the concentrations of water vapor, CO_2, methane, and ozone. We used the parameterization of *Chou et al.* [1991] to calculate absorption and emission spectra. We specified concentrations of CO_2, methane, and ozone at each level in the atmosphere, but the fraction of water vapor was adjusted by the cumulus convective scheme. As shown by *Rennó et al.* [1994a,b], radiative-convective equilibrium achieved with an interactive hydrological cycle differs in important ways from earlier calculations of equilibrium, such as those of *Manabe and Strickler* [1964] and *Manabe and Wetherald* [1967], that prescribed the distribution of relative humidity.

Emanuel's [1991] scheme focuses on accurate prediction of water vapor distributions. Several processes occurring within a convecting region affect the rate at which and height to which water vapor is carried aloft. In order to take into account microphysical processes that modify the transport of water in adjacent updrafts and precipitating downdrafts, fractions of air at each level must be considered separately, at least at levels where clouds can form. First, buoyant air rises and entrains adjacent air with which it mixes. Mixing is parameterized in terms of vertical gradients of buoyancy and a constant entrainment factor λ (= 0.06 $mbar^{-1}$) to weight differences in pressure between layers (see *Emanuel and Zivkovic-Rothman* [1999] for details). As unsaturated air rises from the subcloud layer to the lifted condensation level and cools adiabatically, vapor condenses into droplets or ice crystals. Part of the condensate is converted into precipitation, and the remainder forms, or augments, a cloud. We assume that precipitation forms only when the amount of condensate exceeds an assigned level l_c. The amount converted into precipitation equals the difference between the amount of water in the layer and l_c, with allowance for ice made crudely by letting l_c depend on temperature [*Emanuel and Zivkovic-Rothman*, 1999]. (This approach differs somewhat from *Emanuel's* [1991], which employed a precipitation efficiency to specify a fraction of liquid water that falls.) The downdrafts associated with the precipitation, in turn, comprise only a fraction, σ_d (~ 5%), of the area represented by the 1-D atmospheric column. Moreover, only part, σ_s (~ 15%), of the precipitation is assumed to fall through unsaturated air. The rate of reëvaporation of precipitates is specified in terms of the pressure, temperature, amount of condensed water, and the water vapor mixing ratio in the layer, the last of which must be estimated. Because evaporation cools the air, it can induce a downdraft of what has become unsaturated air, which leads to the familiar phenomenon of a cool breeze preceding a rain storm. Downdraft speeds are governed by (specified) terminal speeds of rain or snow

Table 1. Commonly Used Parameters in Calculations of Radiative-Convective Equilibrium

Parameter	Value
Solar constant, W m^{-2}	1382
Latitude	10°
Surface albedo	0.20-0.46
Concentration of CO_2, ppm	330
σ_d	0.04
σ_s	0.15
Mixed layer depth, m	1.0
Surface wind speed, m s^{-1}	7.0
Time step, min	10
Frequency of radiation calls, hours	6

(Table 1). The transfer of both sensible and latent heat from the surface depends on the wind speed at the surface. To allow for the effect of convection, the assigned background average wind speed (7 m s^{-1} in our runs) was augmented, by a root-mean-square average downdraft speed, which was taken proportional to the downdraft mass flux divided by σ_d.

We used parameter values close to those obtained by *Emanuel and Zivkovic-Rothman* [1999], who exploited temperature and moisture budgets from the Tropical Ocean-Global Atmosphere/Coupled Ocean-Atmosphere Response Experiment (TOGA/COARE) to optimize them. Temperature, specific humidity, and liquid water amount were calculated iteratively under the constraints that the mass-integrated total water and integrated enthalpy over the column be conserved. We used time steps of 10 min between calculations of temperature and mixing ratio at each level. Because radiative heating and cooling is both relatively slow in the atmosphere and relatively time-consuming to calculate, we made such calculations every 6 hours during a run, using the annual average of incident shortwave radiation from the annually averaged zenith angle. We sought steady state conditions using annual average values for solar radiation. Moreover, although clouds clearly play an important part in the calculation of water content in the atmosphere, we made no attempt to guess the feedback between clouds and albedo, which remains an open question. To effect changes in the incident solar radiation, due to differences in latitude or planetary albedo, we varied the assigned value for surface albedo.

Most experiments that have been carried out using *Emanuel's* [1991] scheme, with or without subsequent modifications, have employed a saturated surface [e.g., *Rennó*, 1997; *Rennó et al.*, 1994a,b]. Because oceans do not cover high terrain, we included a coefficient to allow for less latent heat transfer over land surfaces. To simulate surfaces with different degrees of saturation, we multiplied potential evaporation by an evaporation fraction β: (1) β = 1 to allow evaporation of all potentially available moisture from a sea surface, (2) β = 0.5 and (3) β = 0.25 to limit evaporation to half and to a quarter of that potentially available, and (4) β = 0 to prevent evaporation and allow only sensible heat transfer. Because we do not examine seasonal variations of radiation, whose effect on the temperature of the Earth's surface is delayed by its thermal capacity, we used a thin mixed layer (1 m) for the surface layer of the Earth.

3. Results

Before describing the study of calculated temperatures over surfaces at different heights, we summarize tests made to evaluate their precision. First, we sought calculations appropriate for steady states, even if annual variations in radiation render the concept of steady state somewhat irrelevant. In all cases, we calculated the evolution of soundings of temperature and moisture in the atmosphere without including daily or annual variations in solar radiation. Runs lasted at least 1000 days, and longer for many cases. For the vast majority, a steady state was reached after a few hundred days, with the longest times needed for the warmest and wettest environments. Although a few hundred days might seem to be an excessively long period, calculations including daily and annual variations yielded a repeatable annual cycle after 1 year. Thus the long duration stems more from initialization than from a misrepresentation of the atmosphere. For a fraction of runs, surface temperatures did not reach steady state but instead fluctuated, usually periodically. Only in rare cases did amplitudes of fluctuations reach 1°C. Periods varied from days to months, and the form of oscillation commonly was more sawtooth than sinusoidal, but we could recognize no pattern in this oscillatory behavior that correlated with assumed physical parameters. (Following a suggestion of *Rennó* [1997, personal communication, 1999], we found that adding minor diffusion between layers reduced such irregularities to <0.1°C. Calculations presented here, however, do not include such diffusion.)

In general, final calculated soundings, begun from different initial soundings, differed little, by less than a few tenths of a degree Celsius. In tests started from vastly different initial soundings (with surface temperatures 15°-20°C greater and less than the final), some calculated soundings differed by as much as 1°C, to almost 2°C near the top of the troposphere, and differences in calculated specific humidity approach 1 g kg^{-1}. *Rennó* [1997], in fact, reported examples of multiple equilibrium states for some calculations of radiative-convective equilibrium, but differences between temperature profiles for such states were much larger than a few degrees Celsius. In a few cases, and in particular for very high surface temperatures, larger differences in calculated temperatures (but <4°C) and in specific humidity (but <5 g kg^{-1}) were obtained with different start-

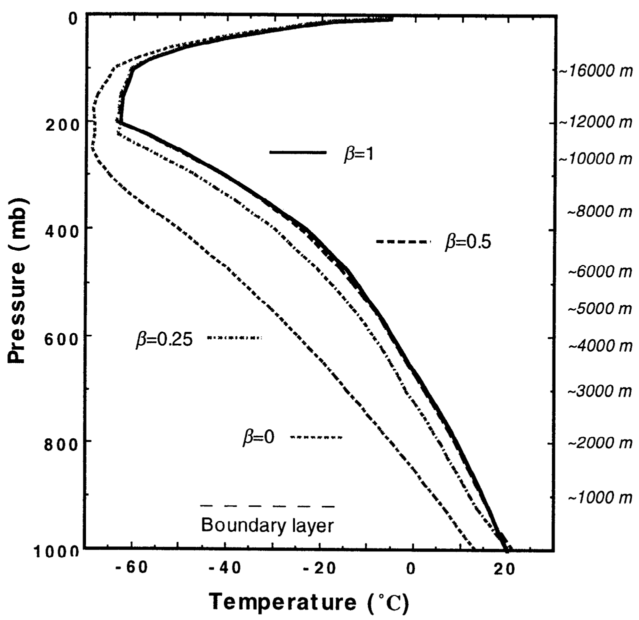

Figure 1. Plots of calculated temperature profiles for different values of β. We used a surface albedo of 0.32 and parameters in Table 1.

ing conditions. From the majority of calculations, however, differences in calculated soundings for different initial conditions imply representative errors (in precision) in temperature of 1°C and in specific humidity of <1 g kg^{-1}. (Again, a small amount of diffusion between layers reduced such discrepancies to <0.1°C.)

Calculated temperatures over sea-level surfaces for $\beta = 1$ (evaporation of all potentially available moisture from the surface, solid line in Figure 1) and $\beta = 0.5$ (half of that amount, long-dashed line in Figure 1) differ by only ~1°C, though they can differ by as much as 5°C for surfaces at 4-6 km ($P_{surface}$ = 600-500 mbar). For less available moisture ($\beta = 0.25$, dash-dotted line in Figure 1), the drier atmosphere makes the air aloft a few degrees colder than for a wetter surface. When no latent heat transfer is permitted ($\beta = 0$, short-dashed line in Figure 1), calculated values of specific humidity are necessarily low, and temperatures must therefore also be much lower than when abundant moisture in the atmosphere absorbs infrared radiation. Note that when $\beta = 0$, any moisture above the dry adiabatic layer is vestigial from the initial conditions and therefore is somewhat arbitrary. Because specific humidity is strongly dependent on temperature, plots of specific humidity are not very revealing. Instead, we show calculations of relative humidity for two surface elevations (Figure 2). When moisture is available, the three lines are similar, with high values in the boundary layer, a minimum high in the troposphere, and another maximum just below the tropopause. When $\beta = 0$, the upper part of the relative humidity profile is similar to those with large values of β.

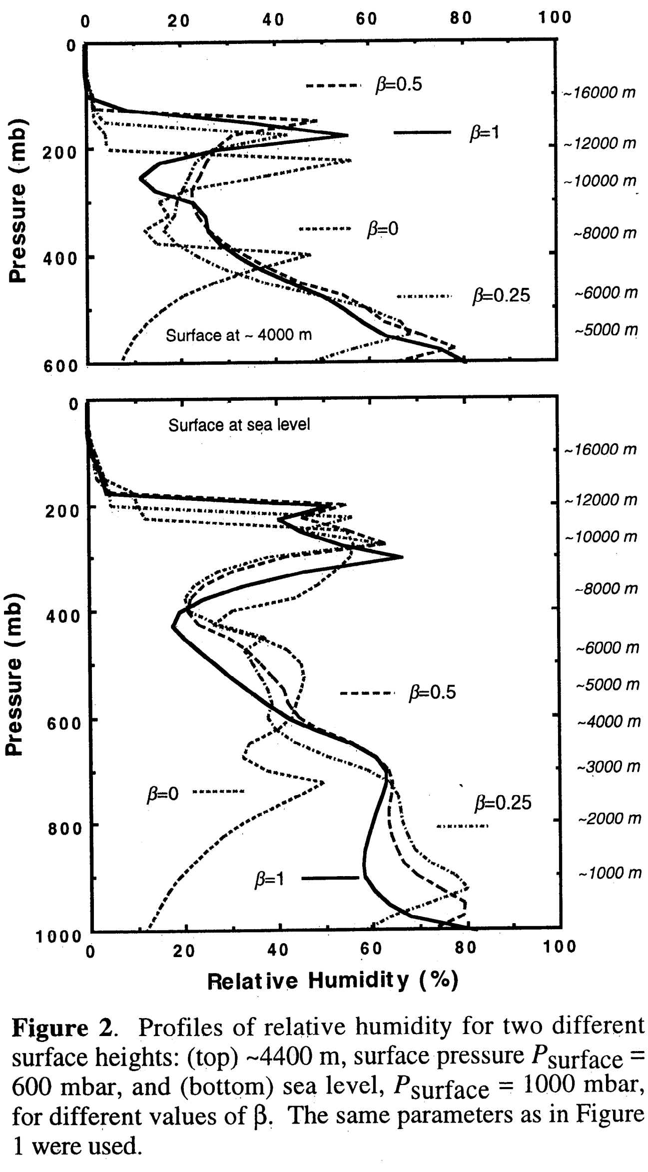

Figure 2. Profiles of relative humidity for two different surface heights: (top) ~4400 m, surface pressure $P_{surface}$ = 600 mbar, and (bottom) sea level, $P_{surface}$ = 1000 mbar, for different values of β. The same parameters as in Figure 1 were used.

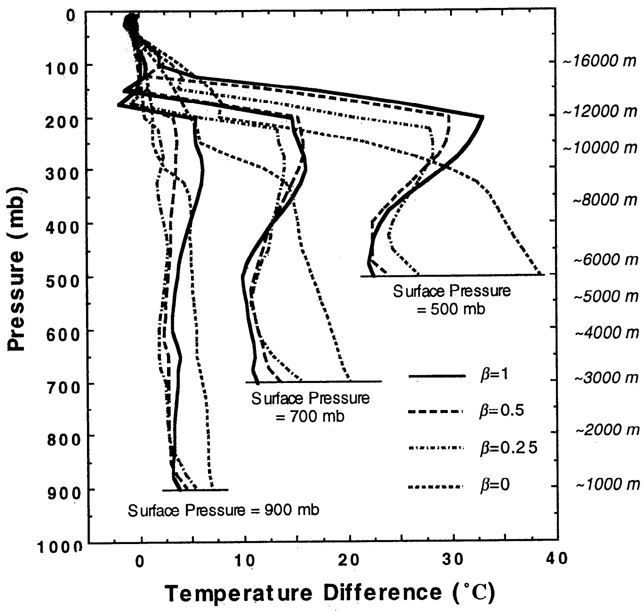

Figure 3. Plots of differences between calculated temperatures above elevated surfaces and those above sea level for different values of β and for different surface heights: $P_{surface}$ = 900 mbar (~1000 m), 700 mbar (~3000 m), 500 mbar (~6000 m). We used a surface albedo of 0.32 and parameters in Table 1.

Calculated temperatures and values of specific humidity just above surfaces at different heights differ little. Thus calculated temperatures of air above them are significantly greater than those at the same pressures but in an atmosphere overlying a surface at sea level. For each value of β we subtracted calculated temperatures above a surface at sea level from those calculated over higher surfaces (Figure 3). Except for $\beta = 0$, such differences increase monotonically as surface height increases (as surface pressure decreases) to reach maxima in the upper troposphere at pressures between 200 and 300 mbar. Moreover, these maximum differences are greatest for a saturated lower boundary and decrease as potential evaporation is suppressed and the equilibrium lapse rate approaches a dry adiabat.

When no latent heat transfer at the surface is permitted ($\beta = 0$), differences between calculated temperatures at all levels are greater for higher surfaces than for lower surfaces, but in the upper troposphere the difference is not as great as when latent heat transfer is permitted (Figure 3). Differences of temperatures just above elevated surfaces from those at the same pressures in the free atmosphere above a

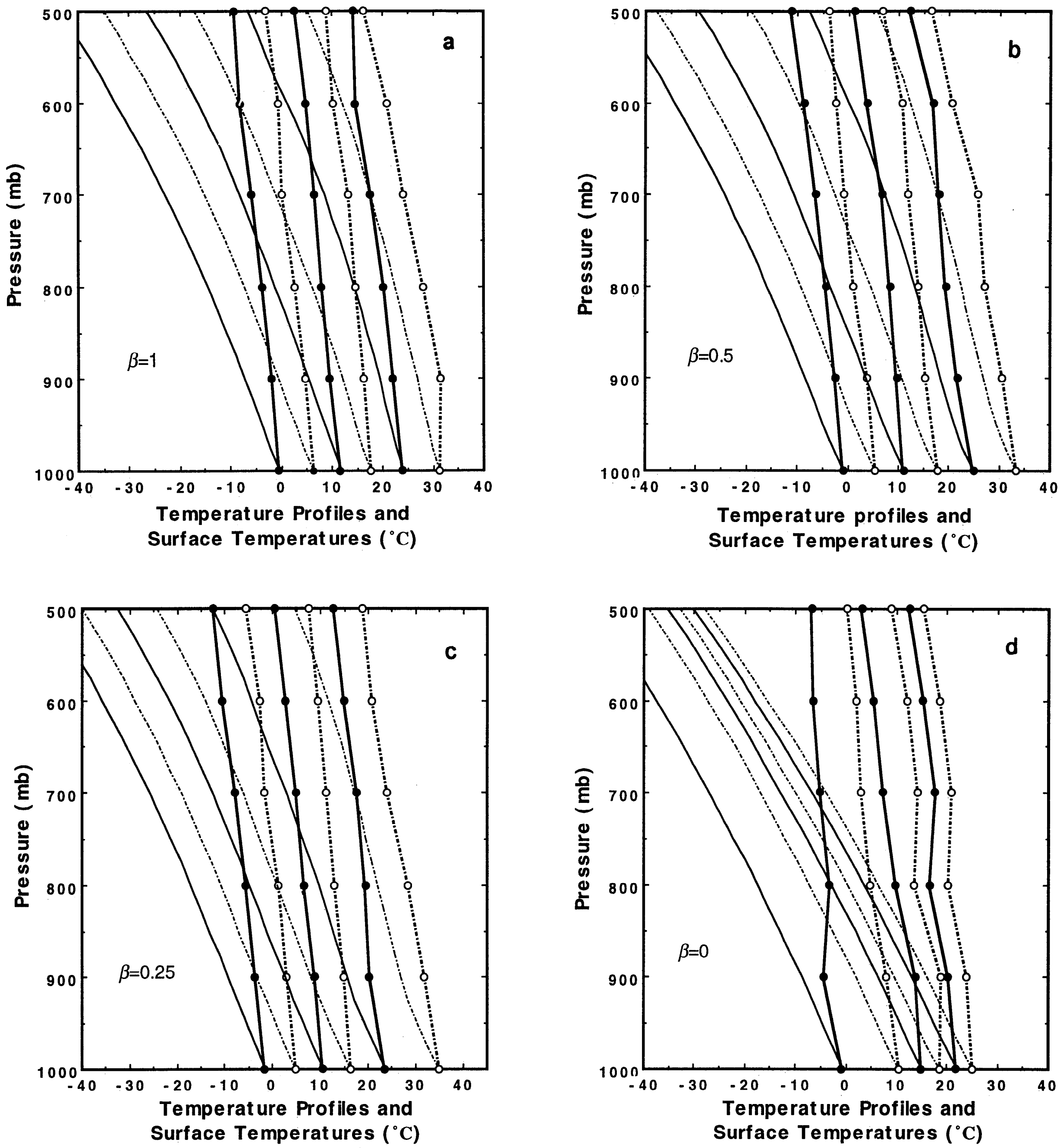

Figure 4. Plots of calculated temperature profiles above surfaces at sea level (thin solid lines) and of temperatures just above elevated surfaces (at the base of the corresponding atmospheric boundary layer) for different values of β: (a) $\beta = 1$, (b) $\beta = 0.5$, (c) $\beta = 0.25$, and (d) $\beta = 0$. Thick dash-dotted lines connect circles corresponding to the same value of albedo used, with alternating open and closed symbols used to distinguish different values. In Figures 4a-4c, values of albedo range from 0.26 to 0.46, and in Figure 4d they range from 0.18 to 0.38, in steps of 0.04. All other assumed parameters are given in Table 1.

sea level base, however, are notably larger when $\beta = 0$ than for $\beta = 0.25$ or more. Two aspects of the profiles without latent heat transfer contribute to the large differences. First, a nearly dry adiabatic lapse rate at all heights above surfaces with no potential evaporation make for very cold free-air temperatures aloft (Figure 1). Second, the combined decrease in short wavelength absorbers in a thinner atmosphere and lack of latent heat transfer increase surface heating. Thus, differences between temperatures immediately above elevated surfaces from those at the same pressure aloft above sea-level surfaces increase not only with surface height but also with decreasing latent heat transfer (Figure 3). This contrasts with the increased differences between temperatures at the same pressure high above surfaces of the same elevation as latent heat transfer increases.

The small variation in temperatures at the base of the atmospheric boundary layer (immediately above the surface) as a function of surface height (Figure 3) applies to a wide range of incident radiation and hence to a wide range of surface temperatures (Figure 4). Whereas free-air temperatures commonly follow gradients between 0.04°C $mbar^{-1}$ (~5°C km^{-1}) and 0.06°C $mbar^{-1}$ (~7°C km^{-1}), gradients of calculated bottom boundary-layer temperatures (essentially surface temperatures) commonly are closer to 0.02°C $mbar^{-1}$ (~2°C km^{-1}). Exceptions arise for particularly hot surfaces (runs with small albedo), for which calculated steady-state temperatures show a marked sensitivity to initial conditions. With $\beta = 0$ (Figure 4d), multiple runs, again in rare cases, indicate a dependence on initial conditions and possible instability in the calculations but, in this case, for very cold conditions (and very low specific humidity).

4. Discussion

Explaining an equilibrium state dictated by conservation laws always challenges arguments that rely on cause and effect, whether that state be one of steady large-scale circulation [e.g., *Emanuel et al.*, 1994] or the state of stress in a slowly deforming viscous medium [e.g., *England and Molnar*, 1993]. One cannot assign the "cause" of the state to one particular mode of energy transfer or one component of stress when a balance must exist. Thus *Flohn's* [1953] argument that infrared radiation will be independent of pressure cannot alone account for a difference in calculated temperature structures in radiative-convective equilibrium, for such an argument assumes no difference in convective heat transport. In fact, our calculations indicate a higher rate of infrared cooling per unit mass over high surfaces than over low surfaces and a greater convective heat transfer from higher surfaces than from lower surfaces.

The degree to which the surface temperature exceeds the black-body equilibrium temperature depends on the total mass of infrared absorbers in the atmosphere. As the mass of atmosphere above the surface is reduced, the total mass of infrared absorbers also decreases, provided that the mixing ratio of absorbing gases, such as water vapor, does not change. Thus, when the surface is elevated, the surface temperature will, in general, decrease. For the Earth the atmospheric column above sea level is sufficiently transparent to infrared radiation that elevating the surface results in only a small reduction in surface temperature. In particular, this rate of reduction is substantially smaller than the moist adiabatic lapse rate, which approximates well the equilibrium temperature profile. Thus, the equilibrium temperature over an elevated surface will always be higher than that at the same pressure over a surface at sea level.

Simple aspects of the hot and cold ends of the spectrum allow some insight into why surface temperature is only mildly sensitive to the elevation of the surface (Figure 5). Where the surface temperature is low, there is relatively little water vapor in the atmospheric column. Thus the reduction of column water vapor associated with a smaller

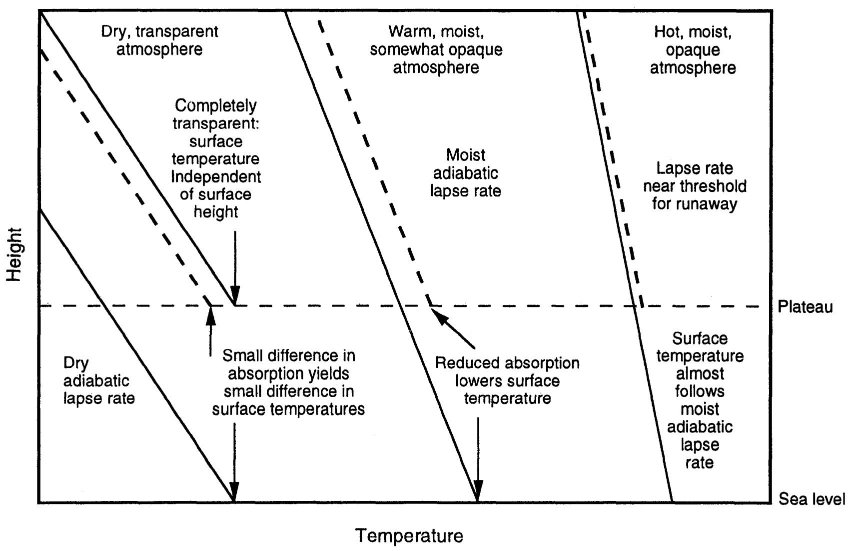

Figure 5. Simple cartoon illustrating effects of different surface elevations on surface temperatures. (left) At low surface temperatures with a dry adiabatic lapse rate, absorption and emission of infrared radiation is weak. Thus both direct and infrared radiation to the surface depend weakly on surface elevation, and surface temperature is nearly independent of height. (middle) For a warm, moist atmosphere the significantly greater infrared absorption and emission over low surfaces than over high surfaces would make those lower surfaces much hotter than higher ones, but convection transfers heat more effectively than radiation and reduces the surface temperature. (right) In the limit of a hot, moist atmosphere near a state of thermal runaway the moist adiabatic gradient limits the rate at which surface temperature can decrease with height. If temperature increased with decreasing height more rapidly than the moist adiabatic lapse rate, then, as temperature increased slightly, evaporation would also increase, leading to an increase of surface pressure, which would, unstably, further increase temperature and evaporation.

total column mass over a higher surface is comparatively small. Consequently, the surface temperature changes only slowly with surface height, as column mass decreases. Where temperatures are relatively high, the surface temperature calculated for radiative-convective equilibrium increases very rapidly with increasing column mass (or decreasing surface height) because each additional layer added at the surface increases the emissivity appreciably. Yet this rate of increase is limited by the moist adiabatic lapse rate, owing to convective adjustment. Thus the increase of surface temperature with surface height is limited by the moist adiabatic gradient, which is relatively small for hot, moist conditions. In the limit of a very hot, moist atmosphere, thermal runaway will occur: increased evaporation will increase the surface pressure, such that at its higher temperature, further evaporation will escalate warming and evaporation.

5. Conclusions

Calculated sounding of temperature and specific humidity in radiative-convective equilibrium for atmospheres over surfaces at different heights indicate notably higher temperatures at the same pressure above elevated surfaces than over sea level, with all other parameters held fixed. Calculated temperatures at the bottom of the atmospheric boundary layer decrease with the height of the surface at only ~2°C km^{-1}, a much lower rate than either typical moist or dry adiabatic lapse rates in the overlying atmosphere (~6°C km^{-1} and 10°C km^{-1}, respectively) and closer to isothermal.

Differences in calculated temperatures over surfaces at different heights approach maximum values in the upper troposphere, at pressures of 200-300 mbar, which is reminiscent of the large difference between measured temperatures in the upper troposphere over Tibet and over India during the summer monsoon [*Li and Yanai*, 1996]. Because large variations in horizontal gradients of temperature are required to drive meridional circulation [*Plumb and Hou*, 1992], the rise of the Tibetan Plateau seems likely to have played a major role in the initiation of the Asian monsoon. Thus our calculations support the assumption, commonly made and built into general circulation models employing coarse grids and simpler schemes for calculating radiative-convective equilibrium, that elevated terrain profoundly affects the temperature structure of the atmosphere above them.

Acknowledgments. We thank R. G. Barry, D. S. Gutzler, N. O. Rennó, and an anonymous referee for constructive comments on the manuscript. P. M. thanks the Cooperative Institute for Research in Environmental Sciences at the University of Colorado for support while this study was completed.

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(Received December 18, 1998, revised June 2, 1999; accepted June 22, 1999.)