Modeling and Observations of Martian Stationary Waves

LEV NAYVELT AND PETER J. GIERASCH
Department of Astronomy, Cornell University, Ithaca, New York

KERRY H. COOK
Atmospheric Science Program, Cornell University, Ithaca, New York

(Manuscript received 16 January 1996, in final form 16 September 1996)

ABSTRACT

A linear $\sigma$-coordinate model is used in conjunction with a Mars GCM northern hemisphere winter zonal mean circulation to study the orographically forced stationary disturbances on Mars. The disturbances are compared to their counterparts on Earth and to the bright streak data from the Mariner 9 and Viking images.

Diabatic heating on Mars is predominantly radiative, and inversion near the surface causes mountains to cool and depressions to heat the atmosphere. Due to the short radiative timescales and large meridional surface pressure gradient, the atmosphere responds to radiative forcing by changing its temperature and pressure gradient. The strong meridional flow near the equator and short radiative timescales combine to make responses to mechanical and thermal forcing of the same order in the tropics and in parts of the northern midlatitudes. In the subtropics, small zonal and meridional flows allow the response to radiative forcing to dominate. All these factors contribute to making the Martian atmospheric responses to orographic forcing different from the responses found on Earth.

The model near-surface winds forced by realistic topography agree well with the Mars GCM stationary wave winds and with the orientation of the bright streaks. The addition of stationary eddies to the zonal mean winds improves all measures of alignment, as does the inclusion of radiative forcing in the linear model. Using the dusty basic-state fields improves wind-streak alignment through a better match between the zonal mean winds and streaks. Even near Olympus Mons, linear stationary waves improve the wind-streak alignment when added to the zonal mean winds.

1. Introduction

Terrestrial stationary waves play an important role in the global circulation and have been the focus of much research. From the $\beta$-plane model of Charney and Eliassen (1949) to the primitive equation, $\sigma$-coordinate model of Hoskins and Karoly (1981), to the improved lower boundary condition, pressure coordinate model of Chen and Trenberth (1988), to the nonlinear model of Valdez and Hoskins (1991) and the GCM simulations of Cook and Held (1992) and Ringler and Cook (1995), there has been a steady improvement in the modeling and understanding of orographically forced stationary waves on Earth. Nevertheless, significant unknowns remain and include a lack of comprehensive understanding of the roles of the nonlinearity and diabatic heating in determining stationary wave structure.

On Mars, the understanding of topographically forced waves has been hampered by a lack of observations. Several authors, including Gierasch and Sagan (1971), Blumsack (1971), Mass and Sagan (1976), Gadian (1978), Webster (1977), Pollack et al. (1981), Hollingsworth and Barnes (1995), and Barnes et al. (1996) have predicted the importance of such waves using theory, linear models, and GCM simulations.

Conrath (1981) found evidence for a global wave in Mariner 9 IRIS temperature data that could be interpreted as a zonal wavenumber 2 stationary wave, but the data did not permit unique determination of both the wavenumber and period. Recently, Banfield et al. (1996) found time-independent, zonally asymmetric temperature variation in the Viking Infrared Thermal Mapper (IRTM) observations during Martian fall and spring.

The purpose of this paper is twofold. The first aim is to contribute to our theoretical understanding of how topography forces stationary waves in the Martian atmosphere by obstructing the low-level flow (mechanical forcing) and by perturbing the radiative heating distribution. The second purpose is to try to identify stationary wave structures in the bright streak data obtained from Mariner 9 and Viking images. A correlation between the modeled near-surface winds and bright streaks...

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Mars</th>
<th>Earth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radius (km)</td>
<td>3393</td>
<td>6378</td>
</tr>
<tr>
<td>Gravity (m s(^{-2}))</td>
<td>3.72</td>
<td>9.81</td>
</tr>
<tr>
<td>(\Omega) (10^{-5} \text{s}^{-1})</td>
<td>0.7088</td>
<td>0.7294</td>
</tr>
<tr>
<td>(R) (\text{J kg}^{-1} \text{K}^{-1})</td>
<td>192</td>
<td>287</td>
</tr>
<tr>
<td>(C_p) (\text{J kg}^{-1} \text{K}^{-1})</td>
<td>860</td>
<td>1000</td>
</tr>
</tbody>
</table>

\(\Omega\) = rotation rate.
\(R\) = gas constant.
\(C_p\) = specific heat at constant pressure.

provides some of the most direct evidence to date of stationary waves on Mars and sheds some light on the longitudinal variability of bright streaks.

In the next section, we give the model justification, model description, and the model parameters used in this study. In section 3, the stationary wave response to simple mountain shapes in a Mars general circulation model (MGCM) low-dust winter basic state is investigated. In section 4, the model response to the Martian topography is examined and compared to the MGCM winds. We compare the near-surface stationary response with the bright streaks and discuss the effects of the dust in section 5. Section 6 contains our conclusions.

2. Model

The numerical model used for this study is based on the linear, steady, primitive equation, \(\sigma\)-coordinate model described by Ting and Held (1990). It has nine sigma levels (\(\sigma = 0.025, 0.095, 0.205, 0.350, 0.515, 0.680, 0.830, 0.940, 0.990\)) and is a spectral transform model in the horizontal with rhomboidal truncation retaining 15 spherical harmonic wavenumbers (R15, with a resolution of 7.5° longitude and 4.5° latitude on the transform grid). The planetary and atmospheric parameters are set to Martian values (Zurek et al. 1992) and are listed along with the Earth parameters in Table 1.

A number of Earth modeling studies (e.g., Chen and Trenberth 1988; Valdez and Hoskins 1991) have highlighted the importance of including nonlinear effects in the mechanical forcing of stationary waves. Cook and Held (1992), for example, showed that linear solutions on Earth deviate significantly from fully nonlinear GCM solutions for mountains higher than 2 km. Nonetheless, linear simulations have proved to be essential in gaining insight into forcing of stationary waves on Earth. Since they can be decomposed to associate features of the full response with individual forcings, linear solutions allow us to better understand the nature of the forcing on Mars, and to compare forcing on different planets.

The range of validity of the linear assumption on Mars should be wider than on Earth. Cook and Held (1992) found that the breakdown of the near-surface linear response occurs when the slope of the mountain exceeds the slope of the zonal mean isentropes. Mars has very steep zonal mean topography with a slope opposite to the isentropic slopes (Fig. 1) between 15°S and 60°N, so the linear model should apply to taller mountains there. The consistency of the linear assumption will be checked by comparison between the linear solution and the bright streaks.

Another current issue in the development of our knowledge about stationary waves on Earth concerns the rate of diabatic heating. For orography embedded in weak low-level flow and/or in a moist environment, stationary waves may be generated more strongly by perturbations of the diabatic heating field than by mechanical forcing (e.g., Ting 1994; Ringler and Cook 1995). The diabatic heating field is complicated on the Earth; the sensible and latent heating are both potentially important, as are the radiative effects if cloud response is included. On Mars, the diabatic heating is simpler. As long as there are no dust storms, the heating is dominated by a radiative effect as discussed below. Since this radiative effect can be calculated, we are provided with a cleaner case than for Earth linear models, which have to take diabatic heating from more complex models or often inadequate observations.

a. Basic state

Mars wind observations are scarce, and the response of the linear model is sensitive to the basic-state temperature and velocity profiles near the surface. Thermal wind balanced basic states can be derived for higher latitudes using temperature data from Mariner 9 and Viking missions (Hollingsworth and Barnes 1995), but this method does not provide a well-determined mean flow for low latitudes. Such a basic state would also not
include the effects of surface friction and would not have meridional winds. For these reasons, we have chosen to use an MGCM-derived basic state.

We linearize the model about the time and zonally averaged circulations ($\bar{U}$, $\bar{V}$, $\bar{T}$, $\bar{\sigma}$, and log $\bar{p}$, or zonal velocity, meridional velocity, temperature, vertical velocity, and log surface pressure, respectively) from the National Aeronautics and Space Administration (NASA) Ames MGCM simulations by Haberle et al. (1993). The MGCM includes the smoothed Mars consortium topography, albedo, and thermal inertia data (available from the U.S. Geological Survey in Flagstaff, Arizona), as well as CO₂ condensation within the atmosphere and on the surface. The zonal mean fields are calculated by averaging over the last 10 days of the 50-day simulations that start from a resting isothermal atmosphere. More details on the MGCM simulations can be found in Haberle et al. (1993) and Barnes et al. (1993).

Low-dust, northern winter zonal mean fields (Haberle’s run 87.58) are used as the primary basic state, although the effects of the dust (Haberle’s run 90.02) are also examined. Greeley et al. (1993) suggest that the northern winter near-surface winds agree best with the bright streak pattern.

The basic states are linearly interpolated to the linear model sigma levels. The low-dust fields are depicted in Fig. 2. The zonal mean surface pressure can be read from the bottom of the plots, and the zonal mean vertical $\sigma$-coordinate velocity $\bar{\sigma}$ is set to zero everywhere.

Haberle et al. (1993) simulate several relevant differences between the low-dust Martian GCM basic state and the Earth northern winter zonal mean climatology.
On Mars, the temperature peaks (~250 K) near the surface at very high southern latitudes (~80°S) and decreases monotonically to 60°N (Fig. 2c). In contrast, on Earth, temperatures decrease toward both poles throughout the year. There is a strong temperature inversion on Mars in the northern midlatitudes, and there is an inversion near the surface at all latitudes between 60°N and 60°S. While Martian wind jet structure is similar to that of Earth’s middle atmosphere, there is a surface westerly jet in southern subtropical latitudes on Mars that has no counterpart on Earth. As on Earth, the Hadley cells on Mars are asymmetrical. In addition, over three-quarters of the mass supplied to the rising branch of the Hadley cell flows through the lowest few kilometers. On Earth, the flow occurs over a deeper layer. The mean meridional motion on Mars is roughly six times stronger than found on Earth (Haberle et al. 1993).

b. Dissipation

Dissipation in the vorticity, divergence, and temperature equations is parameterized by

\[
\frac{1}{\tau_R} + \nu \nabla^2 \zeta',
\]

\[
\frac{1}{\tau_R} + \nu \nabla^2 D',
\]

\[
\frac{1}{\tau_N} + \nu \nabla^2 T',
\]

where \( \zeta' \), \( D' \), and \( T' \) are the perturbation vorticity, divergence, and temperature, respectively. Here, \( \tau_R(\sigma) \) is the Rayleigh friction timescale, \( \tau_N(\sigma) \) is the Newtonian cooling timescale, and \( \nu \) is the biharmonic diffusion coefficient.

Rayleigh friction represents the effects of surface drag as well as wave breaking phenomena (a sponge layer) in the upper atmosphere. Near the surface, the timescales are set to

\[
\tau_R = .25 \text{ days} \quad \sigma = .99
\]

\[
= .7 \text{ days} \quad \sigma = .94.
\]

This results in a 1-km boundary layer column integrated drag with a timescale (assuming a height-independent basic state) of .5 days. As shown by Valdes and Hoskins (1989), this timescale is equivalent to a 1-km Ekman layer with a diffusion constant of 10 m² s⁻¹. This diffusion constant is typical for land on Earth. In the sponge layer, the timescales are set to

\[
\tau_R = 1 \text{ day} \quad \sigma = .095
\]

\[
= .5 \text{ days} \quad \sigma = .025.
\]

The Rayleigh damping time is set to two days in the intermediate model layers. The model–observation agreement discussed later in this paper decreases by 20% if the near-surface Rayleigh friction times are decreased or increased by a factor of 4. The agreement is insensitive to the value of Rayleigh friction timescale in the middle and the upper levels.

Newtonian cooling damps the temperature to \( \sigma \)-surface radiative equilibrium values \( T_E = T_E(\sigma) \), see appendix A]. The cooling timescales represent radiative effects and are calculated using an independent radiative model with equilibrium temperatures close to the mean basic-state temperature profile (appendix B). Figure 3 shows the coefficients derived for the low-dust, winter zonal mean circulation from the MGCM. Increasing the cooling times by a factor of 4 decreases the model wind-streak agreement by 10%. Decreasing the cooling times by the same factor increases the agreement by a few percent, and decreasing by a factor of 10 decreases the agreement by 10%.

Biharmonic diffusion, \( \nu \), is a numerical viscosity used to damp out coordinate resonances. It is set to \( 10^{16} \text{ m}^2 \text{ s}^{-1} \) (equivalent to damping the smallest resolved eddies on a 4-h timescale). This value is similar to some Earth studies (Valdes and Hoskins 1988) but is somewhat lower than others (Cook and Held 1992). The agreement between model winds and streaks studied later in this paper is insensitive to the increase or decrease of \( \nu \) by a factor of 10.
c. Forcing

The topographical relief on Mars is considerably larger than on Earth and exceeds the scale height of the Martian atmosphere in some areas. The one-quarter degree resolution Martian digital terrain model (DTM) topography (USGS–NASA 1993), smoothed to the linear model resolution by low-pass filtering in Legendre–Fourier space, is used in this study and is depicted in Fig. 4. The largest relief occurs for the Tharsis bulge (80°–120°W, 30°S–20°N), Olympus Mons (130°W, 20°N), Alba Patera (110°W, 40°N), Elysium (210°W, 25°N), and Hellas (290°W, 40°S). There are also numerous smaller topographical features.

The effects of topography on Mars include mechanical forcing, associated with obstruction of the low-level flow, and thermal forcing, associated with perturbations of the radiative heating field. On Earth, to a first approximation, mountains can be visualized as being embedded in the background thermal structure of the atmosphere, so the temperature at the surface of a mountain is the same as the atmospheric temperature at the same pressure level away from the mountain. On Mars, however, due to the short radiative timescales (Fig. 3), the surface is about the same temperature whether the surface is elevated or not (Webster 1977). Thus a mountain on Mars not only obstructs flow but acts as a heat source or sink. Except above the poles where CO₂ changes phase, latent heating is not important on Mars.

In the Martian linear model, topographic forcing is imposed through the geopotential ($\phi = gh$) at the lower boundary. The gravitational constant $g$ is assumed not to vary over the topographical height $h$. This is equivalent to the mechanical forcing of near-surface vertical velocity

$$w' = \nabla \cdot \nabla h' + \nu \frac{\partial h}{\partial y},$$

(4)

and (in the presence dissipation of temperature on $\sigma$ surfaces, see appendix A) thermal forcing at all levels,

$$Q = -\frac{1}{\tau_{\nu}} \frac{\partial \theta}{\partial z}.$$  

(5)

Here, $\nabla$ is the zonal mean horizontal wind vector and $h$ and $h'$ are the zonal mean and perturbation components of topography. There is also a less important frictional forcing due the dissipation of velocity on $\sigma$ surfaces (see appendix A) and a small forcing due to the thermal wind imbalance of the basic state. The first term on the right-hand side of (4) is the well-known linear mechanical forcing. The second term, which is negligible on Earth, is the forcing due to the zonal mean topography. If $\tau_{\nu}$ is a good representation of radiative timescales, (5) provides the linear radiative effects of topography.

The radiative effect described by (5) is omitted in Earth linear stationary wave studies (e.g., Valdez and Hoskins 1989; Hoskins and Karoly 1981) by applying...
Newtonian damping on pressure levels, that is, letting $T' \Rightarrow T' - \frac{3T}{(5p)} \log p$ in (3). The observed or modeled three-dimensional diabatic heating fields are used as thermal forcing instead. On Mars, there are no observations of Northern Hemisphere winter diabatic heating fields. Radiative heating dominates, and (5) is the most readily available method for representing the heating effects of topography.

The radiative heating per unit mountain height (in scale heights) is shown in Figs. 5a and 5b. An important and unique feature of this heating is that it is negative near the surface. This causes mountain cooling and depression heating and is opposite to the common radiative effect of topography on Earth. It comes about because of the near-surface inversion apparent in the basic-state temperature (Fig. 2c) on Mars. This inversion has no counterpart on Earth. The radiative heating fields near 2 mb and near the surface, forced by the Mars DTM topography, are shown in Figs. 5c.d.

The use of the Martian DTM topography and the zonal mean fields from the Ames MGCM simulations that use the Mars Consortium topography warrants some discussion. In a linear, $\sigma$-coordinate stationary wave model, the zonal mean and the zonally varying topography are handled separately. The zonal mean topography enters only through the zonal mean surface pressure term, which is in dynamic balance with the other zonal mean fields. The zonally asymmetric topography, on the other hand, enters as the lower boundary condition on the geopotential and provides the direct forc-
Fig. 6. Response to (a) total forcing, (b) mechanical forcing, (c) radiative heating, and (d) all other forcing for a 1-km Gaussian mountain at 45°N with a zonal $e$-folding scale of 15° and a meridional scale of 15/$\cos\theta$.

The forcing mechanisms examined are zonal mechanical forcing [zonal part of the first term in Eq. (4)], radiative forcing [Eq. (5)], and all other forcing (meridional mechanical forcing, thermal wind imbalance, zonal mean topography, velocity dissipation forcing, etc.). These were discussed in section 2c and are associated with the following components of the basic state or dissipation parameterization: near-surface zonal wind, damping of temperature on sigma levels, and the remainder of the basic state (meridional winds, winds above lowest level, zonal mean surface pressure, damping of wind on sigma levels, etc.). Several components are grouped together in the last item because, individually, they are associated with small responses. Together, they may contribute to a response on the same order as one of the first two components.

Care is taken to isolate the desired forcing. For instance, the traditional mechanical forcing is removed by subtracting the near-surface velocities from all model lev-
els to keep the vertical shear the same and by subtracting a geostrophically balancing surface pressure term to keep the geostrophic balance on the lowest layer the same.

The idealized mountains are all 1 km and Gaussian, with an $e$-folding scale of $15^\circ$ in the meridional direction and $15/\cos \theta_0^\circ$ in the zonal direction ($\theta_0$ is the latitude of the mountain top). They are placed at strategic locations to probe different regimes of the basic state. The horizontal extent of the mountains was chosen so that they are easily resolved by the R15 linear model and are geophysically relevant.

a. Northern hemisphere midlatitude mountain

The lowest $\sigma$-level eddy wind response to the full forcing associated with a mountain at $45^\circ$N is shown in Fig. 6a. The term full forcing is used to describe the model with unmodified MGCM basic state and dissipation parameterization as described in sections 2a and 2b. Figures 6b and 6c show the winds due to zonal mechanical forcing and radiative forcing, respectively. Figure 6d shows the winds due to all other forcing.

Near the surface, the full forcing response is dominated by the responses to the zonal mechanical and radiative forcing. In contrast to Earth (Ringler and Cook 1995), where the mechanical forcing dominates over the diabatic heating in midlatitudes, on Mars, the linear model suggests that the magnitudes of the two forcings as well as the responses to these forcings are comparable.

The nature of the zonal mechanical forcing response is similar to the baroclinic example of Hoskins and Karoly (1981). In the midlatitudes, poleward (equatorward) meridional advection to the west (east) of the mountain counteracts the adiabatic cooling (heating) due to rising (sinking) air on the mountain slopes (see Fig. 7 of Hoskins and Karoly 1981). On Earth, the response at these latitudes is more barotropic (Cook and Held 1992).

The response due to radiative forcing is dominated by the southwest–northeast flow due to the change in the meridional pressure gradient caused by the temperature perturbation. There is also a smaller barotropic response (vorticity and divergence created by vertical motions) that gives the flow anticyclonic curvature. Because the radiative timescales on Mars are short, the atmosphere responds to thermal forcing by changing its temperature rather than by advection across a temper-
ature gradient. This changes the density, and through the hydrostatic balance, the existing pressure gradient (e.g., due to zonal mean topography). The eddy winds depicted in Fig. 6c reestablish the geostrophic balance or flow down the gradient. This response is unique to Mars. On Earth, even if radiative heating were dominant, since the dynamical timescales are much shorter than radiative timescales, the atmosphere would respond to heating by advection.

The response of the model to all other forcing is dominated by the response to the thermal wind imbalance of the basic state. The noticeable contribution to the full forcing response is small and is limited to the southeast slope of the mountain.

b. Equatorial mountain

The full forcing response (Fig. 7a) for a mountain on the equator is dominated by the winds due to the mechanical and radiative forcing. At this latitude, however, meridional basic-state winds as well as the zonal winds are important. In contrast to Earth, where the response to the predominantly latent diabatic heating is dominant near the equator (Ringler and Cook 1995), on Mars, the response to the predominantly radiative diabatic heating is of the same order as that to mechanical forcing. The zonal mechanical forcing response (Fig. 7b) is fundamentally different than the response at midlatitudes. The flow is from an area of high pressure associated with colder air on the upslope of the mountain to the area of low pressure associated with hotter air on the downslope. This dipolar pattern is due to the negligible effects of the Coriolis force near the equator. Quasigeostrophic theory does not apply in the Tropics and neither does the explanation offered for the midlatitude response. A similar response was simulated by Webster (1972) using a two-layer Earth model.

The nature of the response to the radiative forcing (Fig. 7c), on the other hand, is similar to the response in midlatitudes. The barotropic response is now stronger but with no vorticity generation, and the flow in response to the enhanced meridional pressure gradient is purely meridional. The equatorial response to heating is very different from the classic (Webster 1972; Gill 1980) response found in many Earth simulations. This difference is due to the more baroclinic nature of the

Fig. 8. Same as Fig. 6 but for a 30°S mountain.
equatorial atmosphere on Mars. As in the northern midlatitude case, the short radiative timescales and strongly sloping zonal mean topography cause the Martian atmosphere to respond to heating at the equator by changing the pressure gradient instead of by advection.

The response to the meridional flow over the mountain dominates the response to all other forcing (Fig. 7d). As expected, in the absence of the Coriolis force, it is similar to the response to zonal flow (Fig. 7b) rotated by 90°. The flow is down the pressure gradient, due to the adiabatically cooled air on the northern upslope and hot air on the southern downslope. South of 20°S in Fig. 7, the flow is due to the leakage of the equatorial stationary waves into the barotropic region centered at 30°S.

d. Streamfunctions near 2 mb

On Earth, linear stationary waves are often studied by examining the eddy streamfunction in the upper atmosphere. The near-2-mb eddy streamfunctions due to simple isolated mountains on Mars are presented here for comparison. The streamfunctions are shown in Fig. 9. At all three latitudes, the responses consist of well-defined wave trains superimposed over m = 1 hemispheric patterns. The wave trains at 45°N and at the equator are qualitatively similar to the responses produced by simple mountains on Earth (Ringler and Cook 1995), while the hemispheric waves for the equatorial mountain are consistent with the response of Martian atmosphere to a simple (Tharsis-like) mountain studied by Hollingsworth and Barnes (1995). The upper atmosphere responses were broken into their components in a manner similar to the lower atmosphere responses. In the interest of brevity, only the total responses are shown. The localized wave trains are due to mechanical and radiative forcing, while the hemispheric patterns arise from the imbalance in the basic state.

For a mountain at 45°N, the well-defined wave train propagates into the Tropics and then dissipates (Fig. 9a). The direction of propagation is, however, more zonal than on Earth, and the wave train is coherent for at least 180° longitude. In contrast to the earth linear model, but in agreement with the Earth GCM (Ringler and Cook 1995), there is no polar wave train on Mars. The anticyclonic flow of the m = 1 hemispheric wave is over the mountain in both hemispheres, but the southern hemisphere flow is much weaker. The overall pattern has a strong wavenumber 2 component.

The upper-atmospheric response due to a mountain at the equator (Fig. 9b) produces two wave trains. The northern hemisphere wave train originates on the northern slopes of the mountain, reaches 40°N, and turns equatorward, reaching the equator slightly to the west of the mountain. The smaller, southern hemisphere wave train originates on the southern slopes of the mountain and propagates north until it merges with the northern wave train. The asymmetry in the wave trains in the two hemispheres is due to the asymmetry in the Mars winter basic state. The m = 1 disturbances are distorted by the interference with the localized wave trains. The anticy-
clones of these disturbances are to the west of the mountain in both hemispheres but are slightly out of phase.

The response to a mountain at 30°S is presented in Fig. 9c. There are two localized wave trains. One is zonal and the other penetrates the northern tropics. Both dissipate within 60° of the mountain and are embedded in an $m = 1$ disturbance. The anticyclone of the southern hemisphere global disturbance sits to the east of the mountain. The northern hemisphere $m = 1$ feature is exactly out of phase with the southern one.

e. Summary of idealized mountains

Several important differences between the stationary wave responses on Mars and Earth emerge from the simple mountain studies. While the nature of the responses to the traditional mechanical forcing is similar on the two planets, the responses to full forcing, near the surface and aloft, are quite different.

Near the surface, the difference comes mainly from the nature of the diabatic heating, the short radiative timescales, the strong meridional flow, the basic-state temperature inversion near the surface, and a strong meridional pressure gradient due to topography. Except above the polar caps, there is no latent heating on Mars and the diabatic heating is predominantly radiative. The radiative timescales are relatively short and meridional flow near the equator is strong. As a result, the responses to heating and to mechanical forcing are of the same order in the Martian Tropics as well as some regions of the northern midlatitudes. Due to the basic-state temperature inversion, mountains on Mars radiatively cool while depressions heat their surroundings. Short radiative timescales and a strong meridional pressure gradient fundamentally alter the way the Martian atmosphere responds to heating. A strong temperature perturbation forces a change in the meridional pressure gradient and directly alters the flow. On Earth, in contrast, latent heating dominates in the Tropics, mechanical forcing dominates in the midlatitudes (Ringler and Cook 1995), mountains heat their surroundings, and diabatic heating is balanced by temperature advection.
Fig. 11. Near-surface (lowest level) quasi-stationary eddy winds for the northern hemisphere winter solstice MGCM simulation with DTM topography (contours). Kindly provided by J. R. Barnes.

Fig. 12. Bright streaks on Mars (data kindly provided by Peter Thomas) and smoothed Martian topography contours for position reference.
Fig. 13. (a) Near-surface total winds, (b) eddy winds, and (c) zonal mean winds at bright streak locations superimposed on smoothed Martian topography contours.
In the upper atmosphere, Martian winter conditions are more favorable than Earth winter conditions to the excitation of the low wavenumber global modes. The sensitivity of the Martian atmosphere to excitation of global low wavenumber modes has also been observed by Webster (1977), Hollingsworth and Barnes (1995), and Barnes et al. (1996).

4. Martian topography results

The near-surface stationary wave winds resulting from a linear model simulation with MGCM basic state and the full Mars DTM topography are shown (along with the topography) in Fig. 10a. The winds are plotted on linear model grid points and show many of the features studied with idealized mountains.

North of 45°, the traditional zonal mechanical forcing dominates. The flows near valleys (30°W, 60°N) and (210°W, 65°N) and near elevations (90°W, 50°N) and (330°W, 50°N) are similar to the response to zonal mechanical forcing due to a simple midlatitude mountain on Mars (Fig. 6b) and on Earth (Fig. 3e of Ringler and Cook 1995). This region of Mars is very similar to its counterpart on Earth.

The northern hemisphere midlatitudes south of 45°, on the other hand, are dominated by the response to radiative forcing. The valleys centered around (160°W, 30°N), (40°W, 35°N), and (250°W, 35°N) as well as the elevated regions near (110°W, 40°N) and (300°W, 35°N) exhibit flow characteristics very similar to Fig. 6c (the valley flows being in the opposite direction to mountain flows, as expected). This behavior stands in sharp contrast to Earth, where mechanical forcing dominates everywhere in the midlatitudes (Ringler and Cook 1995). It is due to a strong surface pressure gradient combined with weak zonal mean zonal winds in the 20°–40°N latitude band.

In the equatorial and southern hemisphere regions, the idealized mountain response can also be observed in several regions. Near the elevations at (130°W, 15°N) and (110°W, 10°N) and near the valley at (270°W, 15°N), the flow is similar to the total response due to the equatorial mountain (Fig. 7a). The meridional mechanical and radiative forcing responses are of the same order here. Near the valley at (290°W, 40°S) the flow has the barotropic signature of the southern hemisphere ideal mountain response (Fig. 8). As expected, the valley flows are in the opposite direction from the mountain flows. The nature of the Martian equatorial and southern hemisphere responses is different from the responses observed on Earth. The reasons for the difference were discussed in section 3.

The near-2-mb eddy streamfunction obtained from the full topography experiment is depicted in Fig. 10b. A strong global, antisymmetric, zonal wavenumber 2 mode is clearly evident. In the southern hemisphere,
there is also evidence of a zonal wavenumber 1 mode. This wavenumber 1 mode dominates at high latitudes and at higher altitudes (not shown) in the southern hemisphere. The global modes persist in simulations with or without heating and with or without mechanical forcing. Webster (1977), Hollingsworth and Barnes (1995), and Barnes et al. (1996) found similar wavenumber 1 and 2 global modes in their simulations. Banfield et al. (1996) also saw the global modes in their analysis of the spring and fall equinox Viking IRTM 15 μm brightness temperature observations. While the models and the observations differ in season and in absolute phases of the modes, it appears that Martian atmosphere is very sensitive to wavenumber 1 and 2 forcing. We will examine the relationship between our model and the Viking data in a later work.

The model near-surface winds and the near-2-mb streamfunction can be compared to the MGCM winter quasi-stationary eddies presented and discussed by Barnes et al. (1996). Qualitatively, the linear model near-surface winds (Fig. 10a) and MGCM quasi-stationary winds (Fig. 11; Barnes 1996) agree quite well over most of the planet. The biggest differences occur near the highest elevations of Tharsis, a region of large heights and slopes, and even these discrepancies are exaggerated by the differences in resolution of the models.

The upper atmosphere responses of the two models are harder to compare as we use the near-2-mb stream-
function and Barnes et al. (1996) use the geopotential heights at 3 mb in their analysis. Qualitatively, however, both models produce a strong wavenumber 2 mode in the northern midlatitudes and a weaker wavenumber 1 mode in the southern midlatitudes. The longitudinal phases of the modes are very similar in the two models but the latitudinal positions of the peaks and valleys differ somewhat.

The good agreement between the linear model and the MGCM results suggests that the linearity assumption is appropriate for stationary waves in most of the the Martian atmosphere. This conclusion is further supported by comparison to bright streak data.

**5. Evidence of stationary waves in bright streaks**

The results of the linear model simulation can be compared to the bright streak data shown in Fig. 12 (Thomas et al. 1984). These streaks are thought to be formed due to dust being deposited in lee of obstacles on the Martian surface during the decay of the dust storms in late winter (Sagan et al. 1973; Greeley et al. 1974; Veverka et al. 1977). While some of the streaks get obliterated and reformed, their direction remains constant from year to year (Thomas and Veverka 1979).

The near-surface winds calculated at streak locations and resulting from a linear model simulation with MGCM nondusty basic state and the smoothed Mars DTM topography are shown in Fig. 13. The effects of using a dusty basic state are discussed below. It is not clear which conditions prevail during the decay of dust storms. The winds were moved to streak locations by transforming to Fourier–Legendre space and back. Since the zonal mean circulation (Fig. 13c) does not account for all the variation in streak direction, it is a natural
step to look at topographically forced stationary waves as a possible explanation.

a. Global results

The visible similarity between the total near-surface winds (Fig. 13a) and the streaks is supported by the calculation of a measure of alignment, $d$:

$$d = \frac{\sum V_i S_i}{\sqrt{\sum|V_i|^2} \sqrt{\sum|S_i|^2}} = \frac{1}{N} \sum \cos \theta_i,$$

where $V_i$ are the wind vectors, $S_i$ are the streak vectors, and $\theta_i$ is the angle between them. The summations are over the total number of vectors $N$ in a given region. In the last step we assume that all the vectors are unit magnitude since streaks do not provide explicit magnitude information. A value of $d = 1$ implies total alignment of wind and streak vectors, and $d = -1$ implies wind and streak vectors point in opposite directions.

The alignment measures calculated for $2 \times 2^8$ overlapping bins using the total and zonal mean winds are shown in Figs. 14a and 14b, respectively. The increase in alignment due to the stationary wave winds is shown in Fig. 14c and due to inclusion of radiative forcing in Fig. 14d. With the exception of a few localized areas (e.g., northeast of Tharsis, 135°W, 20°N), the waves provide a considerable improvement in wind-streak agreement near the areas of large topographical relief in the northern hemisphere. Much of the improvement is due to the radiative cooling or heating by Martian topography, supporting our conclusions about the nature and strength of the unique response of the Martian atmosphere to radiative forcing (section 3).

It is also instructive to look at the correlation of the near-surface winds and streaks. By examining the vector linear correlation coefficient

$$r = \frac{\sum (V_i - \bar{V})(S_i - \bar{S})}{\sqrt{\sum |V_i - \bar{V}|^2} \sqrt{\sum |S_i - \bar{S}|^2}}$$

as a function of latitude, we can test for a relationship between winds and streaks even if the zonal mean near-surface winds used in the linear model are not quite correct for the time of the streak formation. This is achieved by binning the vectors into narrow latitude bands (barred quantities denoting averages over the total number of vectors in a band). The correlation coeffi-

### Table 2. Regional wind-streak alignment measures (d) for low-dust simulations.

<table>
<thead>
<tr>
<th>Region</th>
<th>Zonal mean</th>
<th>No radiative forcing</th>
<th>Full forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td>Global</td>
<td>0.33</td>
<td>0.42</td>
<td>0.62</td>
</tr>
<tr>
<td>Acidalia/Chryse</td>
<td>-0.21</td>
<td>0.</td>
<td>0.66</td>
</tr>
<tr>
<td>Elysium</td>
<td>0.37</td>
<td>0.61</td>
<td>0.82</td>
</tr>
<tr>
<td>Tempe Terra</td>
<td>-0.39</td>
<td>-0.12</td>
<td>0.16</td>
</tr>
<tr>
<td>Terra Tyrhena</td>
<td>0.67</td>
<td>0.69</td>
<td>0.70</td>
</tr>
<tr>
<td>Tharsis/Olympus</td>
<td>0.11</td>
<td>0.11</td>
<td>0.33</td>
</tr>
</tbody>
</table>
coefficients for the normalized total winds and streaks as well as for the normalized zonal mean winds and streaks are shown in Fig. 15. North of the equator and south of 45°N, the streaks are clearly related to the linear stationary wave winds. There is also some correlation evident between 10°S and the equator. South of 10°S, however, the stationary wave winds do not exhibit any consistent relationship to the streaks. The large fluctuations of both the full wind and mean wind correlation coefficients south of 25°S and north of 45°N are most likely due to small numbers of streaks in those regions and their poor spatial coverage.

It is reassuring to note that the stationary wind-streak relationship is most evident in the northern hemisphere where the zonal mean topography is steep and its slope is opposite to the slope of the zonal mean potential temperature contours (Fig. 1). In such a region, even tall mountains can have more shallow slopes than the isentropes and satisfy the Cook and Held (1992) criterion to be well simulated by a linear model.

The vector correlation coefficients for total winds in the absence of radiative heating by topography are also shown in Fig. 15. Except for latitudes 15°–25°N and above 40°N, the correlation between streaks and winds is considerably diminished. There is a strong indication that the northern hemisphere streaks are closely related to the winds generated by mountain cooling and depression heating during northern hemisphere winter on Mars.
FIG. 17. Same as Fig. 16 but for the Tempe Terra region.

FIG. 18. Same as Fig. 16 but for the Tempe Terra region.
b. Regions

The alignment measures for several large-scale regions are displayed in Table 2. For comparison, the average dot products calculated using the zonal mean (basic-state) winds and winds from a "no radiative forcing" simulation are also shown. The regions are depicted in Fig. 4. Although zonal mean winds differ greatly from the streaks in some areas, the addition of stationary eddies always provides an improvement in regional alignment. Except for Terra Tyrrhena, the eddies due to radiative forcing dominate or contribute significantly to the wind-streak alignment in all regions.

1) Acidalia/Chryse Planitia (Fig. 16)

Acidalia and Chryse Planitia provide some of the best support for the connection between stationary waves and streaks and for the importance of radiative heating on Mars. While the zonal mean flow is opposite to the streaks in the 25°–40°N latitude band and deviates significantly over the whole region, the total flow is aligned with the streaks throughout the region.

In Chryse Planitia (west of 20°W, 20°–35°N), the wind pattern matches the flow due to a midlatitude depression dominated by radiative heating (opposite sign of Fig. 6c). In the northwest corner of the region, the winds are similar to flow on the western slope of a mechanically forced midlatitude depression (opposite sign of Fig. 6b). The differences between the near-surface winds and the streaks around 40°N and in the northeastern edge of Fig. 16 can be reduced or eliminated by shifting the latitude where zonal velocity changes direction northward or by increasing the strength of the eddies with respect to zonal mean flow.

2) Elysium Region (Fig. 17)

Elysium is situated near the latitude where zonal velocity changes direction (not a critical layer because of nonvanishing meridional velocity) and near the transition from the midlatitude to the equatorial regimes. This creates a rather complicated pattern. South of 20°N, the winds are similar to the flow over an equatorial mountain in easterly zonal mean surface wind (Fig. 6a). North of 25°N, the winds are like the flow over a midlatitude mountain in a westerly zonal mean wind (Fig. 7a). The flow in the valley to the west of 230°W and north of 20°N (Elysium Planitia) is dominated by the radiative heating response. This effect is similar (but of an opposite sign) to the flow depicted in Fig. 6c.

Some disagreement between the near-surface winds and the streaks can be attributed to the low resolution of our model. The streaks to the south of the Elysium Mons (213°W, 25°N) and to the east of Hecates Tholus (210°W, 33°N) seem to indicate winds strongly affected by these features. Since the features are not resolved by the linear model, similar near-surface winds do not exist in the solutions.

3) Tempe Terra (Fig. 18)

There is considerable agreement between the near-surface winds and streaks to the east and to the northwest of the highest point of Alba Patera (115°E, 35°N). Directly to the west of the mountain, however, there is sharp disagreement. While the regional total wind alignment measure is relatively low, it is a large improvement over the zonal mean wind alignment measure (Table 2). This suggests that we are using incorrect zonal mean winds for the time of streak formation or that eddies dominate the near-surface winds in this region.

Tempe Terra is the feature of Mars closest to the 40°N ideal mountain studied in section 3. Looking back at Fig. 6, we can see that the disagreement between the near-surface winds and the streaks can be greatly reduced by decreasing the importance of the radiative heating compared to the mechanical forcing. Right over the peak of Alba Patera, some streaks, while in close proximity to each other, point in different directions. Some of these streaks are consistent with radiative heating and some with the mechanical forcing. Different forcings may be dominant at this location at different times during the streak formation, or other processes such as nonlinearities or boundary layer flows may also be important. The slope near 120°W longitude is unusually large and may produce strong local or nonlinear effects.

It is interesting to note that the two adjacent regions (Tempe Terra and Chryse Planitia) seem to provide contradictory evidence of the importance of radiative heating in the forcing of stationary waves. One possible resolution of this contradiction comes from comparing the albedo of the two regions. Chryse Planitia is very much darker than Tempe Terra. It follows that the surface in the Chryse Planitia should be hotter and have a relatively greater radiative effect than the surface of Tempe Terra.

4) Terra Tyrrhena (Fig. 19)

In Terra Tyrrhena region, the zonal mean near-surface winds are in good agreement with the streaks, and the stationary eddies improve this agreement only slightly (Table 2). However, east of 300°W, inclusion of stationary waves does provide a noticeable improvement. This improvement increases north and south of the equator to the point where the streak pattern around Isidis Planitia (270°W, 15°N) is opposite of the zonal mean flow but agrees with the wave pattern. Similarly, some streak vectors in the southeastern portion of the region point toward the direction of the wave winds but opposite or perpendicular to the mean flow. The close proximity of streaks pointing in the directions of the zonal mean and eddy components of the wind might indicate a time variability of the relative importance of the two components.
Fig. 19. Same as Fig. 16 but for the Terra Tyrhenia region.

Fig. 20. Same as Fig. 16 but for the Tharsis and Olympus Mons region.
5) Tharsis region and Olympus Mons (Fig. 20)

There is agreement between near-surface winds and streaks on the southern slopes of Olympus Mons (135°W, 20°N). There is also agreement on the northeastern edge of Ascraeus Mons (100°W, 10°N), on the slopes of Arsia Mons (120°W, 10°S), near Sinai/Solis Plenum (95°W, 20°S), and around Lunae Plenum (60°W, 15°N). But to the north of Olympus Mons and to the northeast of Pavonis Mons (110°W, 0°), the total flow and the streaks point in the opposite directions. To the northwest of Ascraeus Mons and to the west of Arsia Mons, the model flow also significantly deviates from the streak direction. The total wind-streak alignment measure for the Tharsis region is low but is significantly higher than the zonal mean wind-streak alignment measure (Table 2).

Some of the wind-streak differences can be minimized by changing the ratio of the wave component to the zonal mean component of the model near-surface winds as well as changing the zonal mean wind direction. For example, decreasing the zonal mean wind compared to the wave amplitude in the 5°–15°S band will considerably improve the wind-streak agreement to the west of Arsia Mons without significantly changing the already good agreement to the east. This is further supported by variability of the streak pattern in this latitude band of the Terra Tyrrhena region discussed above. An increase of the zonal mean wind component with respect to the eddy component, along with a modest change in the zonal mean wind direction, can also diminish the wind-streak differences northeast of Olympus Mons.

The uncertainty in the zonal mean winds at any given time on Mars, the uncertainty in the exact time when the streaks are laid down, and the rough approximation of the dissipation can all contribute to an incorrect ratio of wave strength to zonal mean wind strength in the model.

The disagreement on the northeastern slope of Pavonis and on the western slope of Arsia Mons appears to be of a different nature. The winds do not agree with the streaks regardless of the ratio of the eddies to the zonal mean winds. However, the disagreements can be diminished by increasing the importance of radiative heating over the mechanical forcing (Figs. 7c, b). Furthermore, since the linear model does not resolve Ascraeus Mons from Pavonis Mons, the disagreement around these mountains should also diminish with use of a higher-resolution model.

It is surprising that there is any agreement between the linear model results and streaks near Olympus Mons, a clearly nonlinear feature by Earth standards. Since Olympus Mons is the tallest and one of the sharpest features on Mars, this agreement supports our use of the linear model to calculate the near-surface stationary wave winds.

c. Effects of dust

Greeley et al. (1993), using results from the Ames MGCM dusty northern winter simulation, calculated the average of the cosine of the angle between near-surface GCM winds and streaks bin averaged to 7.5° latitude by 9° longitude grid. They found that for the dusty northern winter MGCM flow, this measure was larger than 0.5 (approximated from their Fig. 9).

We investigate the effects of dust on the near-surface wind-streak agreement by using the zonal mean fields from the MGCM dusty northern winter simulation (Haberle et al. 1993, Fig. 6, run 90.02). The fields are transformed to the linear model grid and the temperatures on the two lowest layers are changed to increase the inversion. Haberle et al. (1993) found that their nondusty northern winter simulations underestimated the inversion seen in the Mariner 9 IRIS data (Santee and Crisp 1993). While they did not do a similar comparison with their dusty simulation, the dusty zonal mean fields they present have a much weaker inversion than the nondusty fields despite the lower surface temperature and stronger atmospheric heating. The near-surface wind-streak agreement improves noticeably when the zonal mean temperature inversion of the dusty basic state is enhanced.

The near-surface zonal mean winds from the dusty basic state as well as eddy and total winds resulting from using this basic state in the linear model are shown in Fig. 21. The zonal mean wind and total wind alignment measures for the regions discussed above are given in Table 3. There is a considerable improvement in the wind-streak alignment when compared to the nondusty winter basic-state results. Almost all of the improvement comes from a better match between the zonal mean winds and streaks. However, as demonstrated by Fig. 22, eddy winds still play an important role in the agreement. The improvement due to the inclusion of radiative heating is significant in all regions but Elysium and Terra Tyrrhena.

6. Summary and discussion

By exploring the near-surface, linear stationary wind patterns created by simple mountains in the Mars GCM zonal mean flow, we find important differences between Martian winds and the winds studied on Earth. The nature of the diabatic heating, the response to the heating, and the relative magnitudes of the mechanical forcing and heating are very different on the two planets. On Mars, diabatic heating is dominated by radiative forcing, and near-surface inversion causes mountains to cool and depressions to heat their surroundings over most of the planet. Short radiative timescales let the atmosphere respond to radiative forcing by perturbing the temperature and thereby changing the meridional pressure gradient. Strong meridional velocity near the equator and short radiative timescales throughout the planet’s atmosphere...
Fig. 21. Same as Fig. 13 but for a dusty basic state.
combine to make the responses to radiative and mechanical forcing of the same order in the tropics as well as some areas of the northern midlatitudes. In the region near 30°N, however, the weak zonal and meridional basic-state winds allow radiative forcing to dominate. On Earth, in contrast, mountains heat, depressions cool, and the atmosphere responds to heating through advection across the meridional temperature gradient (Cook and Held 1992). Winds forced by latent heating dominate in the tropics, and mechanically forced winds dominate in the midlatitudes (Ringle and Cook 1995).

Aloft, the linear stationary responses are also different on Mars and Earth. While the localized wave trains forced by zonal mechanical forcing are similar on the two planets, there are global, low wavenumber responses present that are unique to Mars. These responses arise with and without mechanical and radiative forcing and appear to be inherent to the atmospheric structure of Mars. Webster (1977), Hollingsworth and Barnes (1995), and Barnes (1996) noted similar global, low wavenumber disturbances in their simulations. Banfield et al. (1996) also observed wavenumber 1 and 2 modes in their analysis of the Viking IRTM temperature data.

There is strong evidence of stationary winds in the streak pattern of the northern hemisphere on Mars. When stationary eddies from the Martian linear model are added to the zonal mean winds, the large-scale wind-streak alignment improves by all measures. This improvement is partly due to the radiative forcing. There are only a few local regions where the addition of stationary winds lessens the wind-streak agreement and even fewer local regions where the addition of the eddies due to the radiative forcing does. Even near Olympus Mons, a nonlinear feature by Earth standards, linear eddies provide an improvement over the zonal mean winds. Using the dusty basic-state fields increases the wind-streak alignment even further. Most of the improvement is due to better agreement between the dusty basic-state winds and streaks, but stationary eddies and radiative forcing remain important.

Several steps can be taken to extend the presented work. Finer horizontal resolution models can improve the wind-streak agreement considerably by including topography not resolved by our model. A comparison of nonlinear or GCM stationary near-surface winds with the linear winds for realistic Martian topography can

| Table 3. Regional wind-streak alignment measures (d) for high-dust simulations. |
|---------------------------------|-----------------|-----------------|-----------------|
| Regions                        | Zonal mean      | No radiative    | Full forcing    |
| Global                         | 0.61            | 0.64            | 0.72            |
| Acidalia/Chryse                | 0.70            | 0.53            | 0.83            |
| Elysium                        | 0.82            | 0.90            | 0.91            |
| Tempe Terra                    | 0.28            | 0.42            | 0.53            |
| Terra Tyrrhena                 | 0.63            | 0.75            | 0.71            |
| Tharsis/Olympus                | 0.11            | 0.19            | 0.42            |
better explain the disagreements due to nonlinearity. Further refinement of the zonal mean fields and more data from Mars can significantly reduce the uncertainty associated with the model inputs.

Acknowledgments. The authors would like to thank Don Banfield, Mike Smith, Rebecca Stillwell, and Peter Thomas for helpful comments on the manuscript. We also would like to thank Jeffrey Barnes and an anonymous referee for helpful suggestions and comments. This work was supported by the NASA Planetary Atmospheres Program.

APPENDIX A

Dissipation on \( \sigma \) Surfaces and Radiative Forcing

a. Dissipation on \( \sigma \) surfaces

The effect of dissipation on \( \sigma \) surfaces in a linear \( \sigma \)-coordinate model can be demonstrated using a simple basic-state example. In the absence of the mean surface winds, the solution to the inviscid model linearized about a zonally symmetric, thermal wind balanced basic state was given by Held and Ting (1990):

\[
\frac{\partial}{\partial \log \sigma} (\log p)' = -\frac{\phi'_s}{RT} (\log p)',
\tag{A1}
\]

\[
\phi'_s = -RT (\log p)',
\tag{A2}
\]

\[
u'_\sigma = -(\log p)', \frac{\partial U}{\partial \log \sigma},
\tag{A3}
\]

\[
u'_\sigma = -(\log p)', \frac{\partial V}{\partial \log \sigma} = 0,
\tag{A4}
\]

\[
T'_\sigma = -(\log p)', \frac{\partial T}{\partial \log \sigma},
\tag{A5}
\]

\[
\dot{\sigma}' = -\sigma \frac{\partial (\log p)'}{\partial x}.
\tag{A6}
\]

This is equivalent to a vanishing disturbance in pressure coordinates (as expected in the absence of the mean surface winds) since the solutions in the two coordinate systems are related by

\[
u'_p = u'_\sigma + (\log p)', \frac{\partial U}{\partial \log \sigma},
\]

\[
T'_p = T'_\sigma + (\log p)', \frac{\partial T}{\partial \log \sigma}.
\]

However, the presence of dissipation on \( \sigma \) levels adds terms (neglecting the biharmonic diffusion) of the form

\[
-\frac{1}{\tau_h} (\log p)', \frac{\partial U}{\partial \log \sigma},
\tag{A7}
\]

\[
-\frac{1}{\tau_n} (\log p)', \frac{\partial T}{\partial \log \sigma}
\tag{A8}
\]

to the momentum and temperature equations. These terms force the solution away from (A1)–(A6), and...
therefore, from the vanishing disturbances expected in the pressure models linearized about the same basic state.

The physical significance of the temperature equation term can be ascertained by considering the radiative effect of topography on the atmosphere. This is done in the next section. The significance of the momentum equation term can be found in a similar manner.

b. Radiative forcing by topography

In an atmosphere where the radiative timescale is short compared to the dynamic timescale, a mountain can radiatively warm or cool the fluid in its vicinity. To demonstrate how this effect can be included in a linear, steady calculation, it is instructive to look at a simple mountain case in pressure coordinates.

Using the Newtonian cooling approximation, radiative forcing in a linear model is given by

\[
Q = \frac{c_p}{\tau_N} \left( \frac{T_e - T}{T_e - (\bar{T} + T')} - \frac{T_e - \bar{T}}{\tau_N} \right)
\]

where \(T_e\) is the thermal equilibrium temperature, \(\bar{T}\) is the temperature in the absence of the mountain (basic state), \(T'\) is the temperature due to a mountain (perturbation), and \(\tau_N\) is a radiative time constant. To a first approximation in the Martian atmosphere, the radiative equilibrium temperature at any given pressure level depends only on the distance of that level from the surface (Goody and Belton 1967; Gierasch and Goody 1968):

\[
T_e = T_e\left( \frac{p}{p_s} \right).
\]

Since the total surface pressure is given by

\[
p_s = \bar{p}_s + p'_s,
\]

where \(p'_s\) is the surface pressure due to the mountain, the radiative equilibrium temperature can be written as

\[
T_e\left( \frac{p}{p_s} \right) = T_e\left( \frac{p}{p_s} + p'_s \right) = T_e\left( \frac{p}{p_s} \right).
\]

Ignoring \(O(e^2)\) effects inside the brackets and remembering that in a flat topography basic state in radiative equilibrium

\[
\bar{T}(p) = T_e\left( \frac{p}{p_s} \right),
\]

we get

\[
T_e\left( \frac{p}{p_s} \right) = \bar{T} - \frac{p'_s}{p_s} \left[ \frac{\partial \bar{T}}{\partial \log \left( \frac{p}{p_s} \right)} \right]_{p_s \to \bar{p}_s},
\]

and (A9) reduces to

\[
Q = -\frac{c_p}{\tau_N \bar{p}_s} \left[ \frac{\partial \bar{T}}{\partial \log \left( \frac{p}{p_s} \right)} \right]_{p_s \to \bar{p}_s} - \frac{T'}{\tau_N}.
\]

This is the topographical thermal forcing described by
Webster (1977) with pressure used as a vertical coordinate rather than height plus the Newtonian cooling of the eddies. If we now assume no eddies and realize that in linear $\sigma$-coordinates $(\log p_i)' = \frac{p'}{p_i}$, we can rewrite the radiative forcing at each pressure level using $\sigma$-coordinates:

$$\frac{Q}{c_p} = -\frac{1}{\tau_N} (\log p_i)' \frac{\partial T}{\partial \log \sigma}. \tag{A17}$$

This is just the forcing due to dissipation of temperature on $\sigma$ levels given in (A8) and (5).

**APPENDIX B**

**Radiative Transfer Calculation of Newtonian Cooling Timescales**

The latitude-independent Newtonian cooling timescale profile is calculated by perturbing a 50-level, plane parallel, two-stream, radiative transfer model from its equilibrium temperature. The model includes absorption of the solar radiation by the near infrared (4.3 and 2.7 $\mu$m) bands of CO$_2$ as well as by dust, and absorption of thermal radiation by the 15 $\mu$m band of CO$_2$.

The absorption by the 4.3 and 2.7 $\mu$m bands is treated using a band model where the absorptance is approximated with the help of the exponential kernel and Curtis–Godson approximations (Cess and Ramanathan 1972). The absorption by the 15 $\mu$m band is calculated following Smith (1994). The discrete ordinates and correlated $k$ approximations are used to solve for the radiative flux numerically using a table of correlated $k$ values from NASA Goddard Space Flight Center. Dust is assumed well mixed and the absorption by dust is assumed to be gray.

The planetary constants and atmospheric parameters are set to Martian values. The cosine of the solar zenith angle is set to its average value over the globe and the solar solid angle is halved to account for diurnal variation. The surface temperature and dust opacity are varied until the calculated radiative equilibrium temperature closely resembles the latitudinally averaged MGCM northern hemisphere winter zonal mean temperature profile (Fig. B1). The parameters are listed in Tables 1 and B1. The Newtonian cooling times are shown in Fig. 3.

The near-surface timescales are fairly insensitive to the changes in equilibrium temperature profile and surface pressure. Changing the temperature profile from the average to the extreme southern latitude profile only changes the cooling times on the order of 30% near the surface. Changing from the average profile to the northern midlatitude profile changes the cooling times near the surface by a factor of 2. Increasing the surface pressure to 9 mb, leaves the near surface cooling timescales virtually unchanged.

**REFERENCES**


Hoskins, B. J., and D. J. Karoly, 1981: The steady linear response


