A Comparative Study of Daytime Thermally Induced Upslope Flow on Mars and Earth

Z. J. Ye,* M. Segal and R. A. Pielke

Department of Atmospheric Science, Colorado State University, Fort Collins, Colorado

(Manuscript received 11 November 1988, in final form 5 October 1989)

ABSTRACT

Several characteristics of thermally induced mesoscale upslope flow on Mars and its comparison with that on Earth were investigated using both analytical and numerical model approaches. The conclusions obtained from the analytical and the numerical evaluations for Mars are generally in agreement. The main conclusions are (i) the intensity of the Martian summer daytime upslope flow, with a moderate slope, reaches nearly 10 m s\(^{-1}\) and its depth is about 5 km; (ii) the longwave radiation flux divergence heating, within the lower boundary layer of Mars, has a nonnegligible contribution to the intensity of the upslope flows; and (iii) on Mars, the values of upslope wind speed are, typically, about 2.5 times larger than on Earth under similar conditions. The depth of the daytime upslope flow and the air temperature increase near the surface are 3 to 4 times larger. The vertical eddy exchange coefficient for heat and momentum is about 10 times larger on Mars.

1. Introduction

Relatively little concern has been given to an evaluation of mesoscale thermally induced upslope flow over Mars. The temperature of the Martian surface and the temperature of the lower atmosphere, as estimated from the 7, 9, 11, 15, and 20 \(\mu\)m channels that measure the thermal emission of atmospheric CO\(_2\), and the measurements from the Viking Landers (VL-1 and VL-2) (Kieffer et al. 1976; Nier et al. 1976; Nier and McElroy 1976; Hess et al. 1976, 1977; Martin et al. 1979; and Martin and Kieffer 1979), are known to have very large diurnal variations. The Martian surface is also known to exhibit major elevation differences (e.g., Sagan and Pollack 1968; Pettengill et al. 1973), which are expected to generate thermally forced slope flows and even influence the general circulation of the planet.

Gierasch and Sagan (1971) presented the impact of small horizontal scale topography on thermally induced wind on Mars using a scaling analysis. Their results suggest that the wind velocity is proportional to the temperature perturbation (i.e., the departure from the initial temperature due to daytime warming) and inversely related to both the horizontal and the vertical characteristic scales of the topographic relief. Blumskak et al. (1973) has presented a useful study of the Martian boundary layer (MBL) utilizing a one-dimensional (1-D) model. That study has provided initial knowledge about the oscillation amplitudes of the thermally induced upslope wind, temperature, and boundary layer thickness as affected by the thermal forcing over the Martian sloping terrain.

The wind data obtained at the surface of Mars by VL-1 and VL-2 were suggested by Hess et al. (1976, 1977) to indicate the existence of thermally forced daytime upslope flow. On the other hand, Leovy (1981) suggested that the observed diurnal wind variations at the VL-2 site are largely affected by the Martian atmospheric tides.

The lack of adequate observed data to evaluate the daytime thermally induced upslope flow on Mars suggests that theoretical methodologies, which have been proven successful on Earth, can be attempted for Mars. The purpose of the present study is to further explore the expected mesoscale daytime thermally forced upslope flow structure on Mars, using an analytic procedure and numerical model simulations. The evaluations were carried out for the summer daytime hours since they can be more conclusive compared to those with weaker nighttime thermal forcing. Using an analytical approach the upslope flow characteristics were investigated as a function of the atmospheric background thermal stability, \(\beta_0\), slope steepness, \(\alpha\), and thermal function, \(Q\) (defined by Eq. (15)). Numerical model simulations evaluating the Mars daytime upslope flow were also carried out. Unlike Mars, numerous mesoscale observational and modeling studies evaluating the daytime thermally induced upslope flows are documented for the Earth. A comparison of the relations for those flow over both planets should pro-

* Permanent affiliation: Institute of Atmospheric Physics, Academia Sinica, Beijing, China.

* Corresponding author address: Z. J. Ye, Institute of Atmospheric Physics, Academia Sinica, Beijing 100011 China.

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vide additional insight into these flows over Mars. Such an attempt is pursued in the present study. Results involved with the analytical methodological are presented in section 2. Evaluations of those relations utilizing a numerical model are given in section 3.

2. Analytical evaluations

a. The dependence of the thermally upslope flow features on $\theta_0, \alpha,$ and $Q$.

1) FORMULATION

Six assumptions were made in the present study with respect to the Martian atmosphere: (i) the ambient wind is assumed to be zero (unless stated otherwise) in order to isolate the thermal effect; (ii) the slope is homogeneous and long compared to the vertical scale of the upslope flow, $h$ (therefore, the horizontal variation of the induced upslope flow can be neglected); (iii) the temperature in the MBL is above the freezing point of CO2 $\sim 150$ K—the temperature in the Northern Hemisphere summer according to the VL-1 and VL-2 observations, is for most latitudes, at least 20°K above the CO2 condensation point (e.g., Kondratyev and Hunt 1982, p. 391). Therefore, in the present study, which focuses on the Martian northern hemisphere summer, the release of latent heat of sublimation is not considered; (iv) the atmosphere is cloudless and dust-free; (v) the depth of the upslope flow and the daytime convective boundary layer are closely coincidental [as suggested for the Earth by Johnson and O'Brien (1973); Ye et al. 1987]; and (vi) the MBL is assumed to be nearly adiabatic and the depth of the superadiabatic surface layer is considerably smaller than the MBL depth, $h$, so that

$$\frac{\partial \theta}{\partial z} \approx 0$$

(1)

can be applied to represent the profile of potential temperature, $\theta$, within the MBL during the daytime.

It is assumed that a quasi-steady state of the induced flow occurs following the daytime peak warming in the afternoon hours [a quasi-state flow can be interpreted from some of the hodographs provided in Leovy (1981)]. Under the assumptions listed above, a scaled solution for the upslope flow within the MBL can be obtained. The governing equations for the upslope flow, $u$ and $v$, are approximated as

$$\frac{\partial}{\partial z} K \frac{\partial u}{\partial z} + f u + \lambda \theta' \sin \alpha = 0$$

(2)

$$\frac{\partial}{\partial z} K \frac{\partial v}{\partial z} - f u = 0$$

(3)

with the boundary conditions

$$u = 0 \text{ and } v = 0 \text{ when } z = 0 \text{ and } z = h,$$

(4)

where $\theta'$ is the potential temperature perturbation from the background value $\theta_0$ at the same height. It is defined as

$$\theta' = \theta - \theta_0,$$

(5)

with

$$\lambda = g / \theta_0$$

the buoyancy parameter, where $g$ is the acceleration of gravity,

$$\alpha$$

the terrain slope angle,

$$K$$

the eddy exchange coefficient.

The form of $K$ is similar to that in Ye et al. (1987) and is expressed as

$$K = K_0 h (1 - \xi)^2 (a + b \xi),$$

(6)

where $\xi$ is the nondimensional length, $\xi = z / h$, $a$ and $b$ are constants (depending on $h_L / L$, where $h_L$ is the depth of the surface layer and $L$ is the Monin-Obukhov length), and

$$K_0 = 1.1 \kappa_0 u_*,$$

(7)

where $\kappa_0$ is von Kármán’s constant and $u_*$ is the friction velocity.

Physically, Eqs. (6) and (7) indicate that the turbulent intensity can be scaled by knowing the MBL depth, $h$, and $u_*$. From Eqs. (1) and (5) we have:

$$\frac{\partial \theta'}{\partial z} = -\beta_0,$$

(8)

where the background stability, $\beta_0 = \partial \theta_0 / \partial z$, is assumed constant with height. Integrating Eq. (8) vertically with the boundary conditions:

$$\theta' = \Delta \theta \text{ at } z = 0$$

$$\theta' = 0 \text{ at } z = h$$

results in

$$\theta' = \Delta \theta - \beta_0 h \xi,$$

(9)

where $\Delta \theta$ is the increase of $\theta$ at the slope surface during the daytime hours forced by solar radiative heating.

For the sake of simplicity, however, while keeping the physical meaning expressed by Eq. (6), a scaled constant $\hat{K}$ was approximated as

$$\hat{K} = C_0 h$$

(10)

where

$$C_0 = \int_0^1 K_0 (1 - \xi)^2 (a + b \xi) d \xi.$$

(10a)

In the present study, we assumed that $C_0 = 0.07$ m s\(^{-1}\), which is obtained by using $a = -0.05, b = 5.3$, and $u_* = 0.37$ m s\(^{-1}\). The values of $a$ and $b$ are obtained from Eq. (B1) in Ye et al. (1987) by using $h_L / L \approx -20$. The values of $h_L / L$ are based on the numerical model simulations reported in section 3. The value of $u_*$ reflects an average magnitude as obtained by the numerical model simulations (see Fig. 9).

Substituting $\hat{K}$, as given by Eq. (10), into Eqs. (2)
and (3) and multiplying Eq. (3) by the imaginary unit number, $i$, and then adding the results to Eq. (2) yields:

$$\frac{d^2W}{dx^2} - \frac{W}{C_0} + \frac{\lambda \sin \alpha}{C_0 h} \theta' = 0,$$

(11)

where $W = u + iv$.

Neglecting the Coriolis force (which can be appropriate as a first approximation in low latitudes) results in the following solution of Eq. (11):

$$u \approx \frac{\lambda Q_e \sin \alpha}{C_0} \left( \frac{2 + \xi^2}{3} - \xi \right) \xi.$$

(12)

In the general case, in which the Coriolis effect is considered, the scaled solution for $u$ is given by

$$u = \frac{2\lambda Q_e \sin \alpha}{(f \chi/2 C_0)} \left( (B_1 - B_4) \sin(B\xi) + (B_3 - B_2) \cos(B(2 - \xi)) \right)/p,$$

(13)

$$v = \frac{2\lambda Q_e \sin \alpha}{(f \chi/2 C_0)} \left( \xi - 1 + [(B_1 + B_4) \cos(B\xi) - (B_2 + B_3) \cos(B(2 - \xi))] \right)/p,$$

(14)

where

$$B = (f \chi/2 C_0)^{1/2},$$

$$B_1 = \exp(-B\xi),$$

$$B_2 = \exp(-B(2 + \xi)),$$

$$B_3 = \exp(-B(2 - \xi)),$$

$$B_4 = \exp(-B(4 - \xi)),$$

and

$$p = \exp(-4B) - 2 \exp(-2B) \cos(2B) + 1.$$

and the MBL diabatic heating, $Q_e$, is given by:

$$Q_e = \int_0^h \theta' dz = \int_0^r H_\theta dt.$$

(15)

Here

$$H_\theta = \frac{H_\theta}{\rho c_p}$$

(16)

with

$$H_\theta = R_h - R_0 + \Delta SWR - H_{se},$$

(17)

$$H_{se} = \rho c_p u_m \theta_a.$$

(18)

The relationship between $\Delta LWR + \Delta SWR$ and $H_{se}$ is adopted from Kieffer et al. (1976), which suggests as a first approximation for the daytime MBL:

$$\Delta LWR + \Delta SWR = -n(1 + C_1 \epsilon_d) H_{se},$$

(19)

where $n \approx 0.6$, $C_1 \approx 1.9$, and $\epsilon_d$ is the emissivity of dust [the values of $\epsilon_d$ change between 0 (dust free) and 1 (when the atmospheric optical depth is 0.32); see Kieffer et al. (1976)].

The $\Delta SWR$ and $\Delta LWR$ contribution to the heat balance of the MBL is significant during dust storm conditions, as can be seen from Eq. (19). However, in a nondusty atmosphere (i.e., $\epsilon_d = 0$), which is assumed in the present study, the importance of radiative heating is reduced somewhat as compared to that of the sensible heat flux [about $0.6 \cdot H_{se}$ based on Eq. (19)]. This result is also supported by the numerical model simulations presented in section 3 when the surface winds are strong enough.

Over Earth, the daytime value of $\Delta LWR + \Delta SWR$ can be neglected as compared to the value of $H_{se}$ (e.g., McNider and Pielke 1981). Therefore, Eq. (17) becomes:

$$H_Q \approx \begin{cases} -H_{se} & \text{(over Earth)} \\ -(1 + n)H_{se} & \text{(over Mars)} \end{cases}$$

(20)

Substituting Eq. (9) into Eq. (15) and integrating yields

$$Q_e = (1/2) \beta_0 h^2.$$

(21)

Combining Eq. (9) (with $\xi = 1$) and Eq. (21) results in:

$$Q_e = (\Delta \theta)^2/(2 \beta_0).$$

(22)

Reorganizing Eqs. (21) and (22) leads to

$$h = [2Q_e/\beta_0]^{1/2},$$

(23)

$$\Delta \theta = (2\beta_0 Q_e)^{1/2}.$$  

(24)

2) Evaluations

Equation (12) indicates that for the special situation in which $f$ can be neglected (i.e., near equator latitudes): (i) the intensity of the upslope flow, $u$, is proportional to $Q_e$ and $\sin \alpha$ and is inversely proportional to $C_0$; and (ii) the profile of $u$ is independent of $\beta_0$ when $\xi$ is chosen as the vertical coordinate.

In the general case, in which the Coriolis force is included, the more complicated solution given by Eqs. (13) and (14) as dependent on $Q_e$, $\beta_0$, and $\alpha$, is shown by Fig. 1a. Curve 1 in Fig. 1a presents the values of $U_{max} = (u_{max}^2 + v_{max}^2)^{1/2}$ as a function of $Q_e$ for $\alpha = 0.6$°, $\beta_0 = 4$ K km$^{-1}$, and latitude $\varphi = 23.3$°.

The results indicate that the intensity of $U_{max}$ increases linearly with $Q_e$. The impact of variations in $\beta_0$ on $U_{max}$ with $Q_e = 4.0 \times 10^4$ (m K)—the other
The dependence of $U_{max}$ is almost unaffected by changing $\beta_0$. In addition, the relation between $U_{max}$ and slope steepness, $\alpha$, is shown by curve 3 with $Q_s = 4.0 \times 10^4$ (m K) — with the other parameter values the same as in curve 1.

Due to the lack of appropriate observations, the depth of the daytime MBL is uncertain. Equation (23) provides a quantitative estimation of the relation between $Q_s$, $\beta_0$, and $h$, indicating that $h$ is proportional to $Q_s^{1/2}$ and inversely proportional to $\beta_0^{1/2}$. Equation (24) shows that $\Delta \theta$ is proportional to both $\beta_0^{1/2}$ and $Q_s^{1/2}$. For example, using typical values of $\beta_0 = 4 \times 10^{-3}$, K m$^{-1}$, and $Q_s = 4.0 \times 10^4$ m K results in $h = 4.5$ km.

The maximum value of $u$ as derived from Eq. (12) is

$$u_{max} \approx 0.13 \frac{\lambda Q_s \sin \alpha}{C_0}. \quad (25)$$

Curve 4 in Fig. 1a is computed, based on Eq. (25), using the same values of $\alpha$, $Q_s$, and $\beta_0$, as in curve 1. Comparing profiles 1 and 4 indicates that the values of $u_{max}$ and the dependence of $u_{max}$ on $Q_s$, in both cases, are very similar. Using Eqs. (13) and (14), Fig. 1b shows the impact of a change in latitude, $\varphi$ (i.e., the Coriolis force), on the values of $u_{max}$, $v_{max}$, and $U_{max}$. It indicates that $U_{max}(\varphi = 10^\circ)/U_{max}(\varphi = 90^\circ) \approx 1.1$, where $U_{max}(\varphi = 10^\circ)/U_{max}(\varphi = 90^\circ) \approx 0.76$, suggesting that as a first approximation the impact of the Coriolis force on $U$ for the quasi-steady solution for the upslope flow is small. The dependence of the intensity of the upslope flow on $\alpha$, $Q_s$, and $\beta_0$ (illustrated in Fig. 1a) was computed by Eqs. (13) and (14) and indicated the same results as computed by Eq. (25) [which was derived from Eq. (12)], where the Coriolis force is not considered. Therefore, it is suggested that the scaling, presented in this study as Eq. (12), is a reasonable analytical approximation for the upslope flow intensity.
suggest that: 1) $E$ increases about 5.3 times when doubling $Q_s$ (i.e., $E$ is nearly proportional to $Q_s^{5/2}$; 2) the value of $E$ decreases about 1.3 times when doubling $\beta_0$ while keeping $Q_s$ and $\alpha$ constant, and 3) $E$ increases almost 4 times when doubling $\alpha$. These results agree with those suggested by Eq. (29) (i.e., while neglecting the Coriolis force). The difference of the computed $E$ values, based on Eqs. (13), (14), and (27) (which include the Coriolis force), from the computed $E$ values, based on Eq. (29) (i.e., while neglecting the Coriolis force), is shown to be small (see inset of Fig. 3).

b. Comparison of scaled features of upslope flow on Mars and on Earth based on an analytical solution

Our observational documentation concerning the intensity of the upslope flow on Mars is extremely poor, while features of the thermally induced upslope flow on Earth are obviously more available. Therefore, an attempt is made to extrapolate daytime thermally induced upslope flow on Earth in order to delineate ther-
nally induced upslope flow characteristics on Mars. This is done by assuming similarities in certain scaled parameters. The presented comparisons are related to temperature perturbation, upslope flow intensity, the depth of the MBL, kinetic energy, and turbulent characteristics (turbulent friction velocity and eddy exchange diffusivity). The comparisons are mostly based on Eq. (12) (i.e., while neglecting the Coriolis force). As implied by our previous computations, it only results in small deviations in the quasi-steady, while other scaled parameters are nearly unaffected. This provides a simple and physically sensible expression for the relation between the variables on Earth and Mars. In this subsection the subscripts m and e represent the values on Mars (m) and Earth (e), respectively; constants given in Table 1 are used. From Eqs. (12), (23), (24), and (29) the following ratios exist:

\[ \frac{u_m}{u_e} = \left( \frac{\lambda Q_m \sin \alpha_m}{\lambda Q_e \sin \alpha_e} \right) \frac{C_{0e}}{C_{0m}} \]  

\[ \frac{h_m}{h_e} = \left( \frac{Q_{am}}{Q_{ae}} \right)^{1/2} \frac{\beta_{0e}}{\beta_{0m}} \]  

\[ \Delta \theta_m = \left( \frac{Q_{am}}{Q_{ae}} \right)^{1/2} \frac{\beta_{0e}}{\beta_{0m}} \]  

and

\[ \frac{E_m}{E_e} = \left[ \frac{Q_{am}}{Q_{ae}} \right]^{5/2} \frac{\sin \alpha_m \lambda_m}{\sin \alpha_e \lambda_e} \left( \frac{C_{0e}}{C_{0m}} \right)^2 \left( \frac{\beta_{0e}}{\beta_{0m}} \right)^{1/2} . \]  

Using Eqs. (15) and (16), the following relation between Mars and Earth can be derived:

\[ \frac{(\rho_0 C_p Q_{m})_m}{(\rho_0 C_p Q_{e})_e} = \left( \frac{T_r S_0}{T_r S_0} \right)_e , \]  

where \( T_r \) is the conversion efficiency of solar energy \( S_0 \) into \( H_Q \):

\[ T_r = \frac{H_Q}{S_0} . \]  

Rearranging Eq. (34) results in

\[ \frac{Q_{am}}{Q_{ae}} = R_a \frac{S_{0m}}{S_{0e}} \left( \frac{\rho_0 C_p)_e}{\rho_0 C_p)_m} \right) \]  

where

\[ R_a = \frac{T_r_m}{T_r_e} . \]  

Values of \( \rho_0 \) and \( \lambda \) on Mars and on Earth are calculated, based on the ideal gas law and the definition of \( \lambda \), using the values of \( P_0, R, \) and \( \theta_0 \) listed in Table 1. Using the values provided in Table 1, the following relations exist:

\[ S_{0m} \approx 0.73 S_{0e} \]  

\[ (\rho_0 C_p)_m \approx 9.4 \times 10^{-3} (\rho_0 C_p)_e . \]  

Substituting Eqs. (38) and (39) into Eq. (36) results in

\[ Q_{am} \approx 77.7 R_a Q_{ae} = A_Q Q_{ae} . \]  

Using Eq. (40), Eqs. (31) and (32) can be rewritten as

\[ h_m \approx 8.8 \left[ \frac{\beta_{0e}}{\beta_{0m}} \frac{R_a}{\rho_{0e}} \right]^{1/2} \cdot h_e \approx A_h h_e \]  

\[ \Delta \theta_m \approx 8.8 \left[ \frac{\beta_{0e}}{\beta_{0m}} \frac{R_a}{\rho_{0e}} \right]^{1/2} \cdot \Delta \theta_e \approx A_{\theta} \Delta \theta_e . \]  

Combining Eqs. (20), (35), (37), and (38) yields

\[ H_{0m} \approx \frac{0.73 R_a^2 H_{0e}}{1 + n} \]  

\[ H_{0m} = A_{H} H_{0e} . \]  

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<th>Name</th>
<th>Symbol</th>
<th>Unit</th>
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<th>Earth</th>
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or
\[
\frac{u_{a*m}}{u_{a*e}} \approx 77.7 R_a \frac{\theta_{a*m}}{\theta_{a*e}} \frac{1}{1 + n} \quad (43a)
\]

Assuming that the ratio \(u_{a*m}/u_{a*e}\) is proportional to the ratio \(\theta_{a*m}/\theta_{a*e}\), i.e., \(\frac{\theta_{a*m}}{\theta_{a*e}} = \frac{b}{\theta_{a*m}/u_{a*m}}\) and using Eqs. (7), (10a), and (43a) results in
\[
\frac{u_{a*m}}{u_{a*e}} = \frac{C_{a*m}}{C_{a*e}} = 8.8 \left[ \frac{R_a}{b(1 + n)} \right]^{1/2} = A_{u*a}. \quad (44)
\]

From Eqs. (10), (41), and (44) we have
\[
K_m = 77.7 R_a \left[ \frac{\theta_{a*m}}{\theta_{a*e}} \frac{1}{b(1 + n)} \right]^{1/2} K_e = A_{K*K_e} \quad (45)
\]

Substituting Eqs. (40) and (44) into Eqs. (30) and (33) with the value of \(\lambda\) listed in Table 1 and assuming \(\alpha_m = \alpha_e\), results in
\[
\begin{align*}
A_{u*m} &\approx 4.8 \left( \frac{b}{1 + n} \right)^{1/2} R_a^{1/2} \cdot u_e = A_{u*K_e} \quad (46) \\
E_{m} &\approx 200(1 + n) \frac{b}{3} \left[ \frac{\theta_{a*m}}{\theta_{a*e}} \right]^{1/2} \cdot E_e = A_{E*K_e} \quad (47)
\end{align*}
\]

Equations (40) to (47) indicate that the values of the ratios: \(AQ, A_h, A_A, A_{AE}, A_{a*K}, A_{u*K}, A_{K*K_e}, \) and \(A_{E*K_e}\) are dependent on \(R_a\). The dependence of \(A_{u*K}, A_{Q}, A_{h}, \) and \(A_{H_1}\) on \(R_a\) is relatively small (proportional to \(R_a^{1/2}\)); the dependence of \(A_{K*K_e}, A_{Q}, \) and \(A_{H_1}\) on \(R_a\) is moderate (proportional to \(R_a\)), and the dependence of \(A_{E*K_e}\) on \(R_a\) is proportional to \(R_a^{1/2}\). Only the values of \(A_{Q}, A_{h}, A_{AE}, \) and \(A_{K}\) are dependent on \(\beta_0\) (proportional to \(\beta_0^{1/2}\)). Figure 4, computed from Eqs. (40)–(47), shows that while \(R_a\) is changed from 0.04 to 0.28 with \(\beta_0 = \beta_{0*e}\) and \(b = 1\), the corresponding values of \(A_{Q} \) and \(A_{H_1}\) range from about 1 to 7; and for \(A_{Q}\) and \(A_{H_1}\) range from 2 to 10 to 13 to 18. Figure 4 also illustrates the impact of the variations in \(R_a\), which were stated above, on \(A_{u}\) and \(A_{E}\) (indicated by dashed lines) while being computed from Eqs. (13), (14), (27), (44), and (48) which include the Coriolis force effect. The computed values of \(A_{u}\) and \(A_{E}\) are similar to those computed while neglecting the Coriolis force.

Typical values for the various ratios, presented in Fig. 4, are estimated in the following. Sutton et al. (1978) evaluated that near local noon derived \(H_{0}\) values were typically around 18 W m\(^{-2}\) during the first 45 days following the VL-1 landing (Martian summer). Similar values for \(H_{0}\) are also quite typical in the model simulations presented in section 3 (see Fig. 11). Therefore, using Eqs. (20) and (35) and the value of \(S_{0}\), listed in Table 1, results in \(T_{m} \approx 0.048\). On Earth, \(T_{m}\) is generally considered to be equal to 0.30 (Leovy 1979), which results in \(R_{a} \approx 0.16\). Substituting the values \(n = 0.6\) and \(R_{a} = 0.16\) into Eq. (36) and Eqs. (40)–(47) yields the following typical relations:

\[
Q_{m} = 12.4 Q_{s}\quad (48)
\]

\[
h_{m} = 3.5 \left[ \frac{\beta_{0}/\beta_{0,*}}{\beta_{0}/\beta_{0,*}} \right]^{1/2} \quad h_{e}\quad (49)
\]

\[
\Delta \theta_{m} = 3.5 \left[ \frac{\beta_{0}/\beta_{0,*}}{\beta_{0}/\beta_{0,*}} \right]^{1/2} \Delta \theta_{e}\quad (50)
\]

\[
H_{0,1m} = 0.073 H_{0,1e}\quad (51)
\]

\[
u_{0,*} = 2.8 u_{0,*}\quad (52)
\]

\[
K_{m} = 9.6 \left[ \frac{\beta_{0}/\beta_{0,*}}{\beta_{0}/\beta_{0,*}} \right]^{1/2} K_{e}\quad (53)
\]

\[
u_{m} = 2.4 u_{e}\quad (54)
\]

and

\[
E_{m} = 0.26 \left[ \frac{\beta_{0}/\beta_{0,*}}{\beta_{0}/\beta_{0,*}} \right]^{1/2} E_{e}\quad \text{or}\quad (55)
\]

\[
E_{m} = 20.5 \left[ \frac{\beta_{0}/\beta_{0,*}}{\beta_{0}/\beta_{0,*}} \right]^{1/2} E_{e}.
\]

Equation (48) indicates that the value of \(Q_{s}\) on Mars is about one order of magnitude larger than that over Earth. Equations (49) and (50) suggest that the relations between \(h_{m}\) and \(h_{e}\), (\(A_{h}\)), and between \(\Delta \theta_{m}\) and \(\Delta \theta_{e}\), (\(A_{\theta}\)), depend on the value of \(\beta_{0}/\beta_{0,*}\). For the same values of \(\beta_{0}\) on both planets, \(h_{m}\) and \(\Delta \theta_{m}\) are approximately 3.5 times as large as \(h_{e}\) and \(\Delta \theta_{e}\), respectively. For example, representative values of \(h_{m}\) and \(\Delta \theta_{m}\) are about 1.5 km and 10–15°K, respectively. Thus, the corresponding values of \(h_{m}\) and \(\Delta \theta_{m}\) are about 5 km

Fig. 4. The dependence of the ratios: \(A_{h}, A_{A}, A_{a*K}, A_{Q}, A_{E}, A_{K}, A_{H_1}\), on \(R_a\), based on Eqs. (40) to (47) (solid lines) and with \(\beta_0 = \beta_{0,*}\). In the inset: the relation between \(U_{a*m}\) on Mars (m) and \(U_{a*e}\) on Earth (e) as derived from Eqs. (13), (14), (48), and (52) for the values of \(a * Q_{s}\) indicated on the abscissa of Fig. 1a and with \(\varphi = 23.3^\circ\).
and 40°K estimated from Eqs. (49) and (50). Flasar and Goody (1976), Sutton et al. (1978), and Leovy (1979) infer that the characteristic maximum depth of the Martian diurnal convective boundary layer is likely to be 5 km based on the VL-1 entry temperature profile and the orbiter imaging observations of convective clouds. Burk (1976), in a two-dimensional (2-D) model simulation of the thermally induced flow between the polar cap frozen CO₂ and adjacent bare soil area, found that the depth of the local wind reaches 3–4 km. However, Gierasch and Goody (1968) simulated a value of $h_m = 10$ km during the Martian afternoon using their 1-D radiative-convective model. The observation data of VL-1 indicates that during the summer the value of $\Delta \theta$ near the surface (at 1.6 m above the surface) is about 235 K at 1600 LST (Hess et al. 1976). According to Burk’s (1976) numerical model results the amplitude of the near surface air temperature perturbation is $\approx 40$ K. These facts indicate that the estimates using Eqs. (49) and (50) are reasonable. Sutton et al. (1978), using the wind speed, and ambient and surface temperatures from the VL-1 and VL-2 observations, computed $u_{*m}$ values ranging from 0.4 m s⁻¹ to 0.6 m s⁻¹ in the late morning near midsummer. Flasar and Goody (1976) and Kondratyev and Hunt (1982) suggest that the values of $u_{*m}$ range from 0.45 to 1.80 m s⁻¹. The values of $u_{*m}$, under unstable surface layer stability on Earth based on Eq. (52), are suggested to be in the range of 0.15 to 0.60 m s⁻¹, as typically observed. Equation (53) suggests that the eddy diffusion exchange coefficient on Mars is nearly one order of magnitude larger than that on Earth during the daytime hours.

Equations (54) and (55) suggest that the intensity of upwelling wind on Mars is about 2.4 times as large as that on Earth. The related layer-integrated density normalized kinetic energy, $E$, on Mars is about 20 times as large as that on Earth; however, the layer-integrated kinetic energy on Mars, $E_{m}^*$, is less than that on Earth (about 26 percent of that on Earth when $\beta_0 = \beta_0^*$ is assumed). The values of $U_{max}$ on Mars and on Earth, computed from Eqs. (13), (14), (44), and (48), which include the Coriolis force, are presented in the inset of Fig. 4 suggesting $\frac{U_{max}^*}{U_{max}} \approx 2.34(\frac{U_{max}}{U_{max}^*})$ for different values of $\alpha$, $Q_0$, and $\varphi = 23.3^\circ$, which is in agreement with Eq. (54). The ratio between $(U_{max})_m$ and $(U_{max})_e$ is nearly unchanged while altering the latitude as shown in Fig. 1b.

3. Numerical model simulations

In order to consider further refinements of the analytical evaluations, a set of illustrative numerical simulations (see Table 2 for their description) were performed. These simulations enable an evaluation of some of the analytical results, as well as providing an insight into some additional features of the upwelling flow on Mars. Apparently, no refined 2-D mesoscale numerical model simulations of the daytime thermally induced upwelling flow on Mars had been carried out. Therefore, the simulations presented in this study provide an initial step in this direction. In 2-D simulations, it is possible to account for nonideal terrain features while considering background flows. In addition, the impact of changes in terrain on the surface fluxes, as well as on the radiation flux divergence within the lower atmosphere, can be considered.

A 2-D hydrostatic primitive equations model, whose formulation is given in detail in Pielke (1974), Mahler and Pielke (1977), Mahler and Pielke (1978), and McNider and Pielke (1981) was used; therefore, its formulation is not repeated in the present paper. This model was successfully validated for a considerable number of cases of daytime thermally induced upwelling

<table>
<thead>
<tr>
<th>Case</th>
<th>Plateau height (m) (slope steepness) (deg)</th>
<th>$u_x$ (m s⁻¹)</th>
<th>$u_y$ (m s⁻¹)</th>
<th>Atmospheric longwave flux divergence and solar radiation absorption</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>800 (0.23)</td>
<td>0</td>
<td>0.5</td>
<td>included</td>
</tr>
<tr>
<td>M2</td>
<td>2000 (0.57)</td>
<td>0</td>
<td>0.5</td>
<td>included</td>
</tr>
<tr>
<td>M3</td>
<td>800 (0.23)</td>
<td>3</td>
<td>0</td>
<td>included</td>
</tr>
<tr>
<td>M4</td>
<td>2000 (0.57)</td>
<td>3</td>
<td>0</td>
<td>included</td>
</tr>
<tr>
<td>M5</td>
<td>800 (0.57)</td>
<td>−3</td>
<td>0</td>
<td>included</td>
</tr>
<tr>
<td>M6</td>
<td>2000 (0.23)</td>
<td>−3</td>
<td>0</td>
<td>included</td>
</tr>
<tr>
<td>M7</td>
<td>800 (0.23)</td>
<td>0</td>
<td>0.5</td>
<td>eliminated</td>
</tr>
<tr>
<td>M8</td>
<td>2000 (0.57)</td>
<td>0</td>
<td>0.5</td>
<td>eliminated</td>
</tr>
</tbody>
</table>
flows over Earth (e.g., Segal et al. 1982; Abbs and Pielke 1986; Steyn and McKendry 1987, among others). Worth noting is that in the surface layer similarity functional relations, the related empirical constants are assumed to be the same on Earth and Mars. It is uncertain how adequate this assumption is (which is also commonly adopted in other Martian studies). The laminar near-surface layer was considered following the parameterization in Mahler and Pielke (1977).

Obviously, the various original model input parameters and constants had to be changed to those appropriate for Mars (see Table 1). The longwave and shortwave radiation schemes were rewritten to consider the Martian atmosphere radiative processes. The main gaseous component of the Martian atmosphere, CO₂, exerts a dominant influence on radiative heating. The water vapor contribution to the total solar and longwave radiative heating rate in the Martian atmosphere is very small and is neglected in our numerical model simulations. The transmission function, which parameterizes the attenuation of the solar flux by the absorption of solar radiation by CO₂ and the atmospheric heating or cooling due to the flux divergence of longwave radiation caused by CO₂, is calculated using the formulation given in Pollack et al. (1981). It is briefly described in the following.

The solar radiative heating rate, due to absorption of solar radiation by CO₂, at any given model level N, is given by

$$\frac{dT(N)}{dt} = \frac{1}{\rho C_p} \frac{[S(N) - S(N+1)]}{[Z(N + 1) - Z(N)]},$$

where $S(N)$ represents the fractional amount of solar radiation absorbed by CO₂ above level N which is at height $Z(N)$ above ground. Denoting the pressure and temperature at level $N$ by P(N) and T(N), respectively, then $S(N)$ is approximated as

$$S(N) = 0.005 \dot{P}^{1/2}(N),$$

with

$$\dot{P}(N) = P(N) \left[ 1 + 0.405 \left( \frac{T(N) - 200}{200} \right) \right] \times (2 \cos \mu)^{-0.606},$$

where $\mu$ is the solar zenith angle.

The dependence of $S(N)$ on $P(N)$ given by Eq. (57) was derived based on Fig. A1 in Pollack et al. (1981).

The longwave radiative heating rate at each layer is computed as:

$$\frac{dT(N)}{dt} = \frac{1}{\rho C_p} R_L^{i}(N + 1) - R_L^{i}(N) + R_L^{i}(N) - R_L^{i}(N + 1) / [Z(N + 1) - Z(N)],$$

with

$$R_L^{i}(N) \approx 2\pi \left[ B_{(T_G)} \frac{\Delta \nu}{2} - 0.488 \sum_{j=1/2}^{N-1/2} E(j, N) \times \left[ B\left( j + \frac{1}{2} \right) - B\left( j - \frac{1}{2} \right) \right] \right],$$

$$R_L^{i}(N) \approx 0.488 \cdot 2\pi \left[ B_{(TOP)} E(TOP, N) - \sum_{j=N^{1/2}}^{TOP-1/2} E(j, N) \left[ B\left( j + \frac{1}{2} \right) - B\left( j - \frac{1}{2} \right) \right] \right],$$

where TOP is the top level of the model and $E(j, N)$ is the equivalent width of a 15 $\mu$m band of CO₂ approximated as:

$$E(j, N) = 0.44 \left( \frac{T}{T_0} \right)^{0.879} \ln \left[ 1 + 0.15 \left( \frac{T}{T_0} \right)^{0.256} \right] \times W(j, N)^{0.566} \tilde{F}^{0.323},$$

and

$B$ is the Planck function

$T_0$ is set equal to the mean atmospheric temperature over Mars (≈200°C)

$T_G$ is the ground temperature

$W(j, N)$ is the CO₂ path length between levels j and N

$\Delta \nu$ is the effective CO₂ band width

$\tilde{P}$ and $\tilde{T}$ are the height-weighted mean pressure and temperature between levels N and j, respectively, and given by:

$$\left( \tilde{P}, \tilde{T} \right) = \sum_{i=j}^{N} \left( \tilde{P}, \tilde{T} \right) \left[ Z(i + 1) - Z(i) \right] / \left[ Z(j) - Z(N) \right].$$

The simulations in the present study were carried out using a 2-D model version that consists of 36 levels ranging from near the surface to 28 km (1.6, 5, 15, 50, 100, 150, 200, 250, 300, 400, 500, 700, 900, 1100, 1300, 1600, 2000, 2500, 3000, 3500, 4000, 5000, 6000, 7000, 8000, 9000, 10 000, 12 000, 14 000, 16 000, 18 000, 20 000, 22 000, 24 000, 26 000, and 28 000 m above the surface). The simulation domain extended horizontally for 600 km with a terrain configuration of plain–slope–plateau (see Fig. 5) and was resolved with a horizontal grid interval of 10 km. The background stability, $\beta_0$, was prescribed to be 2°K km⁻¹ in the lower 9 km and 4°K km⁻¹ above this level (similar to that reported in Nier et al. 1976; Gadian 1978). The model simulations were carried out for 23°N and for an aero-centric longitude of the sun $L_s = 110^\circ$. It is worth noting that around this period (from $L_s \approx 54^\circ$ to $L_s \approx 126^\circ$), the solar radiation incident at the top of the Martian atmosphere of the Northern Hemisphere
in latitudes 20°–60°N is nearly the same (Levine et al. 1977). The initial surface pressure along the lower plains of the simulated domain is 7 mb, which is about the observed surface pressure at the VL-1 site during this period of the Martian year (Hess et al. 1980). The thermal inertia of Mars, following Palluconi and Kieffer (1981), is $10^{-3}$ to $15 \times 10^{-3}$ cal cm$^{-2}$ s$^{-1/2}$ K$^{-1}$; an average value of $7.18 \times 10^{-3}$ cal cm$^{-2}$ s$^{-1/2}$ K$^{-1}$ was used in the present simulations. Following the analysis of Sutton et al. (1978) of the Mars surface characteristics at the Viking Landers’ sites we adopted $z_0 = 0.5$ cm for the roughness height. The Martian day was divided into 24 hours denoted as MST (Martian Sun Time) where the model simulations commenced at 0600 MST (around sunrise) with surface temperatures of 208 K. Several illustrative cases were considered (as described in Table 2). Simulated flow and potential temperature fields as well as additional pertinent simulated fields are presented in the following.

a. Negligible geostrophic flow

The purpose of the simulations with the negligible geostrophic flow is to enable isolation of the thermally induced upwarp circulation, while also permitting a comparison with the analytical results previously presented. Two simulations were carried out to illustrate the impact of the slope steepness on the daytime upwarp flow component and on the development of the daytime MBL (cases M1 and M2). Results are presented for 1400 MST, an hour at which these fields are around their daytime peak development. For case M1 the upwarp component reached a value of 4.8 m s$^{-1}$ as compared to 0.4 m s$^{-1}$ for case M2 (Fig. 5a, d). These values are about 2.5–2.7 times as large as the corresponding values obtained for Earth under similar simulations (see Ye et al. 1987, Table 7). This ratio is also in agreement with the analytical scaling [see Eq. (54)]. As will be discussed in section 3e, the magnitude of the surface sensible heat flux on Mars is sensitive to the intensity of the surface wind speed. Therefore, increased values of the sensible heat flux are computed along the slope where the wind speed is relatively strong as compared to the lower plain and the elevated plateau. Consequently, to some degree a “sea breeze” like circulation was involved, supporting the upwarp flow in the lower plain, while opposing it on the elevated plateau. The simulated $v$ component (i.e., cross-slope component, see Fig. 5b, e) is quite small as compared to the $u$ component.

The potential temperature distribution within the $x$–$z$ cross section at 1400 MST is shown in Fig. 5c, f. The larger surface sensible heat flux along the steeper slope (case M2) (see Fig. 11) forces a somewhat deeper MBL as compared to the shallow slope (case M1). The depth of the MBL, as inferred from the potential temperature vertical distributions, is about 4.5 and 5.5 km in the M1 and M2 cases, respectively. Assuming a typical boundary layer depth of about 1.5 km on Earth (e.g., Ye et al. 1987), the depth of the slightly unstable MBL is larger by about a factor of 3 to 3.7 as compared to that on the Earth. The depth of the MBL, as inferred from the potential temperature fields and the depth of the upwarp flow layers is nearly equal, as assumed in
the analytical formulation presented in section 2. Also worth noting is that the simulated thermal stratification within the MBL is slightly unstable (unlike the closely neutral daytime boundary layer on Earth). Sensitivity simulations in which the vertical exchange coefficients were increased by a factor of 2 resulted in temperature profiles that were nearly adiabatic (not shown). This feature of the impact of change in thermal stability on the MBL, with increased turbulence, is similar to that reported in Blumsack et al. (1973).

b. Simulations with geostrophic flows

Coupling with large scale background flows should be considered in most situations involved with the daytime upslope flows. Cases M3–M6 are illustrative situations for both slopes that were used in cases M1

Fig. 6. Numerical model simulated x–z fields for cases M3 and M4 at 1400 MST. (a) $u_3$, with $u_3 = 3$ m s$^{-1}$ (M3), (b) $v_3$, with $u_3 = 3$ m s$^{-1}$ (M3), (c) $\theta$, with $u_3 = 3$ m s$^{-1}$ (M3), (d) $u_4$, with $u_4 = 3$ m s$^{-1}$ (M4), (e) $v_4$, with $u_4 = 3$ m s$^{-1}$ (M4), and (f) $\theta$, with $u_4 = 3$ m s$^{-1}$ (M4).

Fig. 7. The same as Fig. 6 except for cases M5 and M6.
and M2, where background flows, supportive or opposing the upslope flows, are introduced (Figs. 6 and 7). The magnitude of the geostrophic flow was selected based on Leovy's (1981) analysis for the surface wind at the VL-2 site. In general, this analysis implies that the geostrophic flow in the simulated period of the Martian year is typically not exceeding several meters per second. With a supportive geostrophic flow, $u_g$ of 3 m s$^{-1}$, an enhancement of the upslope flow component is simulated (cases M3 and M4; Figs. 6a, d). The enhancement is nearly linear in both cases. In the shallow slope case, case M3, the return circulation aloft was eliminated due to the existence of opposing geostrophic flow (Fig. 6a). A noticeable $v$ component developed in the steep slope case (Fig. 6e). The MBL depths in these two cases (see Fig. 6c, f) are comparable to those simulated in cases M1 and M2. When a downslope geostrophic flow $u_g$ of $-3$ m s$^{-1}$ was introduced, a reduction in the resultant upslope component was simulated (cases M5, M6; Figs. 7a, d), where the relative reduction is about the same in both cases. The $v$ component is noticeable only in the steeper slope case (Fig. 7e). The MBL height is almost unaffected in cases M5 and M6 as compared to cases M1 and M2, respectively.

c. The impact of longwave radiative flux divergence heating and solar radiation absorption on the upslope flows

As evaluated previously and as presented in section 2, the $\Delta$LWR heating is a nonnegligible thermal-forcing

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**Fig. 8.** The same as Fig. 5 except for cases M7 and M8.

**Fig. 9.** The friction velocity, $u_*$, along the simulated slope at 1400 MST, for cases M1–M8. Dark triangles indicate the extent of the slope.
of the MBL as compared to the surface sensible heat flux forcing. The absorption of solar radiation, on the other hand, only has a mild thermal effect (see Fig. 13). In order to evaluate the importance of the ΔLWR heating and the solar radiation absorption heating, simulations similar to cases M1 and M2 were carried out; however, while excluding the radiation heating (cases M7 and M8).

In the shallow slope case, case M7, the drop in the upslope component of the wind (see Fig. 8a) is about 0.5 of its value in case M1, while for the steeper slope, case M8, the drop is 0.65 of its value in case M2 (see Fig. 8d). Similar reductions were simulated also for the \( v \) component (Figs. 8b, e). The corresponding drop in the MBL depth is 0.5–1.0 km in cases M7 and M8 as compared to cases M1 and M2, respectively (see Fig. 8c, f).

d. Surface friction velocity and MBL turbulence

The surface friction velocity, \( u_* \), along the slope for the simulated cases M1 to M8, is presented in Fig. 9. The values of \( u_* \) correlate to the intensity of the surface wind speed, which is presented in Figs. 5–8. In the present simulations the \( u_* \) values range from \( \sim 2 \) cm \( s^{-1} \) to \( \sim 65 \) cm \( s^{-1} \), which are within the range computed by Sutton et al. (1978) based on the VL-1 and VL-2 surface wind observations.

The vertical exchange coefficients for heat, \( K_{ez} \), are computed based on the surface layer similarity relationships and a cubic polynomial functional approximation of \( K_{ez} \) within the MBL, assuming that at the MBL top \( K_{ez} = 0 \) (see Mahrrer and Pielke 1977). The computed vertical profiles of \( K_{ez} \) over the middle of the slope are presented in Fig. 10 where, as anticipated, the maximum values of \( K_{ez} \) increase with the depth of the MBL. The maximum values of \( K_{ez} \) in the various cases are in the range \( \sim 3800–5100 \) m\(^2\) s\(^{-1}\), where the average \( K_{ez} \) within the MBL is \( \sim 2000 \) m\(^2\) s\(^{-1}\). Typical values of \( K_{ez} \) within the Earth summer daytime boundary layer are 200 m\(^2\) s\(^{-1}\) (e.g., McNider and Pielke 1981). Thus, \( K_{ez} \) on Earth is about one order of magnitude lower as compared to \( K_{ez} \) on Mars [as also suggested by Eq. (53) in the analysis in section 2b]. The computed vertical exchange coefficients for momentum, \( K_{me} \), were typically smaller by a factor of \( \approx 6–8 \) compared to the corresponding values of \( K_{ez} \), providing a reasonable agreement with the scaling expressed by Eq. (10). The magnitude of the values of \( K_{ez} \) and \( K_{me} \) is similar to that reported in the model studies of Blumsack et al. (1973) and Burk (1976). Finally, sensitivity simulations in which the values of \( K_{ez} \) and \( K_{me} \) were increased/decreased by a factor of 2 indicated a relatively small effect on the predicted fields.

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**Fig. 10.** Vertical exchange coefficient for heat, \( K_{ez} \), at the middle of the slope at 1400 MST for cases M1–M6.

**Fig. 11.** The surface sensible heat flux, \( H_{so} \), along the simulated domain at 1400 LST for cases M1–M8.
e. The MBL thermal characteristics

The intensity of the daytime upslope flow is proportional to \( Q \) [i.e., the time integrated heating of the MBL since sunrise as expressed by Eq. (15)]. In order to gain a better insight into the characteristics of \( Q \), it is important to evaluate the relative contribution of each of the thermal components: sensible heat flux (\( H_s \)), net longwave flux divergence (\( \Delta LWR \)), and solar absorptivity (\( \Delta SWR \)). Note that in the following, when surface fluxes are considered, we used the sign convention that all the fluxes are positive toward the surface. As mentioned previously, it was found that the values of \( H_s \) are affected considerably by the surface wind intensity. Comparison with the surface components, presented previously (in Figs. 5-8) and with the \( u_0 \) values (Fig. 9), shows a correlation between these variables and \( H_s \). This situation is not similar to that on the Earth where the surface wind speed has a relatively mild effect on \( H_s \) (e.g., Carlson and Boland 1978). Based on \( H_s \) values obtained in the present simulations, it is suggested that horizontal gradients in \( H_s \), due to changes in the surface wind speed, are likely to be nonnegligible on Mars.

The reduction in \( H_s \) as the wind speed drops was shown to be mostly balanced by an increase in the surface temperature and by a corresponding increase in the outgoing longwave radiation (Fig. 11). The typical range of \( H_s \) values in the various simulations is 15°–30°W m\(^{-2}\), which is about 10–15 times smaller compared to those obtained on Earth under similar simulations (e.g., Ye et al. 1987) and is in agreement with Eq. (51). Worth noting is that this magnitude of the daytime Martian surface sensible heat fluxes is comparable to those typical to slopes over Earth during the nocturnal period (e.g., Ye et al. 1989). The net longwave radiation flux divergence, \( \Delta LWR \), within the MBL is given in Fig. 12, indicating an increase in its value with the increase in the MBL depth towards the slope top. When the winds are relatively intense the values of the \( \Delta LWR \) are about 0.5 of the corresponding values of \( H_s \). However, when the surface wind speed decreases, and consequently \( H_s \) also decreases, the values of \( \Delta LWR \) and \( H_s \) become comparable.
The direct heating of the MBL by the absorption of solar radiation (Fig. 13) is typically insignificant compared to the heating amount due to the accumulative contribution of $H_\alpha$ and $\Delta LWR$. Typically, the direct solar heating is about 10% of that amount. As stated previously, in the Earth's boundary layer the surface sensible heat flux dominates the $\Delta LWR$ and $\Delta SWR$ heating contribution.

4. Conclusion

The present study investigated the daytime thermally induced upslope flow characteristics on Mars (including the range of values of the maximum upslope flow, the depths of upslope flow and the convective boundary layer, the temperature perturbation in the atmosphere as well as the impact of background atmospheric thermal stability, and terrain slope on the thermally induced upslope flow). The characteristics of upslope flow on Mars were compared with that on Earth, using analytical evaluations and 2-D numerical modeling. The purpose of the analytical evaluations is to provide a physically rational general insight into the understanding of the problem. The numerical model simulations were carried out in order to provide a detailed evaluation for specific cases. The major conclusions we obtained from this study are as follows:

- The summer daytime thermally induced upslope wind is about 10 m s$^{-1}$ for slope steepness of less than 1°. The depths of upslope flow and the convective boundary layer do not extend above 5 km. Moderate supportive or opposing background flows change the
strength of the thermally induced upslope wind in a nearly additive manner.

- Unlike the Earth, the sensible surface heat fluxes on Mars are noticeably dependent on the intensity of the surface wind speed.

- The longwave flux divergence heating is a significant factor in the daytime development of the upslope flow and of the MBL, mostly when the slope is relatively shallow.

- Comparison of various characteristics related to daytime induced upslope flows on Mars and Earth indicated that, typically: On Mars, the upslope wind speed is about 2.5 times as large; the air temperature warming near the surface, the depths of upslope flow, and the convective boundary layer are 3–4 times as large; the friction velocity is about 3 times as large; the eddy exchange coefficient is 10 times as large; and the kinetic energy is about 0.25 times as large as on Earth.

- There is general agreement between the results of the analytical analysis and those obtained in the numerical modeling simulation.

Acknowledgments. The study was supported by the National Science Foundation under Grant ATM-8414181 and ATM-8616662. The computations were carried out by the NCAR Computer Facility (NCAR is supported by the NSF). M. Moran provided useful editorial comments on the early version of the manuscript. We would like to thank B. Critchfield and D. McDonald for the preparation of the manuscript.

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