

Science

Martian mesoscale and microscale wind variability of relevance for dust lifting

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Abstract

Background: Mars is both a windy and dusty environment. Airborne dust is a crucial climate component on Mars. It impacts atmospheric circulations at large-, meso- and micro-scales, which in turn control dust lifting from the surface and transport in the atmosphere. Dust lifting processes and feedbacks on atmospheric circulations are currently not well understood.

Method: Our purpose is to show how mesoscale models and large-eddy simulations help to explore small-scale circulation patterns which are potentially important for lifting dust into the atmosphere but which are unresolved by global climate models. We focus on variations of friction velocity, u^* , relevant for dust lifting, in particular investigating maximum values and the spatial and temporal variability of u^* .

Conclusion: Meteorological scales between 100 km and 10 km can be studied by high-resolution global circulation and limited-area mesoscale models, which both show strong topographic control of the daytime and nighttime near-surface winds. Scales below 10 km and 1 km are dominated by turbulent gusts and dust devils, two distinct convective boundary layer processes likely to lift dust from the surface. In low-latitude regions, boundary layer depth and friction velocity u^* are correlated with surface altimetry. Further studies will be carried out to parameterize lifting by boundary layer processes and dust radiative effects once transported in the atmosphere.

Introduction

Mars is a windy environment. Strong winds are associated with the rapid response of the thin Martian CO₂ atmosphere to radiative forcing. This is particularly true in the lower troposphere, where large diurnal variations and horizontal gradients of surface temperature, along with topographical contrasts, yield large spatial and temporal variability in those intense winds. The Martian tropospheric circulation is active at all meteorological scales: large-scale (>100s kilometers), mesoscale (>100s meters), and microscale (≤100s meters). On the large-scale, the circulation is characterized by inter-hemispheric meridional Hadley Cells ([Wilson 1997](#); [Forget et al. 1999](#)), high-amplitude thermal tides ([Wilson and Hamilton 1996](#); [Lewis and Barker 2005](#)) and mid-latitude baroclinic wave variability ([Collins et al. 1996](#); [Hollingsworth et al. 1996](#)). On the mesoscale, it is characterized by intense katabatic and anabatic winds, particularly over craters, volcanoes, canyons ([Rafkin et al.,](#)

[2002](#); [Tyler et al. 2002](#); [Toigo and Richardson 2003](#); [Spiga and Forget 2009](#)), thermal circulations induced by soil thermophysical heterogeneities ([Toigo et al. 2002](#); [Kauhanen et al. 2008](#)), local dust storms ([Rafkin 2009](#)) and regional-scale transient eddies, mostly in polar regions ([Tyler and Barnes 2005](#)). On the microscale, it is characterized by daytime convective and nighttime shear-induced turbulence in the boundary layer ([Michaels and Rafkin 2004](#); [Tyler et al. 2008](#); [Spiga et al. 2010](#)).

Mars is also a dusty environment. Mars' atmosphere always has a thin veil of suspended dust particles, the amount varying with location and season. Atmospheric dust on Mars absorbs incoming sunlight mainly in visible wavelengths (and outgoing infrared radiation), which locally warms the Martian troposphere. Even in moderately dusty situations, the influence of dust on Martian thermal structure is critical. When large quantities of dust are injected into the atmosphere by regional or planet-encircling dust storms

(Cantor 2007; Malin et al. 2008), intense atmospheric heating occurs while 80%, or more, of the incoming sunlight is absorbed by dust. Airborne dust is therefore a crucial climate component on Mars which impacts atmospheric circulations at all scales. At the same time, the amount of dust in the atmosphere is controlled through lifting and transport by winds at various scales. Mechanisms by which dust is lifted from the surface and injected into the Martian atmosphere, as well as the resulting feedback on atmospheric circulation, are currently not well understood. This lack of understanding is exemplified by the ongoing mystery surrounding the growth of some local dust storms into planet-encircling dust events with irregular inter-annual variability (Montabone et al. 2005).

Martian missions yielded numerous observations of structures formed by dust lifted and transported by atmospheric winds at all scales. Figure 1 describes this diversity of dust “storms” in the Martian atmosphere and emphasizes how the dust cycle on Mars involves horizontal scales spanning at least seven orders of magnitude: from dust fronts extending over thousands of kilometers to dust devils and turbulence on meter and sub-meter scales. At the same time, measurements of wind remain scarce on Mars. Near-surface winds have only been estimated in-situ in the Chryse (Viking 1, Pathfinder), Utopia (Viking 2) and Vastitas Borealis (Phoenix) plains. Insights into Martian wind

variability at various scales can be gained from such measurements (Barnes 1980; Savijarvi and Siili 1993), as well as from indirect observations of wind activity through imaging of aeolian erosion structures (Nayvelt et al. 1997; Greeley et al. 2003) or analysis of cloud morphology (Michaels et al. 2006). Nevertheless, physically-consistent meteorological modeling remains necessary to fully characterize and understand the intense Martian circulations over a range of various scales, regions and seasons and to diagnose their ability to lift and transport dust away from the surface.

As shown in Figure 1, modeling the Mars dust cycle and the key lifting, transport, radiative processes necessitates the use of different and complementary modeling tools. Each type of meteorological scale has a dedicated three-dimensional atmospheric model that can be used to expand the knowledge of Martian winds contributing to the dust cycle: global circulation models (GCMs) (e.g., Newman et al. 2002), mesoscale models (e.g., Rafkin et al. 2002), and large-eddy simulations (microscale models) (e.g., Michaels 2006). GCMs do not typically generate surface wind stresses of sufficient amplitude directly to initiate significant dust lifting. This is because GCMs explicitly represent only the large-scale, slowly-varying components of the global circulation; if winds associated with these components were sufficiently strong to lift dust routinely, then the Martian atmosphere

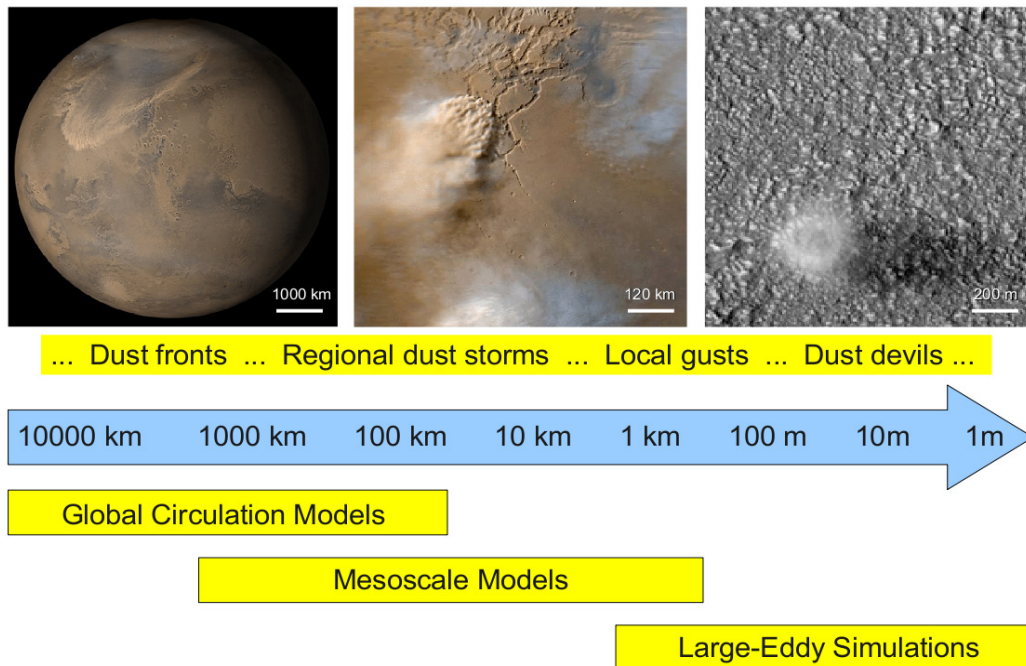


Figure 1. Typical dust lifting events, relevant spatial scales and meteorological models suitable for their analysis. Left: Dust front observed through Mars Orbiter Camera imagery and analyzed by Wang et al. 2003. Middle: Dust regional storm observed with Mars Color Imager as described in Malin et al. 2008. Right: Dust devil observed with High Resolution Imaging Science Experiment; review of such phenomena can be found in Balme and Greeley 2006. Note that 1) the right limit of GCM box keeps on being translated towards the right, thanks to advances in computational resources and modeling techniques, and 2) mesoscale modeling and (turbulent-resolving) Large-Eddy Simulations can be often carried out with the same non-hydrostatic dynamical core.

would be continuously dusty. One possible area of model improvement is to determine the wind variability, relevant to dust lifting, which occurs within grid boxes of order tens to hundreds of kilometers across in global models. The purpose of this paper is to show how mesoscale models and large-eddy simulations might help to explore small-scale circulation patterns which are responsible for injection of dust into the atmosphere but which are left unresolved by global models. From lifting to transport to radiative feedbacks to sedimentation, studying dust processes on Mars remains an open and crucial topic, illustrating how the atmosphere and the surface are strongly coupled in the climate system. In such a vast topic, this report offers some preliminary discussions as a starting point for further studies.

Scope

It is well known that strong winds are able to lift dust from the surface into the air. Yet the injection of dust into the free atmosphere, beyond the lowest couple of meters of the boundary layer, is far from straightforward and not fully understood (Greeley and Iversen 1985). Under current assumptions for the Martian soil characteristics, strong near-surface winds would tend to catch dust particles large in size, of order 10-100 μm (Pollack et al. 1976; Newman et al. 2002, their Figure 1), which might fall relatively quickly to the ground without being transported very far. Only smaller dust particles could remain in suspension for very long times once in the free atmosphere: the predominant particle size in the atmosphere is estimated to be of order 1 μm (Wolff et al. 2006). Nevertheless, owing to electrostatic, intermolecular and weak magnetic interactions causing those particles to stick together, on Mars it appears more difficult to directly lift those from the surface – some studies even suggest that it is virtually impossible given the unrealistically high values of winds required for such events to occur (Pollack et al. 1976; Newman et al. 2002).

This problem leads to the need to make a distinction between fluid and impact thresholds for wind speed, associated with direct lifting of dust particles through wind stress alone, and lifting *via* saltation respectively (Bagnold 1941). Saltation is a process in which larger particles are lifted into the air and quickly fall out by sedimentation, “kicking” smaller particles into the atmosphere while impacting the surface and so overcoming their larger interparticle cohesion (Greeley and Iversen 1985). The latter mechanism may explain how smaller particles can be lifted from the surface and transported into the atmosphere by atmospheric winds on Mars (Greeley 2002). The fluid threshold is generally represented by defining a threshold drag velocity u^*_t , which must be exceeded by the actual friction velocity u^* for lifting to occur. Friction velocity u^* is related to near-surface wind stress τ and atmospheric density ρ through the relationship $\tau = \rho u^{*2}$. In atmospheric levels close to the surface, velocities vary approximately logarithmically with height $ku(z) = u^* \ln(z/z_0)$ where $k = 0.4$ is the von Karman constant and z_0 the roughness length (a value of 1 cm for z_0 , thought to be typical for Martian conditions, is adopted in this study similarly to most studies in the literature, *e.g.*,

Forget et al. (1999), Newman et al. (2002)). Hence u^* can be deduced through the relationship $u^* = ku(z_1)/\ln(z_1/z_0)$ from near-surface winds $u(z_1)$ simulated by a meteorological model at an altitude z_1 above the surface.

In this paper, we focus on mesoscale and microscale variations of friction velocity u^* relevant for dust lifting, in particular maximum values of u^* . Note that we discuss only the day-to-day wind activity that yields dust lifting, leaving out discussions about episodic regional dust storms in the present work. This is obviously only one component of the complex problem of lifting Martian dust from surface to atmosphere by interacting atmospheric phenomena. Another key aspect consists in determining the dependence of threshold velocity u^*_t on particle density, particle size and atmospheric density, usually through semi-empirical relationships retrieved from laboratory experiments covering a wide range of conditions for terrestrial planets. Accounting for the saltation mechanism is also clearly central in dust lifting studies and subject to active research (*e.g.*, saltation may occur for much lower wind speeds than previously thought, according to Kok (2010)). All these aspects are left for discussion in other papers, as well as the finite availability of dust on the Martian surface, variations of roughness length and departures from laboratory-derived idealized saltation scenarios. Instead, we focus on the dynamical processes causing maxima of friction velocities u^* and their temporal and spatial variability. The rationale is to separate unambiguously the influence of dynamical changes of u^* from other factors: we assume that the injection of dust in the atmosphere is only limited by the ability of the atmosphere to generate wind of sufficient strength to lift dust from the surface *via* saltation. Our analysis is based on meteorological maps of friction velocity u^* obtained through atmospheric modeling with radiative feedbacks of lifted dust on circulations not included in the initial analysis.

Mesoscale variability

In this section, we discuss the variability of u^* at typical horizontal scales of ~10s kilometers in low-latitude regions. Mesoscale winds in high-latitude areas are not examined. As detailed in Toigo et al. (2002), near-surface wind variability in polar regions is governed by the combination of storm track activity (*i.e.* baroclinic waves) with thermal circulations induced by icy vs. bare surface contrasts and topographical gradients at the edge of the caps. This requires particular attention in simulating the limits of seasonal polar caps, which is beyond the scope of the present paper. In order to determine wind variability at the mesoscale in low-latitude regions, we adopt a comparative approach with two distinct tools. The beginning of the northern summer season is considered (solar longitude $L_s = 90^\circ$).

On the one hand, we use global circulation modeling with unusually high spatial resolution: the model is run with grid spacing 0.7 degree of latitude / longitude (~ 40 km) instead of the more typical 3-5 degrees of latitude / longitude. GCMs

integrate the equations of fluid dynamics over the whole planetary sphere, normally under the hydrostatic approximation. We use the [Lewis and Read \(2003\)](#) model, also referred to as the UK Mars GCM. This model employs spectral discretization of the geophysical fluid dynamics equations (hence, horizontal resolution is only approximately 40 km, since it is designated through truncation of spherical modes, “T170” here which implies that modes up to a total wavenumber of 170 are retained in the model). The model is not started from rest but from a coarse-resolution T31 (3.75 degree) simulation run for ~ 10 Martian years so as to reach equilibrium (*i.e.*, stable interannual cycles) after spin-up. The GCM used alternatively at coarse or fine resolutions features similar settings, except for a modified hyper-diffusion coefficient and lower timestep in the fine resolution case to avoid numerical instabilities.

On the other hand, we use limited-area mesoscale modeling at typical horizontal resolution for regional-scale studies (*i.e.*, 20 km grid spacing; computations are carried out for a few days at the considered season). Mesoscale models typically integrate the nonhydrostatic¹ atmospheric circulation with improved horizontal and vertical resolutions compared to more typical GCM simulations. Calculations are performed on a limited area (here, the Tharsis region) with initial and boundary conditions derived from GCM predictions. We use the [Spiga and Forget \(2009\)](#) grid-point mesoscale model, also referred to as the LMD [Laboratoire de Météorologie Dynamique] Mars mesoscale model. This model has the notable feature of sharing the same physical parameterizations as the LMD GCM, used to provide initial and boundary conditions (see [Forget et al. \(1999\)](#) for details about parameterizations of the Martian environment, not detailed here for the sake of brevity). This is important for ensuring the best downscaling efficiency possible. Moreover, the UK GCM also features physical parameterizations and Mars Global Surveyor surface properties similar to the LMD mesoscale model, which simplifies comparison. One of the only actual differences in terms of physical parameterizations between the mesoscale and GCM tools used in the present study is the slope insolation scheme featured in the former ([Spiga and Forget 2008](#)).

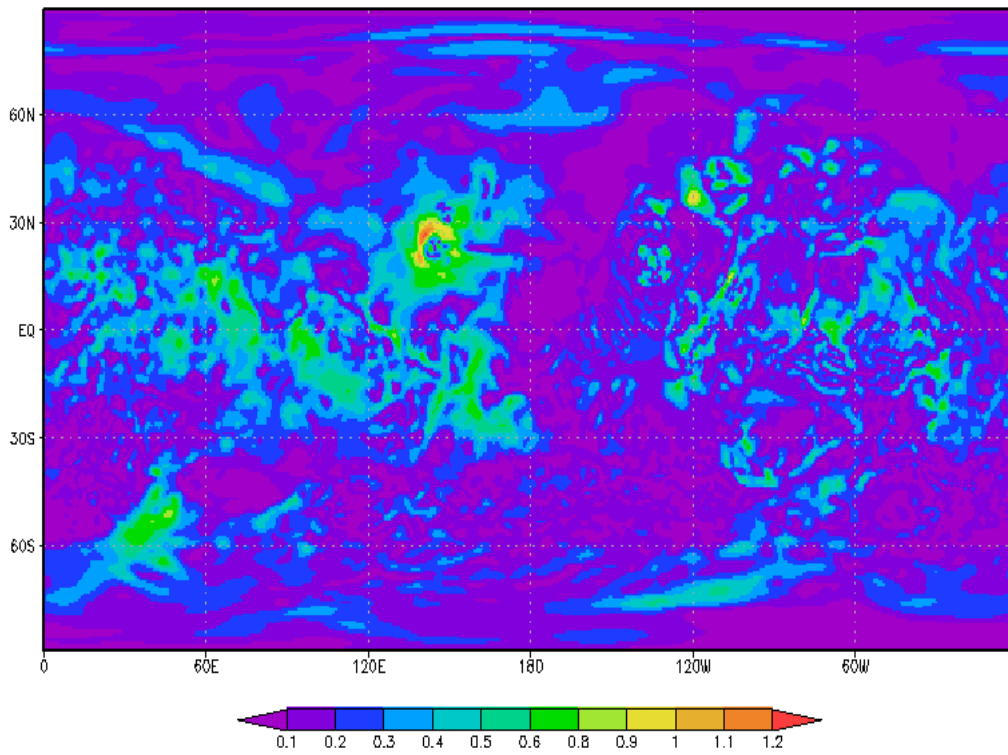
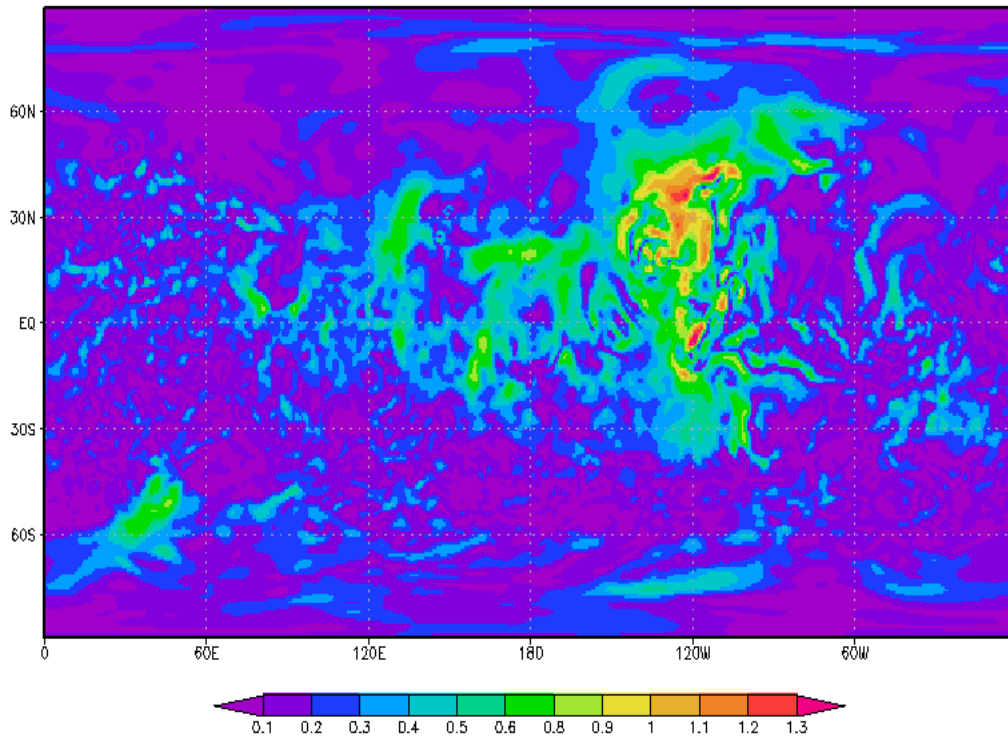
Both models calculate horizontal winds roughly 2-3 meters above the surface. We used those predictions in the first vertical level above the ground to calculate friction velocities according to the method detailed in the previous section. Typical results from high-resolution UK GCM simulations are shown in Figure 2. Both in daytime and nighttime conditions, areas of maximum friction velocity appear to be strongly related to the largest topographical obstacles on Mars. The main regional-scale meteorological phenomena accounting for wind variability near the surface are slope

¹The hydrostatic approximation $\partial p = -g \partial z$, where p is pressure, g is gravity and z is altitude, is valid for moderate vertical motions in the atmosphere. This is the case for large-scale circulations, but not for mesoscale circulations taking place at horizontal scales of ~ 20 km and below. Hence, mesoscale models, contrary to GCMs, feature the complete equations of motion without the hydrostatic approximation so as to accurately compute local-scale circulations.

winds. In the first 100s of meters above the surface, nighttime cooling and daytime warming impose terrain-following behavior of the atmospheric temperature over slopes, hence baroclinic production as gradients of pressure and density are not colinear. A so-called “slope-buoyancy” pressure gradient forms and gives rise to afternoon anabatic (upslope) and nighttime katabatic (downslope) atmospheric motions, two to three times stronger on Mars owing to short radiative timescales and low thermal inertia of the thin CO₂ atmosphere (see [Spiga \(2010\)](#) and references therein).

A key factor to observe clear-cut slope circulations is a relative weakness in large-scale background winds so that the “slope-buoyancy” force dominates the large-scale pressure gradient force. Owing to the significant influence of large-scale circulations, the global maps of friction velocity u^* shown in Figure 2 are not completely correlated with slopes. The influence of thermal tides in subtropical latitudes on friction velocity maxima can be clearly seen on the global maps. The fact that Alba Patera features some of the highest friction velocities, despite its relatively smooth slopes, highlights that large-scale influence. As a combination of both slope winds and thermal tides, the locations and intensities of friction velocity maxima undergo a distinctive diurnal cycle. Overall values of friction velocities are larger in the day than during the night. Notwithstanding this, note that friction velocity is only one aspect of dust lifting: wind stress (density times friction velocity squared) might be higher during the night owing to higher density of colder air. Some areas of high friction velocity do not show significant diurnal variability. This is notably the case for Hellas' southeastern rim, which is known to be an area of a sustained southerly jet (playing a peculiar role in the formation of glaciers in past Martian climate, see [Forget et al. \(2006\)](#)).

The highest value of friction velocity u^* is reached over Arsia Mons by the end of the afternoon. The maximum friction velocity is 1.4 m/s, which is higher than predicted by lower resolution simulations (less than 0.9 m/s, figure not shown for sake of brevity). We chose this region to compare results from high-resolution global circulation models with outputs from mesoscale modeling (shown in Figure 3). Wind patterns are similar in both LMD mesoscale simulations and high-resolution UK GCM computations. It should be noted that GCMs at 40 km resolution only slightly underestimate the anabatic wind maximum compared to the 20 km resolution mesoscale model. This maxima is associated with a strong front (*i.e.*, an area limited in space where variations of temperature and winds are particularly intense) over Arsia Mons, caused by the constructive influence of anabatic flow and a positive maximum in the thermal tide (in comparison, wind intensity two hours before is lower). This feature is present both in high-resolution global and limited-area mesoscale simulations, though it is more sharply resolved in the latter. In addition to topographical gradients, the surface temperature field shows that thermal contrasts could drive a component of the circulation, owing to the local variations of albedo and thermal inertia (see Figure 3c). Given the direction of prominent surface temperature gradients, the contribution of thermally-driven circulation might explain



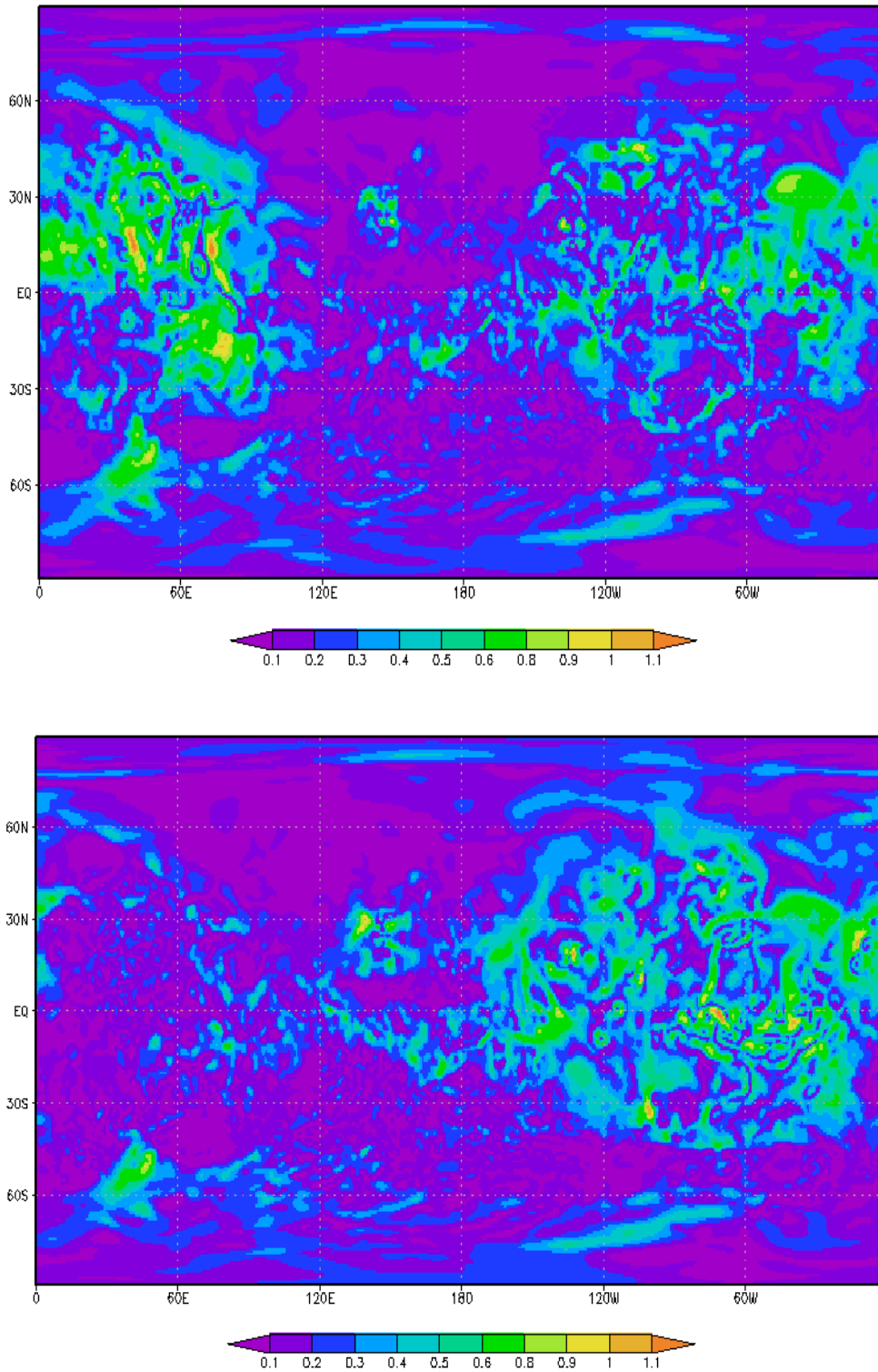
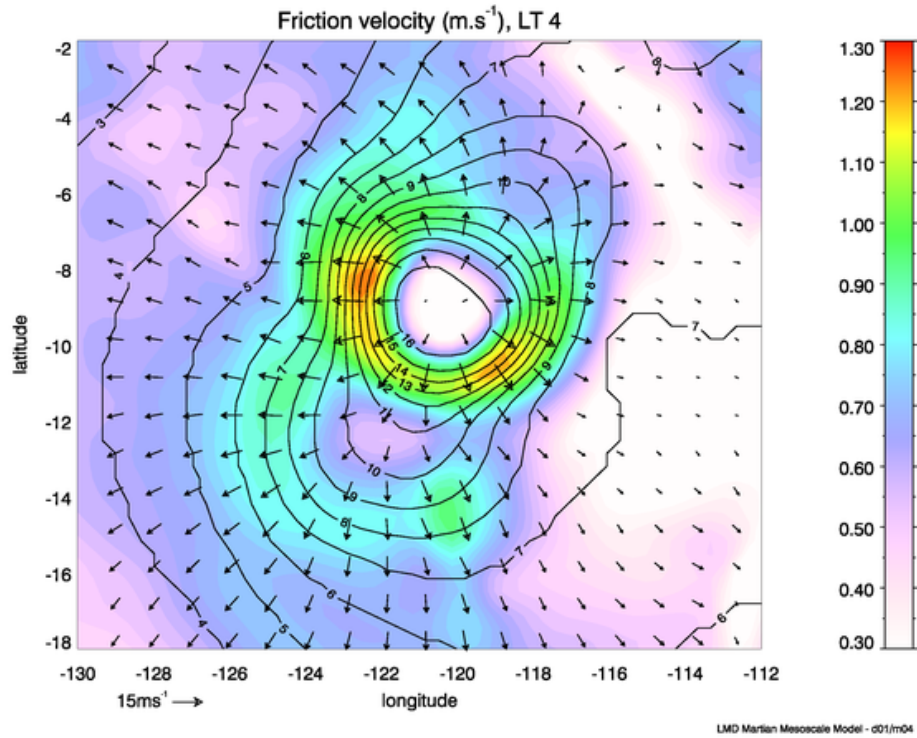
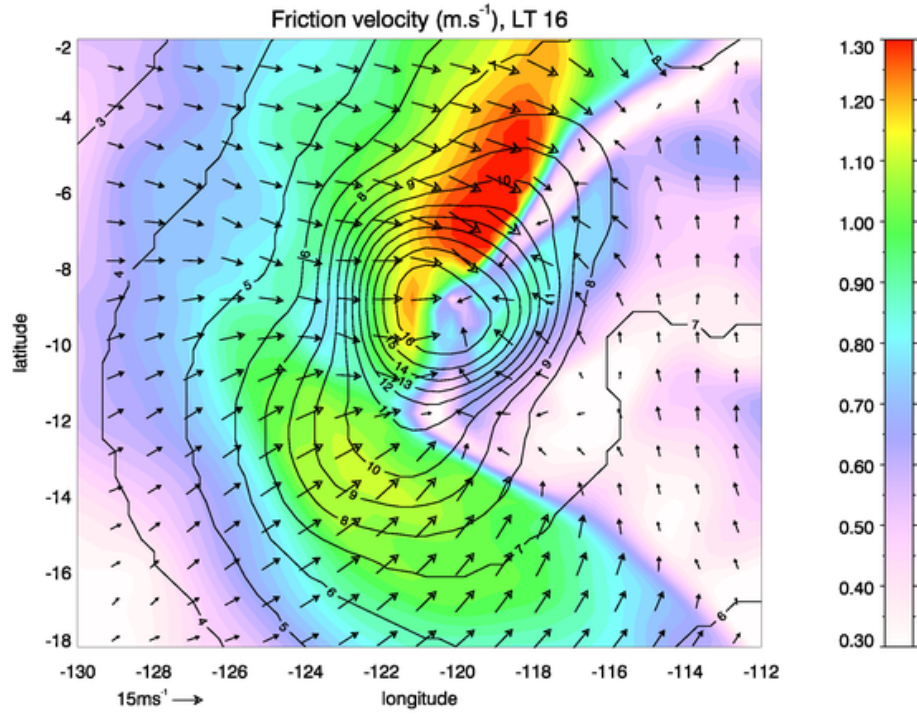


Figure 2. Friction velocities u^* predicted in northern summer ($L_s \sim 90^\circ$) at universal times (local times at longitude zero) 00h (a), 06h (b), 12h (c), 18h (d). High-resolution global circulation modeling by the [Lewis and Read \(2003\)](#) spectral model (40 km resolution). Note that nighttime katabatic winds around Olympus Mons (see map b) have some specific artifacts whose cause is known. A correction is under investigation for future runs (detailed discussions will be included in a forthcoming publication).



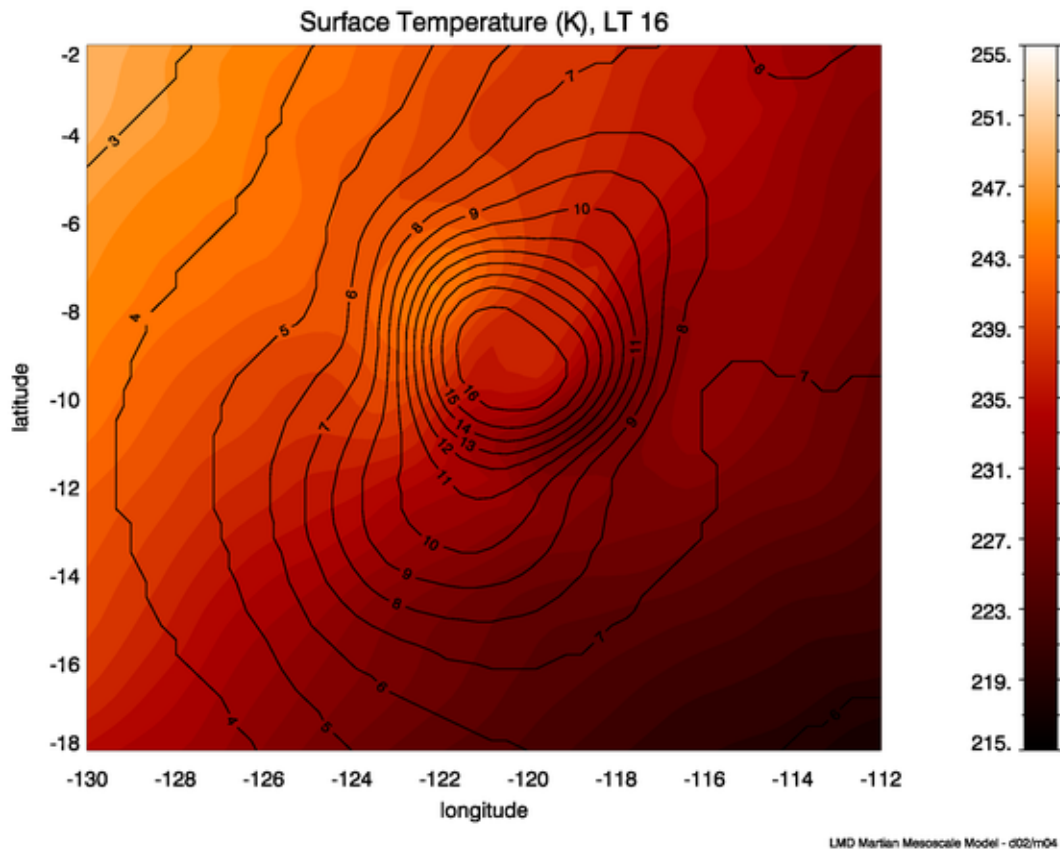


Figure 3. Friction velocities u^* and wind vectors 10 meters above the surface predicted in northern summer ($L_s \sim 90^\circ$) at local times 16h **(a)** and 04h **(b)**, corresponding to universal times $\sim 00h$ (see Figure 2a) and 12h (see Figure 2c). Mesoscale modeling in the Arsia Mons region by the [Spiga and Forget \(2009\)](#) mesoscale model (20 km resolution) forced by LMD GCM fields at its boundaries (note that any value above 1.3 m/s is represented in red; maximum predicted value is 1.45 m/s). In frame **(c)**, surface temperature predicted by the model at local time 16h is shown.

enhancements of winds in a northeast – southwest axis.

All these elements tend to confirm, as suggested by [Rafkin et al. \(2002\)](#), that Arsia Mons might constitute a preferential area for dust lifting from the Martian surface. Above slopes where near-surface winds are developed, these enhanced lifting conditions are associated with potential for transport of dust to large horizontal and vertical distances from the area of lifting (this is not necessarily true in the night since katabatic winds are downslope). It can be noted that Arsia Mons anabatic winds are stronger than over Olympus Mons. Arsia Mons might represent the perfect setting for afternoon slope winds, i.e. a trade-off between slope being steep enough for slope acceleration to be significant, but not too steep to optimize exposure to incoming sunlight. The same combination accounts for friction velocities over Elysium Mons being larger in surrounding slopes than over the central peak (see the global map at universal time 06:00 in Figure 2b). This trend is not observed (as is confirmed by the simulations) for katabatic winds which, being independent of incoming sunlight, merely increase with slope steepness. This might explain why the underestimation of friction velocity in global circulation simulations (where topography is less well resolved than in mesoscale simulations) is more

significant in nighttime than in daytime.

Overall, GCM maps of friction velocity showing significant correlation between u^* maximum and slope steepness tend to show that horizontal resolution is an important element for accurately simulating dust lifting mechanisms and tendencies. However, comparisons between the two modeling strategies adopted in this paper also indicate that large-scale global models run at high resolution allow for correct first-order characterization of the near-surface mesoscale wind variability compared to finer mesoscale modeling. Furthermore, the consistency of results between both simulations builds confidence that a mesoscale model, run with GCM predictions imposed at its boundaries, is able to accurately reproduce features of the large-scale circulation. We must admit there is a caveat in the Martian case: the influence of local-scale circulations so dominates the large-scale influence, especially near steep slopes, that small errors on the large-scale pressure gradient might have very little effect on the small scale winds. For instance, simulations of the Valles Marineris canyon by [Richardson et al. \(2007\)](#), which did not include large-scale forcing, yielded similar quantitative results for the near-surface wind speeds as other studies including the influence of ambient winds

(Tyler et al. 2002; Toigo and Richardson 2003; Rafkin and Michaels 2003; Spiga and Forget 2009). Nonetheless, we find generally good agreement between high-resolution global and mesoscale limited-area modeling.

Microscale variability

The influence of small-scale turbulent winds in the boundary layer is crucial to determine how much dust is lifted from the Martian surface and injected in the atmosphere. Convective circulations are particularly well developed in the Martian boundary layer during the afternoon under the influence of the heated surface. Even in situations of weak large-scale and mesoscale winds, this thermally-driven convection might itself cause significant vertical mixing and wind gustiness, hence much larger wind maxima than what large-scale and mesoscale diagnostics suggest. In GCMs and in mesoscale models, mixing induced by those daytime convective circulations is parameterized, for the turbulent vertical motions and horizontal gustiness cannot be resolved. In such models, maximum winds might be strongly limited and incomplete. Since the beginning of the 2000s, the dynamics of the Martian convective boundary layer has been analysed by means of Large Eddy Simulations (LES) (Rafkin et al. 2001; Toigo et al. 2003; Michaels and Rafkin 2004; Tyler et al. 2008; Spiga et al. 2010), where grid spacing in Martian mesoscale models is lowered to a few tens of meters so as to resolve the larger turbulent eddies, responsible for most of the energy transport within the convective boundary layer. Boundary layer growth, convective cells and dust devils are described in detail by such simulations; the relevance of such simulations in dust cycle studies have been stated in recent studies (Toigo et al. 2003; Michaels 2006). Note that we do not discuss results in nighttime conditions, where a different kind of turbulence (shear-induced) takes place. This variability has not been addressed in Martian studies yet: it requires a significant amount of work and is considered out of the scope of the present paper. Terrestrial studies have shown that nighttime LESs require finer grid spacing and more sophisticated sub-grid scale parameterizations than daytime LESs (e.g., Stull 1988).

In the present discussion, we use the 50m-resolution LESs carried out with the LMD mesoscale model to help interpret Mars Express radio-occultation measurements of boundary layer depth (Hinson et al. 2008). Details of those simulations are given in the Spiga et al. (2010) paper, where encouraging agreement is reported between mixing layer depths measured by Mars Express and predicted through (windless free convection) LESs. In particular, in low to mid-latitudes where surface temperature is rather uniform, the convective boundary layer depth is found to be positively correlated with surface altimetry. This peculiar variability of turbulent convection with pressure is caused by the radiative control of the Martian boundary layer (see Spiga et al. (2010) for further details). For instance, mixing layer depth is 2 km higher over Tharsis plateaus (2.5 km above zero datum) than over the lower-altitude Amazonis plains (3.6 km below zero datum). Of particular interest in the frame of the present discussion about dust lifting is the fact that LESs reveal the

boundary layer dynamics (and, in particular, near-surface winds) associated with the observed regional variability of boundary layer depth.

The evolution of friction velocity u^* in two LESs over Amazonis plains and Tharsis plateaus is shown in Figure 4. In Spiga et al. (2010), it was reported that boundary layer height is respectively 5 km and 8 km, which yields maximum updrafts of respectively 12 and 18 m/s (for similar values of daytime surface temperatures). These facts are useful for dust transport and mixing in the boundary layer, but more importantly for dust lifting, similar contrasts can also be noted for the near-surface turbulent horizontal wind. Figure 3 shows that, while u^* maximum values reach 0.85 m/s in Amazonis plains, values over Tharsis plateaus reach 1.2 m/s. Thus, the two regions are not equivalent in terms of ability to lift dust (note that here we only consider friction velocity, but there is an additional, opposing effect of lower density on Tharsis plateaus compared to Amazonis plains). The strongest winds are predicted between 12h and 14h30. Over short timescales, the behaviour of u^* exhibits turbulent fluctuations, where extreme values are followed by much lower values. Nevertheless, for instance between 13h and 14h, the maximum friction velocity attained in the LES simulation domain (14 km \times 14 km) never goes below 0.6 m/s in Amazonis plains and 0.8 m/s over Tharsis plateaus. In Figure 4, average values of u^* in the simulation domain are also plotted in order to show that maximum values allow for a better characterization of turbulent wind variability. The latter are also the most relevant element for dust lifting studies, where maximum winds are key compared to mean winds. It is however interesting to note that boundary layer gustiness is never zero and is always quite significant in afternoon hours.

It should be remembered that these results are obtained through LESs in windless conditions and represent unresolved wind gustiness within a grid point where “background” winds at larger scales are zero. In other words, where GCM and mesoscale models tend to be the furthest from predicting conditions for dust lifting, turbulent components might account for friction velocities above 0.8 m/s in brief “gusty” episodes. A broader conclusion for the Tharsis regions is that it could represent a key region for dust lifting, because mesoscale and microscale wind variability are strong in this region: both slope winds and turbulent gustiness contribute at different scales to help lift dust from the surface. Running large-eddy simulations with background winds or turbulent-resolving integrations of anabatic/katabatic flow would allow for a better assessment of coupling between the two dynamical effects and its consequence on dust lifting in topographically uneven terrains.

Which boundary layer phenomena correspond to the maximum predicted values for u^* ? In strongly heated conditions, boundary layer circulations follow an horizontal structure in polygonal convective cells. At the intersection of those convective cells, the flow organizes itself into an intense vortical structure with strong updrafts and low

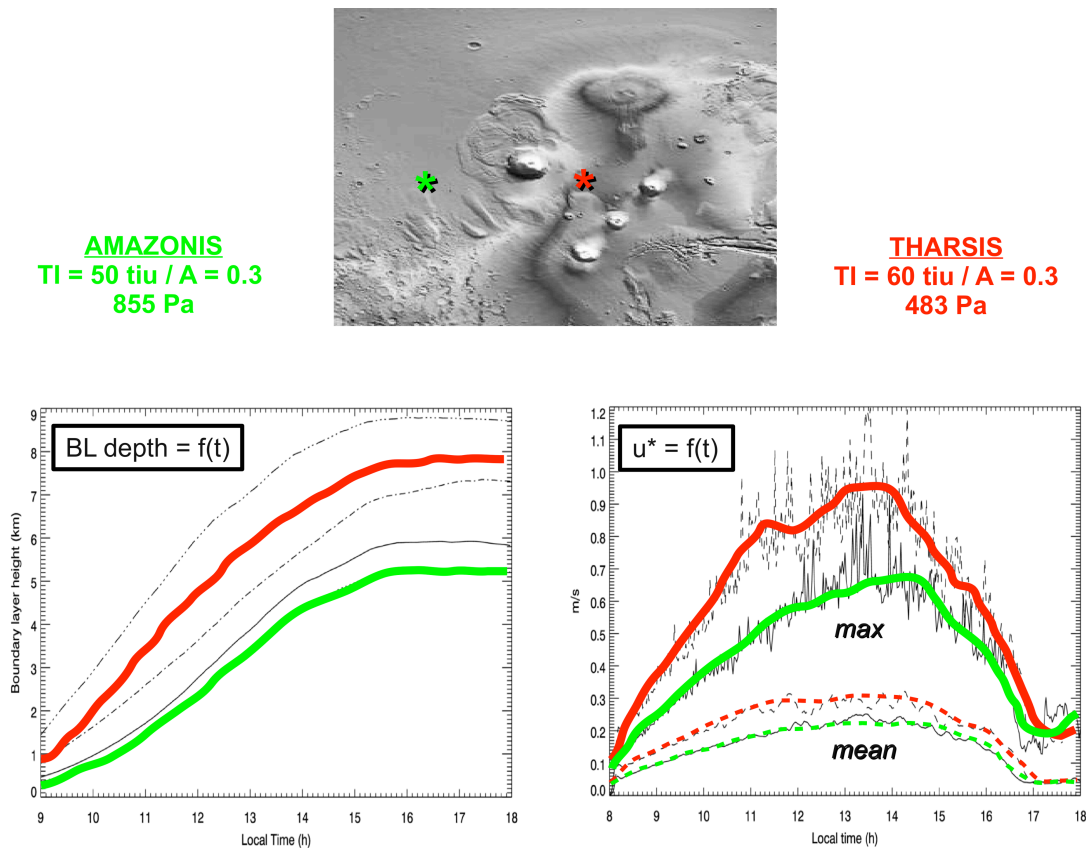


Figure 4. Evolution diagnosed from Large-Eddy Simulations carried out with the [Spiga and Forget \(2009\)](#) mesoscale model of (left) boundary layer depth and (right) friction velocity u^* in two locations with similar soil properties but distinct altitudes (Amazonis plains in green and Tharsis mountains in red). Amazonis and Tharsis large-eddy simulations are detailed in [Spiga et al. \(2010\)](#) with respective labels *b* and *c*. Boundary layer depth at local time 17h00 is predicted in satisfying agreement with values derived from Mars Express radio-occultation experiments. Values of u^* in full (dashed) lines correspond to maximum (mean) values in the simulation domain.

pressure core at its centre. Note that, as mentioned in earlier studies, predicted convective vortices share all the characteristics of observed dust devils and form “dustless devils” in the present LES simulations where dust lifting is not included. Strong winds not related to cyclostrophic formations (“convective gusts”) can also be found near walls of convective cells where significant horizontal convergence occurs. The two distinct phenomena result in u^* being maximum in the convective boundary layer², as shown in Figure 5, but are not exactly equivalent with regards to dust lifting. Convective vortices have an enhanced ability to both lift and inject dust in the atmosphere (which is, actually, confirmed by numerous observations of dust devils events or tracks at the surface of Mars) because (1) values of u^* associated with this kind of event are stronger, for near-surface winds in a convective vortex are larger in a dynamical system governed by cyclostrophic balance; and (2) the magnitude and vertical extent of vertical winds are

²While intensity of convective gusts is similar in 50 m and 100 m horizontal resolution simulations, convective vortices are poorly resolved in 100 m simulations and their u^* signatures are much weaker than those in the 50 m resolution case. Future studies of dust lifting induced through boundary layer turbulence would likely also benefit from 20 m or 10 m resolution simulations.

enhanced in convective vortices compared to convective gusts. The majority of maximum u^* values shown in Figure 4 between 12h and 15h correspond to the activity of convective vortices rather than “convective gusts”.

There are other differences worthy of mention between dust devils and “regular” gustiness. Since the circulation in convective vortices is organized according to the cyclostrophic equilibrium, the formation of a convective vortex is associated with a low-pressure core. It has been suggested that this depression could induce an enhanced lifting of dust, as laboratory measurements seemed to show this effect ([Greeley et al. 2003](#)). In addition, as dust devils concentrate a large amount of dust particles on their walls, triboelectric effects caused by the effective friction of dust particles colliding with each other might be enhanced and could influence the ability of dust particles to be lifted through electrostatic effects ([Farrell et al. 2004](#)).

Our large-eddy simulations show that, owing to the dominant radiative control of the Martian boundary layer, high-altitude terrains such as Tharsis plateaus are conducive to faster near-surface horizontal winds within convective vortices, which could facilitate dust lifting and dust devil formation

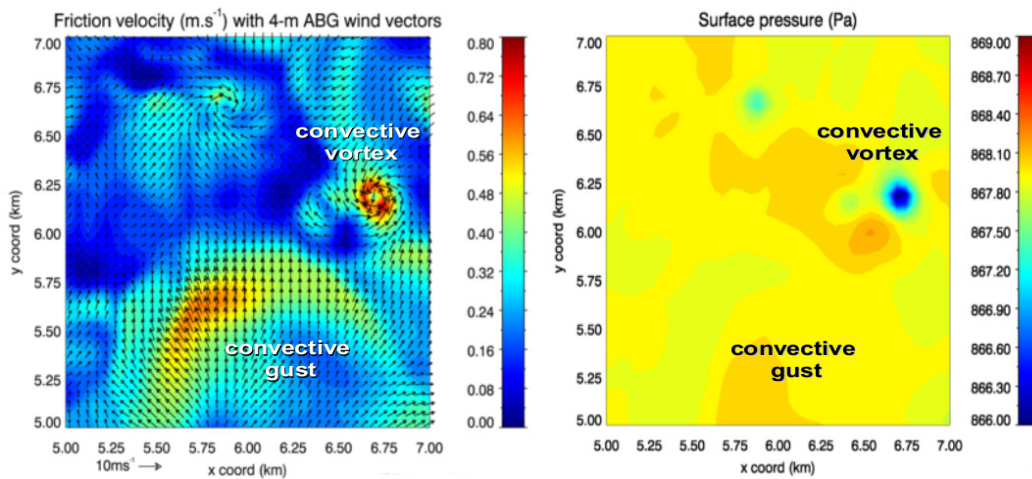


Figure 5. Horizontal map of (left) friction velocity (with superimposed wind vectors 4m above the surface) and (right) surface pressure around local time 13h30 in a chosen area of the large-eddy simulation domain where maximum values for u^* are detected.

compared to lower plains such as Amazonis Planitia. This could explain the presence of dust devils at high surface altimetry on Mars (*e.g.*, over the caldera or slopes of Arsia Mons, as shown in Reiss et al. 2009), while lower density / pressure would normally not favour the lifting of dust from the surface compared to lower surface elevations. Since the convective boundary layer is also deeper over these terrains, dust devils are higher (up to boundary layer depth, *i.e.*, 8 km here) which means they could inject particles up to 10 km above the zero datum reference altitude. This value, calculated for Tharsis plateaus, is far from being an upper limit: dust devils forming over the Arsia caldera as reported by Reiss et al. (2009) could potentially lift and transport dust up to 25-30 km above zero datum.

Getting high values of u^* within convective vortices is only one factor indicating possible dust lifting and dust devil formation. Maps of dust devils occurrence retrieved by MOC imagery confirm that Tharsis and Solis low-pressure terrains are preferential areas for dust devil formation, but high-pressure terrains such as Amazonis or Hellas are also key regions for dust devils formation (Cantor et al. 2006). The next step is thus to include dust lifting and transport by the convective gusts and vortices computed by the turbulent-resolving simulations to gain further insights into the formation of dust devils – and not only convective vortices which are shown by LESs to occur almost anywhere on Mars. The sensitivity of boundary layer circulations and dust lifting to surface properties (roughness, thermal inertia) and background winds needs to be determined, as well as possible radiative feedbacks of lifted dust into dust devils (*e.g.*, Fuerstenau 2006). It would also be useful to further investigate whether dust lifting could be enhanced by low-pressure core or triboelectric effects within an existing dust devil. The contribution of dust devils to the total amount of dust in the atmosphere remains an important issue to address in the Martian environment. Dust devils might be important contributors to maintaining a continuous component of background dust loading in the atmosphere, even in the

absence of larger dust storms (as others have previously noted, *e.g.*, Balme and Greeley 2006).

Conclusions

We have investigated meteorological scales between 100 km and 10 km using both the high-resolution UK GCM and the LMD mesoscale model (forced by the LMD GCM) with similar physical schemes. Where the models almost overlap in resolution and coverage (resolutions of 20-40 km), results seem reassuringly robust comparing the two models, despite the GCM being hydrostatic and the mesoscale model nonhydrostatic. Both show strong topographic control of the daytime and nighttime near-surface winds. The higher topographic slopes accessible in high resolution models are likely to be important for dust lifting, especially by anabatic and katabatic winds.

Scales below 10 km and 1 km, resolved through LESs, are dominated by turbulent gusts and dust devils, two distinct convective boundary layer processes likely to lift dust from the surface. Convective vortices have an enhanced ability to both lift and inject dust in the atmosphere. In low-latitude regions, boundary layer depth and friction velocity u^* are correlated with surface altimetry. This could explain the presence of dust devils at high surface elevations on Mars.

Directory of supporting data

[root directory](#)

[spiga_mars_2010_0006.pdf](#)

[LMD_MM_MARS_sol193_start6z.nc](#)

LMD mesoscale model results (NETCDF file)

[LMD_MM_MARS_ustar_LEScaseB.nc](#)

LMD mesoscale model Large-Eddy Simulations (NETCDF file)

[LMD_MM_MARS_ustar_LEScaseC.nc](#)

LMD mesoscale model Large-Eddy Simulations (NETCDF file)

[UK_MGCM_ustar.nc](#)

UK global circulation model results (NETCDF file)

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