TWO AERODYNAMIC ROUGHNESS MAPS DERIVED FROM MOLA DATA AND THEIR EFFECTS ON BOUNDARY LAYER PROPERTIES IN A MARS GCM. N.G. Heavens¹, M.I. Richardson², and A.D. Toigo² ¹Division of Geological and Planetary Sciences, California Institute of Technology, MC 150-21, Pasadena, CA, 91125 (heavens@gps.caltech.edu, mir@gps.caltech.edu); Center for Radiophysics and Space Research, Cornell University, Ithaca, NY 14853 (toigo@astro.cornell.edu).

Introduction: An essential characteristic of a lower atmospheric model for a terrestrial planet is a scheme to describe how heat and momentum are exchanged between the surface and the atmosphere. In a General Circulation Model (GCM), these exchange processes are often modeled by elaborations on the boundary layer turbulence theory first described by Obukhov [1]. One familiar result of this theory is the logarithmic profile of wind speed with height in a neutrally stratified layer from a level of no motion near the surface. The height of this level of no motion is alternately called “the roughness parameter” or “the aerodynamic roughness height/length” but is typically denoted $z_0$. $z_0$ is usually treated as a spatially-varying intrinsic property of the surface, corresponding to the effects of surface roughness elements such as rocks and trees on the atmospheric flow.

An accurate map of $z_0$ is extremely useful for certain types of GCM simulations. Because of $z_0$’s importance in boundary layer theory, the representation of many boundary layer processes within a GCM are highly sensitive to it, e.g. turbulent eddy diffusion of heat and momentum. If turbulent eddy diffusion is vigorous (high $z_0$), surface heat will be transported efficiently to the rest of the boundary layer and vice versa. However, low $z_0$ may make the boundary layer more prone to convect, enhancing vertical heat transport over smooth areas in some cases. Thus boundary layer atmospheric temperatures may be non-linearly sensitive to $z_0$. Of interest for both the Earth and Mars is the effect of $z_0$ on the transfer of momentum from the wind to sand and dust-sized potential aerosols on the surface. Smoother areas are more resistant to erosion and are less likely to be sites of initiation for dust lifting activity such as dust storms.

Remote sensing provides a variety of avenues for investigating the roughness characteristics of Mars. However, the transfer function between the physical roughness characteristics that interest geologists and $z_0$ is not always straightforward. In this abstract, we first will describe two ways of conceptualizing an important non-linearity in the transfer function. Next, we construct two different maps of aerodynamic roughness length based on Mars Orbital Laser Altimeter (MOLA) data, one based on topography and the other based on the scattering of the MOLA beam by the surface. Finally, we will use each roughness map to force a Mars GCM, the Mars implementation of the Planetary Weather Research and Forecasting Model (MarsWRF) and examine differences in their output boundary layer temperatures. We also consider the implications of the differences between the maps for dust devil activity.

The Effects of Roughness Element Spacing on $z_0$: In wind tunnel experiments, $z_0$ is typically a fraction of the height of the roughness elements. Moreover the magnitude of this fraction is variable, ranging from ~0.01-0.1 [2]. Some of this non-linearity is due to the fact that $z_0$ is not really an intrinsic characteristic of the surface but also depends on the turbulence of the flow. Additional non-linearity results from the distribution of roughness elements (Figure 1).

![Figure 1: A schematic showing how variations in the spacing of roughness elements along a line affects $z_0$ (after Greeley and Iversen [2]).](3218.pdf)

Note that this one-dimensional approach to the effect of roughness element distribution on $z_0$ suggests the relation $z_0 \propto |dh/dx|$. Actual roughness element distributions, however, typically are two dimensional. As a thought experiment, let us take a statistical approach to the two-dimensional problem. Consider an area with a number of sites, $N_s$. Some number of these sites, $N_0$, are occupied by a roughness element of height, $h$. We make no assumption about how uniformly distributed the roughness elements are, but it is helpful to pretend that all the roughness elements are not clumped all to one side. We then define $F$ as the fractional occupation of sites, $N_0/N_s$. It can be shown that the topographic variance of the area, $\sigma_h^2 = h^2(F-F^2)$ and the standard deviation $\sigma_h = h(F-F^2)^{1/2}$ (Figure 2). Wind tunnel experiments show that $z_0(F)$ typically has a form very similar to $\sigma_h$ or $\sigma_h^2$, though some skewing of the maximum from $F=0.5$ is observed [3]. Thus, the statistical properties of physical roughness may have a simple relation to $z_0$. 

Figure 2: Variation in the variance (blue) and standard deviation (red) of topography with fractional occupation of roughness elements of h=1.

A Topographically-Derived Aerodynamic Roughness Map for Mars: Krevlasky and Head have published maps at 0.125°x0.125° resolution of a topographic roughness parameter derived from MOLA measurements called C [4,5]. C is in the interquartile width of the distribution within each cell of parameter c, which is equal to (h_{i+1} + h_{i-1} + 2h_i)/4L^2, where h is the height above the MOLA datum and i, i+1, and i-1 represent the center and ends of a baseline L. Krevlasky and Head’s publicly available maps show roughness at L=0.6, 2.4, and 9.2 km. Note that the definition of C implies a relation between C and σ_h.

We then assume that the fractal self-affinity of roughness at meter to kilometer scales such that σ_h^2 = aL^2H, where a is a proportionality constant and H is the Hurst exponent, which has a range of (0,1) for natural surfaces [6]. If we assume that c within a given grid box follows a normal distribution, we then can use the fact that the interquartile-width =1.38 σ_h. If we next assume that h_{bas} = h_{bar} - h_{bar, i} - h_{bar, i}, i.e., the mean of all possible heights used to calculate h is the mean height measured in the grid box, we can derive a complex statement of C:

\[
C = \frac{1.38}{4L^2} \left[ \left( \sigma_{i+1}^2 + 2\sigma_{i+1} \sigma_{i-1} r_{i+1,i} - 4\sigma_{i+1} \sigma_{i-1} r_{i+1,i} \right)^{1/2} + \sigma_i^2 + 4\sigma_i^2 \right]
\]

where σ_i^2 = variance of h_i and r_{i,i+1} = correlation between h_i and h_{i+1}. C is maximized when the heights of the ends of the baseline are correlated and the heights of the ends anti-correlate with the height of the baseline center. C is minimized when the heights of the baseline ends and the center all correlate perfectly, i.e., flat terrain. Provided that σ_h is invariant within the grid box, C^max = (1.38/4L^2)σ_h and C^min = 0, implying that C^bar = (1.38/2L^2)σ_h. We expect that σ_h^2 = C^2L^4 within a constant of proportionality. However, when we rearrange C^2L^4 = aL^{2H} in a form suitable for linear regression and fit each point on Krevlasky and Head’s roughness maps, we generally obtain values of H in the range (0,2), implying that σ_h^2 = CL^2, not C^2L^4. Using the parameters a and H for each point (assuming σ_h^2 = CL^2), we then extrapolate C to a baseline of 0.01 km. (10 m.). This baseline corresponds to the approximate depth of mechanical turbulence and thus the approximate scale of interactions between roughness elements and near-surface atmospheric flows. We still do not know why our theoretical derivation of σ_h^2 disagrees with the empirical result, but it may stem from our assumption of a normal distribution and our assumption that C=C^bar.

Figure 3: Distribution of LOG(C(10 m.)), based on randomly extracted C at 50000 grid points, showing the range of C(10 m.) around the Mars Pathfinder site. The outlier is an artifact associated with very high-latitude data.

The distribution of C(10 m.) spans ~2.5 orders of magnitude (Figure 3). Sullivan et al. expect that z_0 should range from ~1 mm. to 10 cm [7]. This is a somewhat narrower range than observed on Earth. Terrestrial values in poorly vegetated terrain range from ~50 μm. to >15 cm. [8,9]. Therefore, if z_0 is proportionate to σ_h (same as proportionality to C^{1/2}), we likely obtain an overly narrow range for Martian roughness (1.25 orders of magnitude). Moreover, by eye, the relationship between z_0 and F in experiments typically does not bulge upward to the extent seen in the σ_h curve in Figure 2. Thus, we will assume that z_0 is proportionate to σ_h^2 and thus to C. We then calibrate C(10 m.) to obtain z_0 by using the inference from Pathfinder wind measurements that z_0 at the Pathfinder site is ~ 3 cm,
thought to be an unusually high value for plains units [7]. An interpolation of this data to a 36x64 grid is shown in Figure 4(a).

Figure 4: (a) \( \log_{10}(z_0) \) (m) inferred from C(10 m.) interpolated to the standard 36x64 MarsWRF grid; (b) \( \log_{10}(z_0) \) (m) inferred from MOLA optical pulse width rms slope-corrected roughness interpolated to the standard 36x64 MarsWRF grid; the color stretch is identical for both figures.

A MOLA Pulse Width-Derived Aerodynamic Roughness Map: We obtained RMS total vertical roughness maps for Mars at 0.25°x0.25° resolution from J. Garvin. Total vertical roughness is inferred from slope-corrected MOLA optical pulse width returns as described by Garvin et al. [10]. The approximate baseline of these measurements is 160 m. Some corrections to this method of inferring total vertical roughness were made by Neumann et al. [11], but we do not use them here. We then arbitrarily chose 15 cm. to be the maximum \( z_0 \), corresponding to an RMS roughness of 11 m. Note that the range of \( z_0 \) is roughly an order of magnitude (~1-10 cm.) with typical values ~2 cm. \( z_0 \) in the vicinity of the Mars Pathfinder site ranges from ~1.6-3.8 cm. in this map. An interpolation of the pulse-width \( z_0 \) to a 36x64 grid is shown in Figure 4(b).

Differences Between Roughness Maps and Implications for Dust Devil Activity: There are definite similarities between the two maps of \( z_0 \) such as the high \( z_0 \) associated with the lava flows surrounding Olympus Mons, the Tharsis Montes, and Aeolis Mensae. Moreover, both maps generally show smoother terrains in the high latitudes of both the northern and southern hemispheres. The major difference between the two maps is the ranges of \( z_0 \) they show, which are due to different assumptions about the relation between \( z_0 \) and some measure of \( \sigma_h \). The topographically-derived map assumes \( z_0 \propto \sigma_h^2 \), while the pulse width map assumes \( z_0 \propto \sigma_h \). If one of these maps were a better representation of the true \( z_0 \) of Mars, there not only might be implications for boundary layer temperatures but also for dust devil activity. In Heavens et al. [12], we investigate the possibility that the number density of dust devils per unit area, \( N_{DD} \), may be proportionate to \( \exp(-D_{Dec}/(h/|L|)) \), but the fraction of dust devils larger than some diameter \( D_{DD} \) may be proportionate to \( \exp(-D_{DD}/(2|L|)) \), where \( L \) is the Obukhov length (a strong function of \( z_0 \)), \( h \) is the convective boundary layer height, and \( D_{De} \) is some critical \( h/|L| \) for dust devil formation. Thus, it is possible that the high density of dust devil tracks observed at high latitudes [13] could be due to the low values of \( z_0 \) there, and indeed this effect would be enhanced for \( z_0 \propto \sigma_h^2 \). Very smooth terrain, however, might limit erosion by dust devils, producing a contrary effect. Roughness-related effects also might explain the high density of large dust devils observed on NW Amazonis Planitia but the relatively low density of small dust devil tracks observed there [14, 15]. Note that the area of Amazonis repeatedly observed by MGS MOC corresponds to the unusually rough “teardrop” terrain [11] (~36, -160), clearer in the topographically-derived map. From our assumptions about the relations of \( N_{DD} \) and \( D_{DD} \) to \( |L| \), it is quite simple to derive \( |L|^b_{WA} \), the optimal Obukhov length for the formation of of dust devils resolved by MOC Wide Angle imagery:

\[
|L|^b_{WA} = \sqrt{\frac{R_{DDWA} h}{D_{De}}} \tag{2}
\]

where \( R_{DDWA} \) is the minimum radius of a dust devil resolved in MOC WA imagery (~250 m.), \( h \) is typically ~10000 m. and \( D_{De} \) is ~1000, so \( |L|^b_{WA} \) is ~50 m. A factor of 2 difference in \( z_0 \) is roughly equivalent to a
factor of 2 change in |L|, so a factor of 2 change in z₀ will produce an order of magnitude lower NDD observable in MOC WA imagery. NW Amazonis Planitia thus may be a rough “sweet spot” for large dust devils with [L−|L|]_{WA} as opposed to a site of intense dust devil activity at all scales.

**Boundary Layer Temperature Comparisons with MarsWRF:** We ran MarsWRF simulations using the two different roughness maps, in order to test the effects of each roughness map on boundary layer temperatures. The simulations use a passive dust forcing based on MGS TES temperature data [16], so interannual variability is limited. All data presented here is from year 8 of the simulations. We first assess zonally-averaged daytime temperatures throughout the year at 3.7 mbar as in [17] (Figure 5a). Daytime temperatures at 3.7 mbar were derived from the maximum daily potential temperature at the model level with the mean pressure closest to 3.7 mbar during that time of year. Figures 5b and 5c show differences between the simulations in zonally-averaged boundary layer temperatures. The main effect of z₀ on boundary layer temperatures appears to come from eddy diffusion. Latitudes that are smoother on the C(10 m.) map tend to have cooler boundary layer temperatures and *vice versa.* This effect might be weak at night because of the negative surface heat flux. A slightly different effect is seen at the poles, especially at the south pole, which is much smoother on the C(10 m.) map than the pulse width map. The mean warming in the C(10 m.) forced simulation at the south pole may result from the slowing of sublimation in spring and summer due to weak eddy diffusion. Over the course of the year, the magnitude of these temperature differences are small (<1 K), but they may be larger seasonally (5-20 K).


**Figure 5:** (a) Difference over the course of the year between zonally-averaged daytime temperatures (K) at 3.7 mbar in simulation year 8 with C(10 m.) derived roughness forcing and pulse width roughness forcing; (b) Difference between latitudinally averaged daytime boundary layer temperatures (K) in year 8 of the two simulations. The abscissa is approximate height of the model layer above the surface; (c) Difference between latitudinally averaged nighttime boundary layer temperatures (K) in year 8 of the two simulations. The abscissa is approximate height of the model layer above the surface.