Hydrogen content of sand dunes within Olympia Undae

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Abstract

Neutron currents measured using the Mars Odyssey Neutron Spectrometer, seasonally varying temperatures measured using the Thermal Emission Spectrometer, and visible images measured using the High Resolution Imaging Science Experiment (HiRISE) are studied to determine the water content and stratigraphy of Olympia Undae. Both the neutron and thermal infrared data are best represented by a two-layered model having a water–ice equivalent hydrogen content of 30 ± 5\% in a lower semi-infinite layer, buried beneath a relatively desiccated upper layer that is 9 ± 6 g/cm\textsuperscript{2} thick (about 6 cm depth at a density of 1.5 g/cm\textsuperscript{3}). A model that is consistent with all three data sets is that the dunes contain a top layer that is relatively mobile, which overlays a niveo-aeolian lower layer. The geomorphology shown by the HiRISE images suggests that the bottom layer may be cemented in place and therefore relatively immobile.

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1. Introduction

The most extensive sand dune deposit on Mars encircles the north-polar residual water–ice deposit. It has its largest contiguous areal extent between 155\° and 230\° E and 78\° and 83\° N, which has been recently renamed Olympia Undae (Tanaka et al., 2005). Although the dunes have been studied extensively using visible and infrared imaging data of the Viking orbiters (e.g., Cutts et al., 1976; Tsoar et al., 1979; Ward and Doyle, 1983; Thomas, 1987; Lancaster and Greeley, 1990), many fundamental issues regarding the origin, evolution, and internal structure of the dunes remain unknown. For example, the apparent thermal inertia of the dunes is modest to low (Paige et al., 1994) suggesting that they are made up of irregularly-shaped cemented dust and sand fragments (Greeley, 1986) or mixtures of ice and silicate dust (Saunders et al., 1985; Saunders and Blewett, 1987) that are not cemented in bulk. Alternatively, the erg is thought to be predominantly composed of sand sized andesite or weathered-basalt fragments (Edgett and Christensen, 1991; Bandfield, 2002; Edgett et al., 2003). In addition, the east end of Olympia Undae has a strong gypsum signature in OMEGA (Observatoire pour la Mineralogie, l’Eau, les Glaces et l’Ativité) data (Langevin et al., 2005). The highest concentration of gypsum is located upwind of the dune field, extends across the dunes, decreasing in concentration downwind, suggesting deposition via an aeolian dust plume. Alternate sources of gypsum signature detected in the dunefield are thought to be from the polar-layered deposits’ basal unit, exposed between the dunes and along the ice cap edge. However, this source is not supported in the OMEGA or CRISM (Compact Reconnaissance Imaging Spectrometer for Mars) data analysis (Langevin et al., 2005; Roach et al., 2007).

All the foregoing models suggest that the surface sediment of these dunes should be mobile, and recent findings indicate
significant sediment transport is possible from small dunes in the north polar region (Bourke et al., 2008a). However, it does appear that most polar dunes are relatively immobile and may be stabilized by surface crusting (Schatz et al., 2006) or cemented by niveo-aolian deposits (Bourke, 2004).

In order to help resolve some of these issues, we report here on the results of a study of the hydrogen content of the north-polar sand dunes within Olympia Undae. Thermal and epithermal neutron currents measured using the Mars Odyssey Neutron Spectrometer (MONS) are analyzed for this purpose. These data yield an estimate of the water-equivalent hydrogen (WEH) mass fraction in near surface material, which can provide limits on the degree of ice content given an assumption of dune mineralogy and porosity.

Although the Olympia Undae Formation is large (∼300 × 800 km, which corresponds to the central contiguous portion of the dune fields in the geologic map of Tanaka, 2005) it is smaller than the MONS instrument footprint. It can therefore not be well characterized by the MONS, which has an intrinsic Full Width at Half Maximum spatial resolution of about 600 km. However, a spatially deconvoluted version of the MONS data may attain a resolution of 300 km. We present here our preliminary deconvolved maps of the water content and stratigraphy of Olympia Undae for this purpose and interpret them in terms of potential water–ice content and layering. The resultant layering is checked using an analysis of the seasonal variation in temperature of the dune field using the Thermal Emission Spectrometer (TES) (Christensen et al., 2001; Bandfield, 2007).

2. Neutron data reduction procedure

Thermal and epithermal neutron currents measured using MONS between February 2002 and December 2006 were analyzed in multiple ways to determine limits on the WEH mass fraction of Olympia Undae dune material and the potential layering of these deposits. Results returned using two different data-reduction algorithms (Prettyman et al., 2004; Maurice et al., 2007) were inter-compared and then averaged in order to provide an estimate of systematic errors resulting from this part of the analysis procedure. Next, we improved the spatial resolution of resultant WEH maps using two different deconvolution techniques in order to place limits on the modifications of the counting-rate data as measured, that are produced by application of these procedures. Our final products consist of lower bounds of the water content of these dunes by applying: (1) a single-layered model of surface soils using the measured epithermal neutron currents alone, and (2) a two-layered model using both the measured thermal and epithermal neutron currents. As shown in Fig. 1, the single-layered model is a special case of the two-layered model. Whereas the burial depth, D, of a water-rich permafrost layer is chosen to be zero in the single layered model, it is nonzero in the two-layered model. In both models, the formation is assumed to extend uniformly to the edge of the MONS field of view. In order to constrain this model using only two measured quantities (the thermal and epithermal neutron currents), all measured data have been corrected for a constant 16 g/cm² atmosphere, the water mass fraction of the upper layer is chosen to be 0.01, and the elemental composition of both soil layers is chosen to be an average of that measured at the Viking 1, Pathfinder, and Mars Exploration rover (at both Meridiani and Gusev) landing sites (see Diez et al., 2007; Feldman et al., 2007 and references therein).

The first deconvolution technique used here is very similar to the van Cittert deconvolution method (Van Cittert, 1931; Jansson, 1997), $I_{k+1} = I_k + r(O - p \otimes I_k)$, where $I_{k+1}$ is the current estimate of the restored image, $I_k$ is the previous estimate, $r$ is a relaxation function, $O$ is the original smoothed image, $p$ is the total effective point spread function (equivalent to the Gaussian smoothed MONS response function) and $\otimes$ denotes a convolution operation. The iteration procedure was terminated just before the water content of the residual polar cap exceeded 100%. Whereas the epithermal neutron map was deconvolved using the response function of the downward-looking prism of MONS by itself (see Feldman et al., 2002 and Boynton et al., 2004 for a description of the instrument), the thermal neutron map was deconvolved in two steps. The first step involved deconvolving the response functions of the forward- and backward-looking prisms separately. The second step reconstructed the deconvolved thermal neutron map by taking the difference between the separately deconvolved forward and backward facing prism counting rates.

The second technique used the Pixon method developed by Puettter (1995) and explored by Lawrence et al. (2007) using the Th map of the Moon measured using the Lunar Prospector Gamma-Ray Spectrometer. In contrast to the Jansson method that uses a fixed grid of measurements, the Pixon technique is a spatially adaptive, image restriction method. This method returns the smoothest possible image that is consistent with the uncertainties in the data. This goal is reached by varying its mesh size to match the information content of an image relative to its noise uncertainties (Puettter, 1995) and is applied to the neutron counting rates in all three prisms separately.

![Geometry of the two-layer model of martian surface used to analyze thermal and epithermal neutron currents measured using MONS.](image-url)
3. Results

Stereographic projections of the WEH content of surface layers at high northern latitudes are shown in Fig. 2. Single-layered WEH mass fractions superimposed on a shaded relief map measured using Mars Orbiter Laser Altimeter (MOLA) (Smith et al., 2001) poleward of 45° N are shown in Fig. 2a. Inspection shows a water–ice residual cap that is slightly offset from the north pole toward 0° longitude. A secondary maximum in WEH covers the Scandia Colles formation (Tanaka et al., 2005) centered at about 70° N and −130° E. At our spatial resolution, this formation is connected to an arc of WEH-rich terrain that extends at constant latitude to about 150° E. Between this arc and the residual water–ice cap is terrain that has a relative minimum in WEH content co-located with the Olympia Undae formation. The minimum WEH mass fraction in the one-layered model amounts to $WEH_{\text{min}} = 0.272 \pm 0.025$ at 81° N and −171° E.

In order to place this position in perspective, it is located by an 'X' in the geologic (Tanaka et al., 2005) and elevation (Zuber et al., 1998) maps of northern high latitudes, shown in Figs. 3a and 3b, respectively. This minimum is seen to be near the azimuthal center of the Olympia Undae formation (Fig. 3a) but offset towards its equatorial edge in latitude. It is also seen to be on the sloping margin of the sand sea, slanted towards the equator, shown in Fig. 3b.

Fig. 2b gives the WEH mass fraction of the semi-infinite lower layer, $W_d$, using a two-layered model (Diez et al., 2007; Feldman et al., 2007). The top layer is assumed to have a WEH mass fraction of 0.01. Although both single-layer and two-layer models give the same overall lateral structure of the WEH deposits, the two-layer model gives a slightly higher minimum WEH mass fraction, $W_d_{\text{min}} = 0.35 \pm 0.03$.

Fig. 2c gives the burial depth of the bottom, WEH-rich layer, in g/cm². Inspection shows a ring centered at 60° N of apparent maximum burial depth (Feldman et al., 2007) that encircles a surface deposit of water ice that covers the north pole. The minimum WEH deposit that covers the sand dune formation in Olympia Undae corresponds to 7 ± 3 g/cm², which is at 4.7 cm depth if the density is 1.5 g/cm³.

The results of the Pixon deconvolution procedure are shown in Fig. 4. Meridional transects of measured epithermal neutron counting rates and the WEH content derived from the one-layer model through the center of the deposit at −171° E, are shown in Fig. 4a. A relative maximum counting rate occurs at about 80° N, which corresponds to a minimum WEH mass fraction of 0.20. Maps showing the deconvolved spatial distribution of counting rates and WEH abundances are shown in Figs. 4b and 4c, respectively. Both these maps and the transects can only be considered approximate at this time because a study of the residual counting rates between the measured rates and the convolved model rates do not appear to be random (they appear to contain clumps, see Lawrence et al. (2007) for a discussion of residuals in the context of spatial deconvolution). Although we do not know what the origin of this result is, we note that the difference in minimum WEH abundance derived from the two deconvolution techniques (Jansson and Pixon) amount to 0.07 with an average value of 0.235.

Quantitative results of all determinations of WEH mass fractions and burial depths at the spatial minimum located at 81° N and −171° E are collected in Table 1. Also included is an estimate of the open pore volume, $PV$, between soil grains that would be needed to hold the minimum water mass fractions of WEH or Wdn. The density of interstitial water ice, $\rho_I$, was assumed to be 0.92 g/cm³ and that of the sand grains, $\rho_s$, was 2.65 g/cm³ giving

$$PV = (\rho_s Wdn) / (\rho_I + (\rho_s - \rho_I) Wdn).$$

4. Discussion

Our best fitting neutron model of Olympia Undae requires at the least, a two-layered structure in order to fit both the measured thermal and epithermal neutron currents. The topmost
layer is relatively desiccated, chosen here to contain 0.01 by mass of water-equivalent hydrogen, has an average thickness of $7 \pm 3 \text{ g/cm}^2$, and overlays a semi-infinite lower layer that contains $0.295 \pm 0.05$ by mass WEH.

The seasonal variation in surface temperatures measured by the TES spectrometer are also consistent with a layered surface. Ice-cemented materials will have a higher thermal inertia than poorly consolidated sand or soil. At high latitudes, both the diurnal and seasonal temperature variations are prominent. For example, Titus et al. (2003) and Armstrong et al. (2005) used these combined cycles apparent in seasonal day/night temperature data to distinguish layered and nonlayered materials. Bandfield (2007) also used seasonal temperatures to predict water ice depths, but without the use of the daytime (low angles of solar incidence) temperatures. The daytime temperatures were not used because of the dominant influence of slopes, surface albedo, and atmospheric properties and their associated uncertainties on the derivation of surface thermophysical properties. This analysis relies on the variable seasonal influence on nighttime surface temperatures during the frost-free summer season to predict water ice burial depths.

TES spectrometer-derived surface temperatures acquired at a local time of 0200–0300 from 79.5° to 80.5° N and −165° to −175° E were averaged into 4.5° bins of $L_s$. The data included all three martian northern summers when the TES spectrometer was operating and emission angles were restricted to less than 20°. Using a two-layered thermal model (developed and provided by H.H. Kieffer and used in Titus et al., 2003; Armstrong et al., 2005; Fergason et al., 2006a, 2006b; Bandfield, 2007), the seasonal TES temperatures were fit using a nonlinear least-squares fitting routine (model parameters are listed in Table 2). The data were fit between $L_s$ of 85° and 170° (avoiding seasonal CO$_2$ frost) allowing top layer thermal inertia and depth to a water–ice rich high inertia layer to vary as free parameters. The best-fit model determined a top layer thermal inertia of 280 J m$^{-2}$ K$^{-1}$ s$^{-1/2}$ with the ice-bearing layer at 7.5 cm depth (Fig. 5). The fit is considerably worse (RMS errors of 6.4 K versus 2.8 K) with the absence of a high inertia layer at depth. The determination of depth is not precise, however. Top layer inertias and water ice depths can be changed to 200 J m$^{-2}$ K$^{-1}$ s$^{-1/2}$ and 3 cm or 340 J m$^{-2}$ K$^{-1}$ s$^{-1/2}$ and 11.6 cm without significantly increasing the RMS error (<0.5 K; see Fig. 5). This range is well within the water ice table depth predicted by the MONS determination. Previous work assuming a nonlayered surface and using Viking Infrared Thermal Mapper (IRTM) data determined thermal inertia values of 200 to 350 J m$^{-2}$ K$^{-1}$ s$^{-1/2}$ in the Olympia Undae region near 80° N, −170° E, consistent with the top layer inertia results shown here (Paige et al., 1994).

In addition, the model assumes a level surface, when in fact the dune surfaces have high slopes of a variety of azimuths, adding a potential surface temperature bias. To gain an estimate of the upper magnitude of these effects, the modeled radiance of equal proportions of north/south/east/west oriented 30° slopes were averaged and converted to apparent brightness temperatures at 25 µm (approximately the same wavelength used for
Fig. 4. (a) Meridional transects of measured epithermal neutron counting rates and the WEH content derived from the one-layer model through the center of the deposit at $-171^\circ$ E. Maps showing the deconvolved spatial distribution of counting rates (b) and WEH abundances (c). The slanted dashed line in (b) gives the location of the transect.

Table 1

<table>
<thead>
<tr>
<th>Model</th>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>One layer, as measured</td>
<td>WEH mass fraction</td>
<td>0.272 ± 0.025</td>
</tr>
<tr>
<td>One layer, deconvolved</td>
<td>WEH mass fraction</td>
<td>0.235 ± 0.035</td>
</tr>
<tr>
<td>Two layer, as measured</td>
<td>WEH lower layer</td>
<td>0.35 ± 0.030</td>
</tr>
<tr>
<td>Two layer, as measured</td>
<td>Burial depth</td>
<td>7 ± 3 g/cm$^2$</td>
</tr>
<tr>
<td>Two layer, deconvolved</td>
<td>WEH lower layer</td>
<td>0.295 ± 0.05</td>
</tr>
<tr>
<td>Two layer, deconvolved</td>
<td>Burial depth</td>
<td>6 ± 3 g/cm$^2$</td>
</tr>
<tr>
<td>One layer, as measured</td>
<td>Pore volume</td>
<td>51.5 ± 3.5%</td>
</tr>
<tr>
<td>Two layer, deconvolved</td>
<td>Pore volume</td>
<td>54.5 ± 5.7%</td>
</tr>
</tbody>
</table>

TES data surface temperature determinations). This distribution of slopes resulted in a slightly shallower water ice (6.5 versus 7.5 cm) and lower top layer thermal inertia (240 versus 280 J m$^{-2}$ K$^{-1}$ s$^{-1/2}$) in the best fit model. While slopes may have little effect on the precision of the surface temperature modeling, they almost certainly influence the local depth of the water ice table (e.g., Sizemore and Mellon, 2006; Aharonson and Schorghofer, 2006). South facing slopes receive the greatest amount of annual insolation and north facing slopes receive the least, resulting in relatively deep and shallow water ice stability depths, respectively. It should be assumed that the modeled water ice table depths represent an intermediate value within a heterogeneous region.

In support for our conclusion that a single-layered model is not sufficiently complex to fit the neutron and thermal data, newly acquired images of the sand dunes within Olympia Undae using High Resolution Imaging Science Experiment (HiRISE) show a rather complicated geomorphology, as shown in Fig. 6. The dunes form transverse linear arrays separated by apparently flat basement material, which in patches has a rel-
in polar deserts on Earth and consist of inter-bedded sand and ice that are often indurated and resistant to erosion. On Earth, niveo-aeolian deposition occurs due to the burial of precipitated snow and frost layers by rapidly aggrading sand. Alternatively, snow can be transported simultaneously with sedimentary particles and deposited on the dune. For martian aeolian landforms, the definition has been extended to include diffusion of vapor molecules from the atmosphere to fill sub-surface pore volumes (Bourke, 2004). Coupled with the occurrence of niveo-aeolian deposits is the potential for melting (perhaps during previous high obliquity cycles), evaporation and sublimation of volatiles that are located close to the surface. This phase change should be associated with morphological changes that are visible on the surface of the dunes. Indeed, we have observed geomorphic features on the dune and interdune surfaces that suggest subsurface volatiles are, or have been present.

For example, circular depressions are located periodically along the dune crest (Fig. 8b). These are between 5 and 7 m in diameter and have estimated depths (using shadow length) of 0.7 m. Additional elongated depressions (Fig. 8c), also located along the dune crest, have rounded end planforms and are proposed to form by the coalescence of individual circular depressions. These forms occur frequently along the dune crests and modify significant lengths of dune crest morphology (Fig. 8a). These features may be similar to sinkholes found on polar desert dunes on Earth (Koster and Dijkmans, 1988). Terrestrial dune sinkholes are formed by the loss in volume of underlying volatiles. These features are found elsewhere in the martian polar deserts. Their position along the dune crest and not elsewhere may be linked to larger diurnal insolation receipts at that location on the dune. These pits, then, identify the locations of denivation, but do not necessarily map out the potentially wider distribution of sub-surface volatiles.

Sinuous, narrow, rectilinear and branching depressions, measuring 25 to 70 m long and <1 m wide, are observed on the lee face of the larger dunes and on both flank slopes of the smaller orthogonal dunes (Fig. 9). They truncate the ubiquitous small-scale transverse ripple features (λ = 2.5 m, with estimated heights of <0.5 m) on the dune surface. These features suggest cohesion of the surface sediment and the operation of tensional stresses. They are similar to tensional cracks reported on dune surfaces in polar deserts. On Earth, the cohesion is a function of moisture content (from snowmelt) and the source of tensional stress is a loss of subsurface volatile volume (Koster and Dijkmans, 1988). On Mars, cohesion of the surface sediment may be provided by crusting (Sullivan et al., 2004; Richter et al., 2006). These cracks, along with the absence of active dry grain flow on the dune avalanche face, suggest that sand transport has not been active recently. High-albedo surfaces exposed in the interdune area have a number of associated morphologies. High albedo arcuate ridges trending in the direction of the larger dunes may be indurated remnants of barchan dune strata that remained in situ as the dunes migrated downwind during an earlier dune-building phase. Similar features are reported from the permafrosted dunes in Victoria Valley Antarctica (Morris et al., 1972), and from the chemically indurated dunes of White Sands National Park, New Mexico.

<table>
<thead>
<tr>
<th>Layer properties</th>
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<tr>
<td>Density of top layer</td>
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<tr>
<td>Density of bottom layer</td>
<td>2018 kg m$^{-3}$</td>
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<tr>
<td>Specific heat of bottom layer</td>
<td>1040 J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>Specific heat of top layer</td>
<td>837 J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>Thermal conductivity of bottom layer</td>
<td>2.5 W m$^{-1}$ K$^{-1}$</td>
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<table>
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<tr>
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<td>First layer thickness (skin depth)</td>
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<td>Succeeding layer thickness ratio</td>
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<tr>
<td>Number of layers</td>
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<tr>
<td>Model iterations before output (80 iter./yr)</td>
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</tbody>
</table>

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<td>Surface albedo</td>
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</tr>
<tr>
<td>Surface emissivity</td>
<td>0.98</td>
</tr>
<tr>
<td>Visible albedo of CO$_2$ frost</td>
<td>0.65</td>
</tr>
<tr>
<td>Thermal infrared emissivity of CO$_2$ frost</td>
<td>1.0</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>Dust properties</th>
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</thead>
<tbody>
<tr>
<td>Dust opacity (visible wavelengths)</td>
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</tr>
<tr>
<td>Visible to 9 µm wavelength opacity ratio</td>
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<tr>
<td>Henyey–Greenstein asymmetry factor</td>
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<tr>
<td>Single scattering albedo</td>
<td>0.90</td>
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<tr>
<td>Twilight extension angle</td>
<td>1.0$^\circ$</td>
</tr>
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</table>

| Model iterations before output (80 iter./yr) | 160 |
Fig. 5. TES-derived and model seasonal surface temperatures at 80° N and −170° E. (a) TES measured temperatures (solid) compared with modeled temperatures with a top layer thermal inertia of 280 (MKS units) and ice at 7.5 cm depth (dashed). Dotted lines give an estimate of error in the TES measurements based on ±3 K at 150 K. (b) TES-measured temperatures (solid) compared with a variety of modeled temperatures: top layer inertia of 280 (MKS units) with ice at 7.5 cm depth (short dash); top layer inertia of 280 (MKS) with no ice present (long dash); top layer inertia of 200 (MKS) with ice at 3 cm depth (dotted); top layer inertia of 340 with ice at 11.5 cm depth (dash–dot).

(McKee, 1966). However the occurrence of these arcuate strata in Olympia Undae with polygonal terrain (Fig. 10b) combined with the modeled stability of ground ice at these northern latitudes, suggests that they are more likely to be exposed indurated permafrost layers rather than geochemically cemented strata.

The polygonal pattern of the high-albedo interdune strata observed in HiRISE images (Roach et al., 2007) suggests to us the presence of volatiles. In the lower part of Fig. 10b the polygons are bound by raised ridges, composed of sediment similar in albedo to the adjacent aeolian sand. This morphology is suggestive of permafrost and cryoturbation processes (Mangold, 2005).

Adjacent to the polygonal pattern is a patch of amorphous sand (Fig. 10b). This is striking because of the ubiquitous organization of the adjacent sand into ripple features and polygon ridges. The patch forms an irregular boundary with the ripples but does not appear to overlie them. We suggest that this morphology combined with its location suggests a loss of volatiles...
Hydrogen content of sand dunes

resulting in the collapse of polygon ridges (or indeed ripples) at this location. This morphology therefore suggests that the enrichment and loss of volatiles from the surface sediment is an active process on Mars and one that affects surface morphology.

Another potential component of the composition of Olympia Undae sand dunes is gypsum. The OMEGA experiment aboard Mars Express observed a concentration of gypsum at the eastern margin of Olympia Undae with decreasing concentrations westward across the dune field (Langevin et al., 2005). Subsequent analysis using CRISM data found a relatively stronger gypsum signature along the dune crests than in the inter-dune areas. This signature may be a result of a coarser grain size along the crest, cemented grains, or higher concentrations of gypsum (Roach et al., 2007). If this deposit (or in fact any other contaminant deposit that consists of hydratable minerals) has intercalated the sand grains then this would be reason by itself why a simple two-layered model should not apply. However, gypsum may not indicate a problem for the two-layer model because the material visible in CRISM data is confined to a thin mantle on top of the crests of the dunes and does not appear to be a significant contaminant of the macroscopic dune formation on a scale of the neutron field of view (Roach et al., 2007). However, its distribution at depth is presently unknown.

Nevertheless, if the sand grains are composed of sediments containing other hydratable minerals, then our estimated open pore volume will be lower. However, if the sand-composition model of mixtures of ice and silicate dust (Saunders et al., 1986; Saunders and Blewett, 1987) is adopted, and an open pore volume of sand before ice cementation is chosen to be 50%, then our estimated WEH content of a single semi-infinite sand deposit could fit the current paradigm of these sand dunes. For example, our results are consistent with a fully ice-filled pore volume at depth covered over by a relatively desiccated, loose sand cover. Such a structure would then have a relatively low thermal inertia, as determined using Viking Orbiter (Paige et al., 1994) and TES data, yet be relatively immobile because it would be fully cemented at depth. This suggests that frozen water (ice or snow) is present in the polar sand dunes on Mars and supports the assertion that some dunes in the north polar region are composed of niveo-aeolian deposits (Bourke, 2004).

The emplacement mechanism of these deposits is still under discussion but may be diffusion, precipitation or a combination of both.

Table 3
Summary of Olympia Undae morphometrics

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Large dune wavelength $\lambda_L$ (m)</td>
<td>470</td>
</tr>
<tr>
<td>Large dune width $W_L$ (m)</td>
<td>375</td>
</tr>
<tr>
<td>Large dune height *$H_L$ (m) (33° and 25°)</td>
<td>75 (100–75)</td>
</tr>
<tr>
<td>Area of substrate exposure (% of total area)</td>
<td>0.8</td>
</tr>
<tr>
<td>Small dune wavelength $\lambda_S$ (m)</td>
<td>230</td>
</tr>
<tr>
<td>Small dune width $W_S$ (m)</td>
<td>80</td>
</tr>
<tr>
<td>Fractional area of large dune $f_1 = (W_L/\lambda_L)$</td>
<td>0.8</td>
</tr>
<tr>
<td>Fractional area of small dune $f_2 = (W_S/\lambda_S)$</td>
<td>0.35</td>
</tr>
<tr>
<td>Fractional area of interdune $(1-f_1)(1-f_2)$</td>
<td>0.13</td>
</tr>
</tbody>
</table>

Notes. Dune morphometry at the sample site in Olympia Undae. However the presence of small ripples on the lee slope suggest that the dune flank does not lie at the angle of repose (assumed to be 33°). A lower angle is consistent with dune inactivity. We have therefore estimated dune height assuming an alternative slope of 25°. This lower estimate is likely more appropriate for dunes in this study area.

* Dune height was estimated using the slip face length method outlined in Bourke et al. (2006).
Fig. 8. Section of dune crest from HiRise image PSP_001736_2605. The Sun is shining from the left. (a) Black arrows indicate locations where the dune crest has been altered by hypothesized sublimation pit formation. (b) Two individual pits located on the crest. (c) An elongated depression 5-m wide and 27-m long formed by coalescing of several pits along a dune crest.

5. Summary and conclusions

Thermal and epithermal neutron currents measured using MONS, temperatures measured using TES, and geomorphic observations from HiRISE, were analyzed to determine the likelihood, average content, and stratigraphy of water ice within the sand dunes in Olympia Undae. The WEH mass fraction of surface soils is found to be a relative minimum at 81° N and −171° E. Both the neutron and thermal infrared data are best represented by a two-layered model having a water–ice equivalent hydrogen mass fraction of 0.295 ±0.05 in a lower semi-infinite layer, buried beneath an upper layer having 0.01 WEH mass fraction that is 9 ± 6 g/cm² thick (about 6 cm depth at a density of 1.5 g/cm³).

A model that is consistent with all three data sets is that the dunes contain a relatively desiccated top layer, which overlays a niveo-aeolian lower layer. The geomorphology shown by the HiRISE images suggests that the bottom layer may be cemented in place and therefore relatively immobile although permafrosted dunes in Antarctica have been shown to migrate in recent times (Bourke et al., 2008b). The composition of the lower layer may be: (a) an indurated, hydrated mineral such as gypsum, (b) pore-volume water ice emplaced by vapor diffusion, or (c) multiple layers of water ice and dust delivered as snow combined with sand during previous obliquity cycles.

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Fig. 10. Characteristics of the interdune area from HiRISE image PSP_001736_2605. The Sun is shining from the left. The top images shows narrow arcuate ridges (black arrows) that may be indurated stratigraphy of small transverse dunes. The locations of sub-panels (a) and (b) are indicated by black boxes. (a) Apparently indurated transverse aeolian ridges formed of high albedo material overlain by darker rippled sediment. (b) Bottom of image: Polygonal terrain with dark circumferal ridges. White arrow: Amorphous dark sediment amongst ripples and polygonal terrain.

References


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