Icarus 221 (2012) 289–295

Contents lists available at SciVerse ScienceDirect

Icarus

journal homepage: www.elsevier.com/locate/icarus

Locations of thin liquid water layers on present-day Mars

Akos Kereszturi [a,b,*], Edgard G. Rivera-Valentin [c]

[a] Konkoly Observatory, Research Centre for Astronomy and Earth Sciences, Hungarian Academy of Sciences, H-1121 Budapest, Konkoly Thege M. 15-17, Hungary
[b] New Europe School for Theoretical Biology and Ecology, H-1038 Budapest, Hollos Korvin 5, Hungary
[c] Department of Geological Sciences, Brown University, 324 Brook St., Box 1846, Providence, RI 02912, United States

ARTICLE INFO

Article history:
Received 18 January 2012
Revised 2 August 2012
Accepted 3 August 2012
Available online 14 August 2012

Keywords:
Mars
Exobiology
Ices
Mars, surface
Mars, polar caps

ABSTRACT

CRISM indicates the presence of water ice patches in Richardson crater, located on Mars' southern polar region at the area of the seasonal ice cap. Numerical simulations suggest that the maximum daytime temperature of the ice at these locations is between 195 and 220 K during local spring. Previous studies suggest that at these temperatures liquid interfacial water could be present. Here, for the first time, we provide an example where the environmental conditions allow for the formation of such liquid films on present day Mars at the southern hemisphere. The upper bound estimated H_2O loss during the presence of these water ice patches is approximately 30 μm between Ls = 200 and 220, though it may be as low as 0.1 μm depending on the ambient water vapor. The upper bound value is larger than the expected condensation thickness in autumn; however, it may still be realistic due to CO_2 gas jet generated deposition and possible subsequent accumulation on mineral grains. The presence of this interfacial water may have impact on local chemical processes along with astrobiological importance.

1. Introduction

The mounting evidence that liquid water is present on Mars today is one of the most important regarding active geochemical processes and the planet's astrobiological potential. Some theoretical models predict that liquid water may be present (Clow, 1987; Haberle et al., 2001; Hecht, 2002), above all as interfacial water (Mohlmann, 2004, 2010; Kossacki and Markiewicz, 2008) where water ice is in physical contact with mineral grains. Salts may influence the appearance of liquid water by decreasing the melting point. The presence of brines is suggested by the observations at the Phoenix landing site and other locations (Kossacki et al., 2004a,b; Chevrier et al., 2009; Hecht et al., 2009; Renno et al., 2009), which may act as a possible agent for the formation of some recent flow-like features (Brass, 1980; Mellon and Phillips, 2001; Knauth and Burt, 2002; Motazedian, 2003; Kossacki et al., 2004a,b) at polar dunes too (Kereszturi et al., 2009; Szynkiewicz et al., 2009) – although dry mass movements could also produce these (Hansen et al., 2010).

In this study, we analyze one special type of location where brines and interfacial water are possible on Mars, the Dark Dune Spots. These spots, which we abbreviate as DDS (Horvath et al., 2001), are part of the great number of interesting seasonal features that appear during local spring in the polar regions of Mars. Larger spots in this group contain a water ice layer on their ring-shaped outer area (Kereszturi et al., 2011) surrounding the darkest and barren central core. In this ring-like feature, the surface is darker than the CO_2 ice covering surrounding terrains and as a result it absorbs more sunlight and thus experiences higher temperatures than the bright frost. Temperature measurements of the martian surface are available with spatial resolution substantially worse than the diameter of these spots (50–100 m). Since the temperature during the presence of water ice cover may be close to the threshold limit (180 K) for interfacial water formation along the mineral–water ice interface, it is important to elucidate the possible appearance of liquid interfacial water at the DDS. In this study, we modeled the temperature at these small areas in order to investigate the possibility of liquid water formation.

2. Methods

Observational data and modeling results were used to analyze the possibility of interfacial water at the target regions. Imaging data was acquired from the Mars Reconnaissance Orbiter (MRO) High Resolution Imaging Science Experiment (HiRISE), and topographic data from MOLA and HRSC digital terrain models (DTMs) based on stereo images.

2.1. Observational data analysis

Using CAT-ENVI software (Morgan et al., 2009), we analyzed CRISM spectral data (Murchie et al., 2007) within Richardson crater

* Corresponding author at: Konkoly Observatory, Research Centre for Astronomy and Earth Sciences, Hungarian Academy of Sciences, H-1121 Budapest, Konkoly Thege M. 15-17, Hungary.
E-mail address: kereszturi.akos@csfk.mta.hu (A. Kereszturi).

0019-1035/$ - see front matter
http://dx.doi.org/10.1016/j.icarus.2012.08.004

(72°S179°E) using CRISM "FRT" observations (Full Resolution Targeted), which are characterized by a high spatial and spectral resolution (18 m and 7 nm respectively). Near-IR wavelengths were considered between 1 and 4 μm, observations were corrected for photometry and atmospheric gas absorptions (McGuire et al., 2009), and a filtering method (Parente, 2008) is used to reduce noise. The interpretation of spectra measured by push-broom sensors may be influenced by the effect called "spectral smile": the mean wavelength of spectels (spectel is the contents of a spectral channel for a particular pixel) and the spectral resolution can slightly change from one spatial pixel to the next (Ceamanos and Douté, 2009). We do not correct for the spectral smile because we take spectra near the center of the CRISM image where band shifts are lower than 0.002 μm (McGuire et al., 2009), which has a minor impact on wide and deep absorption bands.

For the albedo measurements, HiRISE images were also used and the gained results were correlated to TES based albedo estimations in order to have more than one source of information – although the worse spatial resolution TES data made possible was used only to correlate the trends in albedo changes. Absolute albedo values were approximated from HiRISE images using the *DN* values of certain pixels from the RED channel images. We calculated the reflectivity with the formula $I/F = (DN * \mathrm{SCALING\_FACTOR}) + \mathrm{OFFSET}$, and divided it with $\cos i$, in order to have approximated Lambertian albedo, which shows the same radiance when viewed from any angle. The Lambertian albedo is a good approximation of the albedo of the ice covered surface of Richardson crater at this time of the year. The resulting value, though, depends on several factors, such as the surface roughness, and especially on the incidence angle used in the correction. Although small-scale topography might affect the observed and calculated albedo values, such small-scale topography on Mars is unknown and cannot be taken into account – the used averages are the best approaches available. Beyond this theoretical argumentation the comparison between TES based measured and model based calculated values in this study (see Section 4) suggest that our model for the temperature estimation is realistic. We assume the terrain is a horizontal plane and also made several control measurements at various points of the analyzed "penumbra-like" area of the spots where water ice is covering the surface.

Albedo values were calculated and averaged for small 10×10 m square shaped homogeneous sections of the analyzed ring-like features. These sections were located manually, as rarely do small CO_2 patches appear at the ring-like or the barren central dark part. Based on manual analysis, the derived average albedo values are representative of the ring-like structure. The measured albedo values were scattered between 0.23 and 0.26 and on average 0.25. The TES based albedo values were also analyzed to validate and correlate the HiRISE based values and their changes. Unfortunately, TES data's spatial resolution is far worse than what is required for the analysis of structures as small as the ring-features. The correlation of HiRISE and TES based albedo values showed parallel trend, validating the analyzed albedo changes, but because of the low spatial resolution of TES data, HiRISE values were used for the model calculations in this study.

Temperature data was obtained from two sources: measured values with low spatial resolution (for comparison) and modeled values with high spatial resolution (for detailed analysis), where the latter could account for the small size of the target features. Observed values were derived from the Thermal Emission Spectrometer (TES) (MGS) measurements (Christensen et al., 1992), with "vanilla" software. This command line software is produced by the Arizona State University for the planetary science community to read and query binary data from TES dataset, correlate between various data tables, and was used presently to acquire surface temperature values for daytime between 12 and 14 local true solar time with spatial resolution of approximately 3–8 km. As a result, these values can only be considered a rough approximation of the surface temperature throughout the analyzed terrains and therefore are used only to analyze annual trends and to correlate them with model estimated values. For the analysis of certain solar longitude intervals at the surveyed locations, Thermal Emission Imaging System (THEMIS) and OMEGA (Visible and Infrared Mineralogical Mapping Spectrometer) data are not sufficient because their rare acquisition date and different local time.

2.2. Thermal modeling

Surface and subsurface temperatures were calculated by solving the one-dimensional thermal diffusion equation using a finite element approach as described by Rivera-Valentin et al. (2011), Rivera-Valentin (2012) and Ulrich et al. (2010). This method allows for the high spatial resolution required within this study. The vertical extent of the homogenous regolith column modeled was considered down to several times the annual skin depth in order to accurately reach convergence. Thus, the model simulated temperatures to a depth of 10 m with finite element thickness of 0.01 m and a corresponding time step of 10 s. The surface boundary condition for the column was considered radiative where the incoming solar heat flux is given by:

$$Q_{solar} = (1 - A)\frac{S_0}{r^2}\cos\zeta T(\zeta, \tau) \tag{1}$$

where A is albedo, S_0 is the solar flux at 1 AU, r is the instantaneous Sun–martian distance in AU, ζ is the solar angle to zenith, and $T(\zeta, \tau)$ is the transmission coefficient, which is a function of the zenith angle and the atmospheric opacity (τ). As is shown in Blackburn et al. (2009), the transmission coefficient is a polynomial fit to the data from Pollack et al. (1990) as presented by Rapp (2008). Atmospheric perturbations to the incoming heat flux considered were the indirect solar illumination due to scattering ($Q_{scattering}$) and atmospheric thermal emission (Q_{IR}):

$$Q_{scattering} = (1 - A)\frac{S_0}{r^2}(1 - T(\zeta, \tau))f_{scat} \tag{2}$$

$$Q_{IR} = (1 - A)\frac{S_0}{r^2}f_{atm}\varepsilon\cos(\delta - \phi) \tag{3}$$

where f_{scat} (0.02) and f_{atm} (0.04) are the fractional amounts of the relevant flux reaching the martian surface (Schmidt et al., 2009), δ is the solar declination, and ϕ is latitude (Kieffer et al., 1977; Applebaum and Flood, 1989; Aharonson and Schorghofer, 2006; Blackburn et al., 2009; Schmidt et al., 2009; Rivera-Valentin et al., 2010; Ulrich et al., 2010; Rivera-Valentin, 2012). The lower boundary condition includes a modest geothermal heat flux from below assigned as 30 mW/m^2 as previously applied by Ulrich et al. (2010).

We assumed a homogenous regolith column whose thermal properties were obtained as a weighted mass fraction of water ice and soil such that:

$$k = k_{ice}f_{ice} + k_{soil}(1 - f_{ice}) \tag{4}$$

$$C = C_{ice}f_{ice} + C_{soil}(1 - f_{ice}) \tag{5}$$

where k is thermal conductivity, C is volumetric heat capacity, f_{ice} is the fractional amount of ice within the soil column, and the subscripts denote the material. The thermal conductivity of water ice at the average surface temperature at the latitude considered is 3.4 W m^{-1} K^{-1} (Petrenko and Whitworth, 1999) with a volumetric heat capacity of 1.4×10^6 J m^{-3} K^{-1} (Giaque and Stout, 1936). The soil's thermal properties were obtained from the Phoenix Lander results, which found a thermal conductivity of 0.085 W m^{-1} K^{-1} and volumetric heat capacity of 1.05×10^6 J m^{-3} K^{-1} (Zent et al.,