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Theoretical prediction of the thermal conductivity and temperature variation inside mars soil analogues

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Abstract

Mars soil analogues, in dry and frozen conditions, are investigated, as far as the thermal conductivity prediction and the temperature variation, along its depth, are concerned. The thermal conductivity is theoretically predicted with the cubic cell model, which requires the knowledge of the thermal conductivity of the solid particle and of the materials present, i.e. atmospheric gas and/or frozen ice, and the porosity of the soil analogue. The soil mineral composition allows to evaluate the thermal conductivity of the solid particle. The heat capacity of the soil analogue is evaluated with the knowledge of its physical properties, the porosity and the specific heats of the materials present. The thermal diffusivity is calculated as the ratio of the thermal conductivity and heat capacity and results to be a function of the porosity and the ice mass content of the soil analogue. The temperature variations, in dry and partially frozen soil analogues, are predicted during a Martian day. The temperature variation, at different depth, is attenuated, as compared to the surface variation and a phase delay is present, depending on the soil thermal properties. The temperature variation, as well as the derivative of the temperature variation with the depth, is dependent on the thermal diffusivity of the soil analogue. In conclusion, the temperature measurement, along the depth of a Martian soil analogue, can be used to verify its physical status, i.e. dry or partially frozen.

Keywords: Martian soil analogue; Thermal conductivity prediction; Cubic cell model; Temperature variation; Thermal diffusivity

1. Introduction

Images of the surface of Mars, taken by different spacecraft (Viking, Pathfinder), suggest that the upper few kilometres of the crust consists in a thick porous megaregolith, mainly composed by ejecta of impact craters. The lithology varies from loose fine-grained regolith to meter-sized block of ejecta, all with a significant amount of volcanics. If a large quantity of volatiles still resides on Mars, these volatiles might still be concentrated in the Martian regolith and megaregolith as clay mineral, and as ground and underground condensates such as brines on the surface and ice in the underground. With a dry periglacial-type climate, Martian ground ice should be stable at some depth over virtually the entire planet while in the first meters the soil is able to exchange volatiles with the thin atmosphere of the planet. In this soil layer of ice free regolith there will be fluxes of volatiles in and out and the deposition or the evaporation of volatile constituents influence the physical parameters of the soil. Thermal properties of the regolith and megaregolith should be closely approximate by those of terrestrial frozen soil and basalt. Laboratory measurements of thermal conductivity of 144 samples of frozen soil and 301 samples of basalt are reported in Clifford and Fanale (1985). The samples vary in composition, lithology, porosity, and degree of pore saturation, but all represent conditions that are likely to exist somewhere within the top crust of Mars. The thermal conductivity of the soil samples range from 0.08 to 4 W/m/K, with an average value of about 1.66 W/m/K, while the basalt conductivity exhibits a greater spread: 0.09–5.40 W/m/K, with an average value of about 2.06 W/m/K. If the assumption that the outer crust of Mars contains a significant amount of volcanics is correct, then the thermal properties will be most strongly influenced by basalt. Clifford (1993), suggests a column averaged thermal conductivity of about $2.0 + 0.1$ W/m/K, with lower values likely near the surface, where the fine-grained materials may be present in abundance. In contrast, previous estimates were often based on the thermal properties of

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terrestrial analogue. For example, Fanale (1976), assumed that the properties of the megaregolith were well represented by a hard-frozen limonitic soil (0.8 W/m/K), while Crescenti (1984), assumed a purely basaltic composition (2.09–2.5 W/m/K). Thermal diffusivity is also uncertain, being unknown the value of density and specific heat, but in different environments it has been assumed values of about 10^{-6} m^2/s (Fanale et al., 1986).

The present investigation is aimed to evaluate theoretically the thermal conductivity of extraterrestrial soil analogues, as in planets and comets, on the base of the physical parameters. The theoretical prediction of the thermal conductivity of terrestrial soils, subjected to permafrost conditions, has been carried out in Gori (1983a), with reference to a four-phase soil, where solid particle, ice, unfrozen water and air are present. The model is based on the physical assumption that the unit cell is made of a cubic solid particle, surrounded by the other three phases, disposed in a prescribed shape, according to adsorption and capillarity. The prediction of the thermal conductivity has been extended to non-spherical particles in Gori (1983b), with special emphasis to soils with large values of porosity. The cubic cell model has been used to study other kinds of porous media in Gori (1986). The hypothesis of parallel isotherms has been used to solve the one dimensional conduction equation in Gori (1983a, b, 1986), while in Gori and Corasaniti (1999) the heat flux lines have been employed. The results of the model have been compared to those of other models, proposed in the literature, in Tarnawski et al. (2000), while in Gori et al. (2001) a comparison has been done with the data of experiments carried out in the same laboratory. Radiation, coupled to conduction, has been introduced in the model in Gori and Corasaniti (2001a), while the importance of the coupled phenomenon of mass transfer, added to heat conduction, has been studied in Gori and Corasaniti (2002a); Tarnawski and Gori (2002). The application of the cubic cell model to extraterrestrial conditions, taking into account the thermal radiation, has been further done in Gori and Corasaniti (2001b). In conclusion, the cubic cell model can be considered as a reliable theoretical model, able to study the thermal conductivity of soils in extraterrestrial conditions, like those present in other planets or comets, where the physical conditions can be different from the Earth, as far as pressure, temperature, and fluids in the atmosphere are concerned. The theoretical model can provide a relationship between the structure of the soil and the materials present (in solid or vapour phases). For example, the soil in contact with the atmosphere, in the hottest period of the day, has the lowest value of thermal conductivity because the materials is present only in gas phase. The presence of frozen fluids within the soil, increases its thermal conductivity, as compared to the value on the surface. The theoretical models can assist in the prediction of higher values of thermal conductivity, and thermal diffusivity, and finally in the temperature distribution in the soil analogue.

2. Theoretical model for the prediction of the thermal conductivity of Mars soil analogues

The theoretical model considers the dry soil as composed of a cubic cell, with a solid cubic particle at the centre and gas around it, as sketched in Fig. 1a.

The thermal conductivity of the soil, in the present work, has been evaluated by the solution of the one-dimensional heat conduction equation, under the assumption of parallel isotherms. The effective thermal conductivity of the dry soil analogue, or two-phase medium, is then given by

$$\frac{1}{k_{\mathrm{eff}}} = \frac{\beta - 1}{k_{\mathrm{atm}} \cdot \beta} + \frac{\beta}{k_{\mathrm{atm}} \cdot (\beta^2 - 1) + k_{\mathrm{s}}} \tag{1}$$

where k_{eff} is the effective thermal conductivity of the soil analogue, k_{atm} the thermal conductivity of the atmospheric gas, k_{s} the thermal conductivity of the solid particle, and β is given by

$$\beta = \frac{l_{\mathrm{t}}}{l_{\mathrm{s}}} = \sqrt[3]{\frac{\rho_{\mathrm{s}}}{\rho_{\mathrm{d}}}} = \sqrt[3]{\frac{1}{1 - p}} \tag{2}$$

where p is the porosity, ρ_{s} the density of the solid particle, ρ_{d} the dry density of the soil, l_{t} and l_{s} the dimensions indicated in Fig. 1a.

The first term in Eq. (1) corresponds to the thermal resistance of the gas in the cross section l_{t}^2, of length $(l_{\mathrm{t}} - l_{\mathrm{s}})$, while the second term is the thermal resistance of the materials gas and solid in the cross section l_{t}^2, of length l_{s}.

If the Martian soil analogue is not dry and the volume of ice content, V_{i}, is low, as compared to the volume of the void, V_{v}, the ice is considered adsorbed around the solid particle, as represented in Fig. 1b. In terrestrial soils the amount of liquid water adsorbed around the solid particle is a fraction of the water content at the permanent wilting point Gori and Corasaniti (2002a). The value of this fraction has been found empirically on the base of the experimental data on several terrestrial soils and is variable according to the type of soil. Because of the absence of reliable data on Martian soil analogue, the present paper assumes the ratio of adsorbed ice to void volume, or ice volume content equal to $f = V_{\mathrm{i}}/V_{\mathrm{v}} = 0.083$, as done in Gori (1983a). The upper limit of f is 1 when ice occupies all the void and the soil is considered saturated with ice. It must be noted that this is the only empirical constant of the theoretical model that can be removed with better knowledge of the Martian regolith.

The effective thermal conductivity is then given by

$$\frac{1}{k_{\mathrm{eff}}} = \frac{\beta - 1 - \delta/3}{\beta k_{\mathrm{atm}}} + \frac{\beta\delta}{3[k_{\mathrm{atm}}((\beta^2 - 1) + k_{\mathrm{i}})]} + \frac{\beta}{k_{\mathrm{s}} + \dfrac{2}{3}\delta k_{\mathrm{i}} + k_{\mathrm{atm}}(\beta^2 - 1 - \frac{2}{3}\delta)} \tag{3}$$

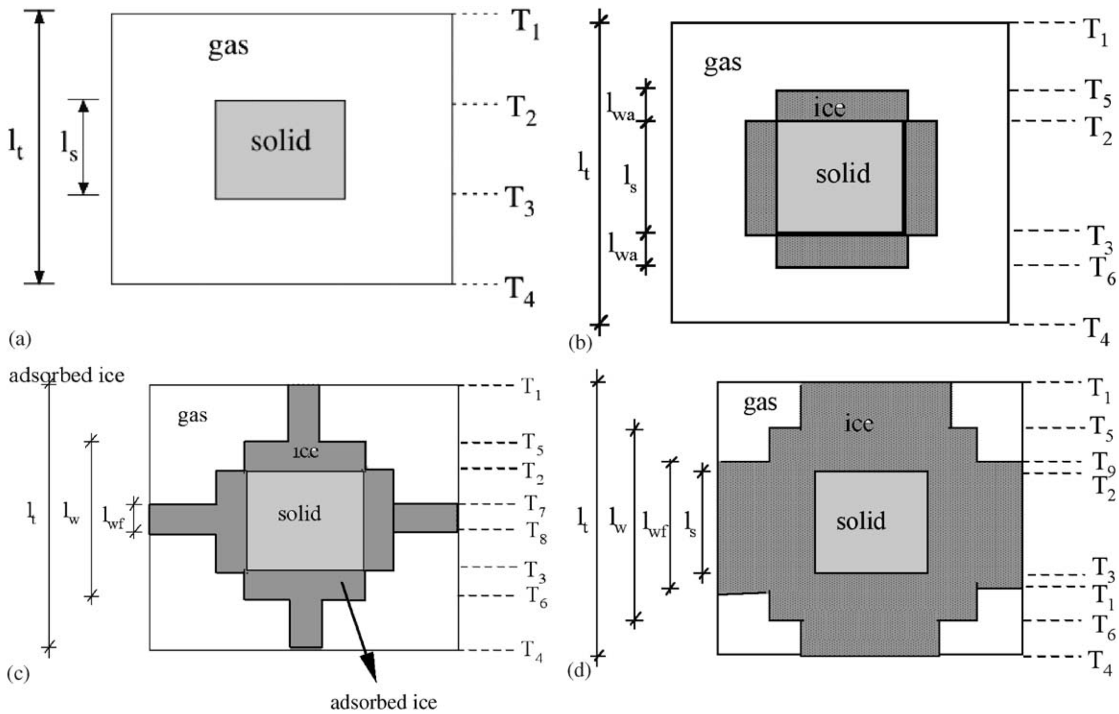

Fig. 1. (a) Cubic cell model for a dry soil analogue. (b) Soil analogue with adsorbed ice and gas. (c) Soil analogue with ice adsorbed around the solid particle and disposed among adjacent particles, for $\gamma_f < 1$. (d) Soil analogue with ice disposed among adjacent particles, for $\gamma_f > 1$.

where k_i is the thermal conductivity of ice and δ is given by

$$\delta = \frac{V_i}{1 - p} \tag{4}$$

with V_i the volume of ice.

The first term of Eq. (3) is the thermal resistance of the gas in the cross section l_t^2, of length $(l_t - l_s - 2l_{wa})$. The second term is the thermal resistance of the materials ice and gas in the cross section l_t^2, of length $(2l_{wa})$. The third term is the thermal resistance of the materials ice, gas and solid, in the cross section l_t^2, of length l_s.

Fig. 1c presents the distribution of ice and gas in the soil analogue when the volume content of ice is higher than the adsorbed value, $f = 0.083$, but the amount of ice among the adjacent solid particles is enough low, such that $\gamma_f < 1$, as defined in the following Eq. (5).

The following variables are then defined:

$$\gamma = \sqrt[3]{\frac{V_i}{V_s} - \frac{V_{if}}{V_s} + 1}, \qquad \gamma_f = \sqrt{\frac{V_{if}/V_s}{3(\beta - \gamma)}}\,, \tag{5}$$

where V_i is the ice volume, V_s the volume of the solid particle and V_{if} is the ice volume accumulated among adjacent solid particles and called funicular, according to Luikov (1966). In order to have a simple model, V_{if}/V_s is assumed linearly variable with the porosity of the soil analogue, between 0.183, for $p = 0.4764$, and 0.226, for $p = 0.2595$. The resulting expression is then:

$$\frac{V_{if}}{V_s} = \frac{V_{if}}{V_v}(\beta^3 - 1) = \left[0.183 + \frac{0.226 - 0.183}{0.4764 - 0.2595}(0.4764 - p)\right](\beta^3 - 1) \tag{6}$$

In the configuration of Fig. 1c, for $\gamma_f < 1$, k_{eff} is given by

$$\frac{1}{k_{eff}} = \frac{\beta^{2\cdot} - \beta\gamma}{k_{atm}(\beta^2 - \gamma_f^2) + k_i\gamma_f^2} + \frac{\beta\gamma - \beta}{k_{atm}(\beta^2 - \gamma^2) + k_i\gamma^2} + \frac{\beta - \beta\gamma_f}{k_{atm}(\beta^2 - \gamma^2) + k_i(\gamma^2 - 1) + k_s} + \frac{\beta\gamma_f}{k_s + k_i(\gamma^2 - 1 + 2\beta\gamma_f - 2\gamma\gamma_f) + A} \tag{7}$$

where

$$A = k_{atm}(\beta^2 - \gamma^2 - 2\beta\gamma_f + 2\gamma\gamma_f). \tag{8}$$

The first term of Eq. (7) is the thermal resistance of the materials gas and ice in the section (l_t^2), of length $(l_t - l_w)$. The second term is the thermal resistance of the materials gas and ice in the section (l_t^2), of length $(l_w - l_s)$. The third term is the thermal resistance of the materials gas, ice and solid in the section (l_t^2), of length $(l_s - l_{wf})$. The fourth term is the thermal resistance of the materials gas, ice and solid in the section (l_t^2), of length (l_{wf}).

Table 1
Thermal conductivity of Martian atmosphere

T (K)	k_{atm} 10^3 (W/m/K)
148	6.86
173	8.39
223	11.87
273	14.49
298	16.01

Table 2
Porosity and ice mass content

$a = M_i/M_t$ for ice saturated soil	
p	$a10^3$ ($T = 223$K)
0.20	73.78
0.30	120.15
0.40	175.20
0.50	241.63

Fig. 1d presents the distribution of ice and gas in the soil analogue for a large volume of ice, such that $\gamma_f > 1$.

The effective thermal conductivity, k_{eff}, is then given by

$$\frac{1}{k_{eff}} = \frac{\beta^{2\cdot} - \beta\gamma}{k_{atm}(\beta^2 - \gamma_f^2) + k_i\gamma_f^2} + \frac{\beta\gamma - \beta\gamma_f}{k_{atm}(\beta^2 - \gamma^2) + k_i\gamma^2} + \frac{\beta\gamma_f - \beta}{A + k_i(\gamma^2 + 2\beta\gamma_f - 2\gamma\gamma_f)} + \frac{\beta}{k_s + k_i(\gamma^2 - 1 + 2\beta\gamma_f - 2\gamma\gamma_f) + A}, \quad (9)$$

where A is given by Eq. (8).

The first term of Eq. (9) is the thermal resistance of the materials gas and ice in the section (l_t^2), of length ($l_t - l_w$). The second term is the thermal resistance of the materials gas and ice in the section (l_t^2), of length ($l_w - l_{wf}$). The third term is the thermal resistance of the materials gas and ice in the section (l_t^2), of length ($l_{wf} - l_s$). The fourth term is the thermal resistance of the materials gas, ice and solid in the section (l_t^2), of length (l_s)

3. Martian soil analogue and atmospheric gas

The Martian soil analogue used in the present work is the olivine tested in the Heat Transfer Laboratory (Gori and Corasaniti, 2002b). The thermal conductivity of the solid particle has been evaluated theoretically, on the base of the experiments, and has been found equal to $k_s = 2.94$ W/m/K (Gori and Corasaniti, 2002b). The evaluation of the thermal conductivity of the solid particle is based on two experimental measurements of the thermal conductivity of olivine, i.e. in dry and water saturated conditions. The two experimental values are then used in Eq. (1) as k_{eff} to give two equations which are then solved to calculate the values of β and k_s.

The Martian atmospheric pressure is about 6 mbar and the atmospheric gas is composed by CO_2 (95%), N_2 (3%), Ar (1.5%) and traces of water vapour. The thermal conductivity of the Martian atmospheric gas has been evaluated as a weighed average, and Table 1 reports its values, at several temperatures.

A relation between ice volume content, $f = V_i/V_v$, ice mass content, $a = M_i/M_t$ and soil porosity, p, can be found, where M_i is the mass of ice in the porous medium, M_t the total mass of the porous medium, V_i the ice volume in the porous medium, and V_v the void volume of the cubic cell. The ice mass content, a, is given by

$$a = \frac{1}{1 + \frac{\rho_{atm}}{\rho_{ice}}\frac{1-f}{f} + \frac{\rho_s(1-p)}{\rho_{ice} f p}}. \quad (9)$$

For a given soil porosity, p, the ice content $a = M_i/M_t$, corresponding to a saturated frozen soil, is reported in Table 2.

4. Thermal conductivity evaluation

Fig. 2 presents the thermal conductivity of the Martian soil analogue versus porosity, at several ice mass contents and at the average temperature of Mars, assumed equal to 223 K.

In dry condition ($M_i = 0$) the thermal conductivity decreases with the increase of porosity, from the value of the solid particle (2.94 W/m/K), at $p = 0$, to that of the atmospheric gas (0.012 W/m/K), at $p = 1$. At each ice mass content, M_i/M_t, the thermal conductivity decreases from the maximum value, predicted at low porosity, when the cubic cell is full of ice, to the minimum one, at high porosity, when the void space inside the cell is occupied by the atmospheric gas. It must be noted that the maximum thermal conductivity is around 2.7 W/m/K, almost independently of the ice mass content.

Fig. 3 presents the thermal conductivity of a partially frozen Martian soil analogue, versus the ice mass content, at several porosities and $T = 223$ K. For a porosity equal to 0.2, the thermal conductivity spans from the minimum value, in dry conditions, equal to about 0.144 W/m/K, to the maximum one, in saturated conditions, equal to 2.7 W/m/K. For a porosity equal to 0.5, the thermal conductivity spans from the minimum value, in dry conditions, equal to about 0.05 W/m/K, to the maximum one, in saturated conditions, equal to 2.7 W/m/K.

Fig. 3 reports also the lines with the same ice volume content f. The thermal conductivity, at the same f, decreases with the increase of the ice mass content, because the thermal conductivity of the solid particle is higher than the ice value.

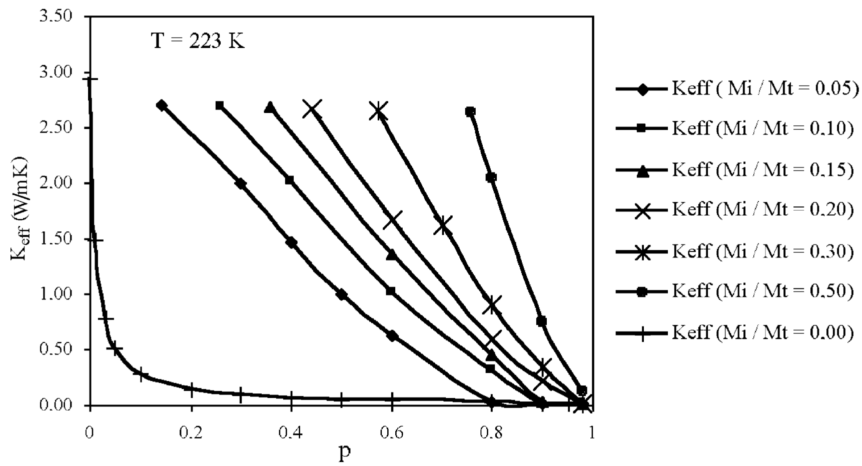

Fig. 2. Thermal conductivity of the Martian soil analogue versus porosity, at several ice mass contents and 223 K.

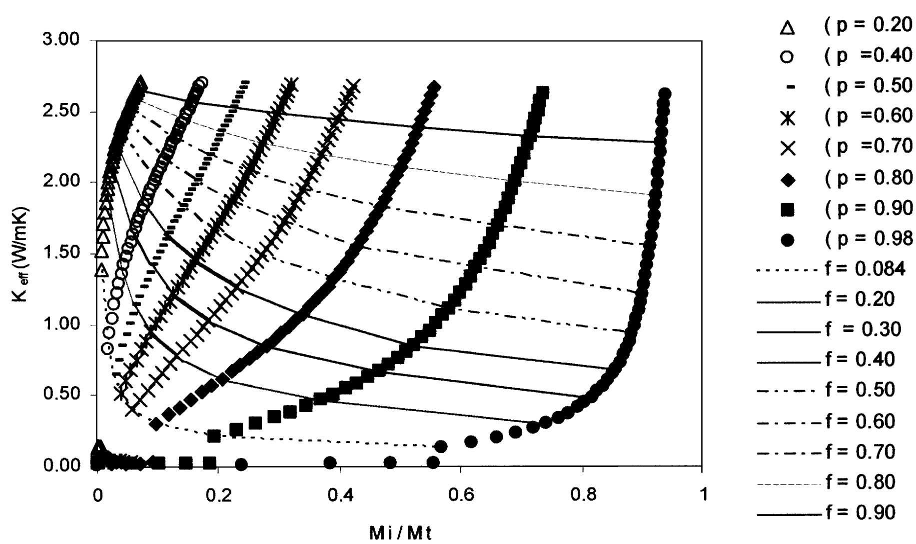

Fig. 3. Thermal conductivity of a partially frozen Martian soil analogue, versus ice mass content, at several porosities and 223 K.

5. Thermal diffusivity evaluation

The heat capacity of the Martian soil analogue is evaluated with the following Eq. (10), and the values are reported in Fig. 4.

$$(\rho c_{\rm p})_{\rm eff} = (\rho c_{\rm p})_{\rm s}(1-p) + (\rho c_{\rm p})_{\rm ice} f p + (\rho c_{\rm p})_{\rm atm}(1-f)p. \qquad (10)$$

The heat capacity of the Martian soil analogue increases, for a fixed porosity, with the ice mass content, with a higher increase for higher porosities.

The thermal diffusivity of the Martian soil analogue is evaluated with the ratio of $k_{\rm eff}$, given by Fig. 3, and ρc_p, given by Fig. 4. Fig. 5 presents the thermal diffusivity versus f, at several porosities and 223 K. The thermal diffusivity has a step increase with the passage from the condition of ice adsorbed around the solid particle, Fig. 1b, for $f < 0.083$, to the ice disposed among the solid particles, for $f > 0.083$, Fig. 1c, because of the ice bridges present among the solid particles. The thermal diffusivity increases with f, with a higher increase for higher porosities.

Fig. 6 reports the thermal diffusivity of the Martian soil analogue, versus porosity, at several ice mass contents and 223 K. The results of Figs. 5 and 6 show that the thermal diffusivity of the Martian soil analogue changes of more than an order of magnitude from dry to frozen conditions.

Fig. 7 presents the thermal diffusivity versus the ice mass contents. The lines at constant f are reported.

6. Theoretical temperature distribution in the soil analogue

The differential equation of heat conduction in the soil

$$\frac{\partial}{\partial z}\left(K\frac{\partial \vartheta}{\partial z}\right) = C\frac{\partial \vartheta}{\partial t} \qquad (11)$$

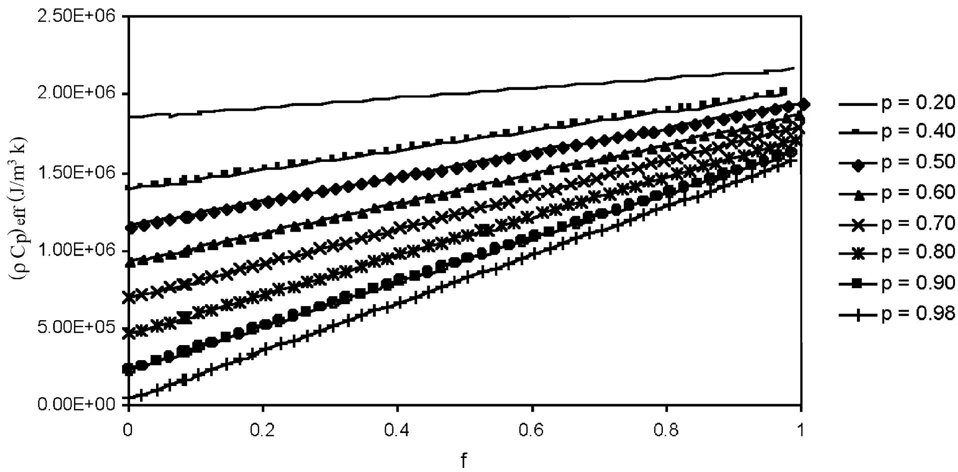

Fig. 4. Heat capacity of the Martian soil analogue versus ice volume content at several porosities and 223 K.

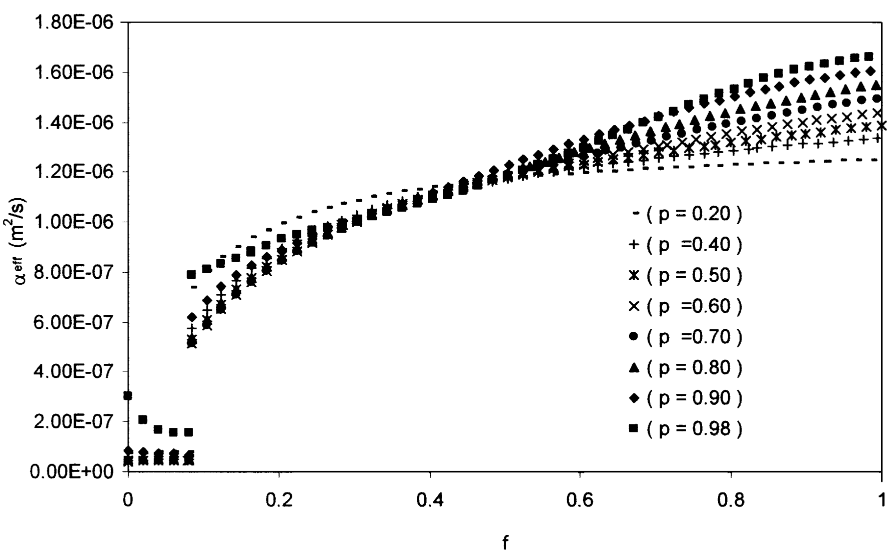

Fig. 5. Thermal diffusivity of the Martian soil analogue versus ice volume content at several porosities and 223 K.

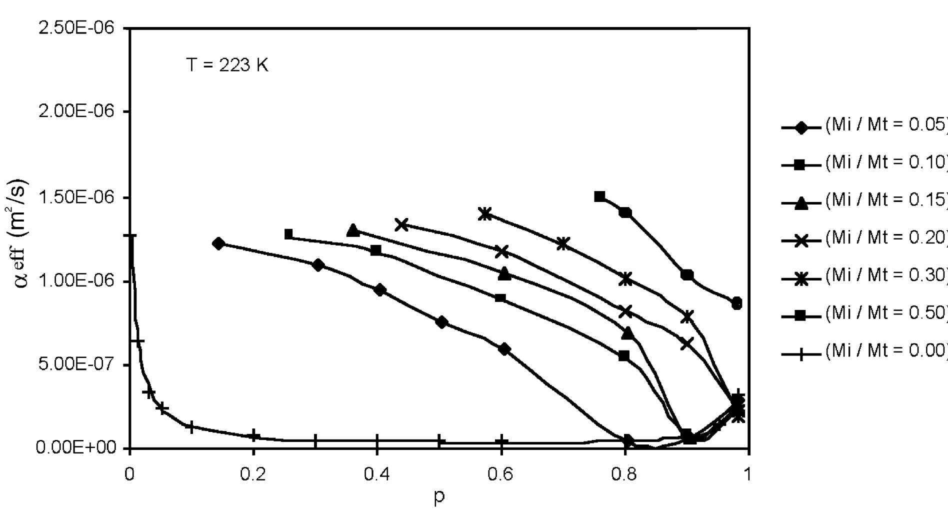

Fig. 6. Thermal diffusivity of the Martian soil analogue versus porosity at several ice mass contents and 223 K.

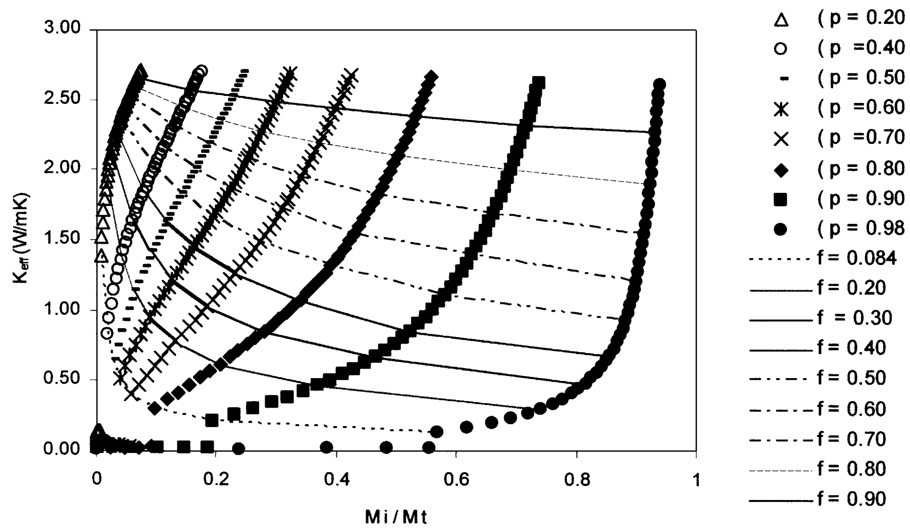

Fig. 7. Thermal diffusivity of the Martian soil analogue versus ice mass content at several porosities and 223 K.

has the solution

$$\vartheta(z,t) = \vartheta_a + \vartheta_s \exp\left(-\frac{z}{D}\right) \cdot \sin\left(\omega t - \frac{z}{D} + \varphi\right) \quad (12)$$

under the following boundary conditions:

- on the surface, $z = 0$, the temperature variation is

$$\vartheta(0,t) = \vartheta_a + \vartheta_s \cdot \sin(\omega t + \varphi) \quad (13)$$

- at an infinite depth the temperature fluctuation decreases to zero

$$\lim_{z\to\infty} \vartheta = \vartheta_a \quad (14)$$

where ϑ_a is the average temperature in the soil, ϑ_s the amplitude of the temperature variation on the surface, t the time, ω is the radial frequency

$$\omega = \frac{2\pi}{T} = \frac{2\pi}{86400} = 7.27 \times 10^{-5}\ \mathrm{s}^{-1} \quad (15)$$

where the Martian day has been assumed equal to 24 h and φ is the phase delay.

Further on:

$$D = \sqrt{\frac{2K}{C\omega}}, \quad (16)$$

where k is the thermal conductivity of the soil, C the heat capacity and $\alpha = k/C$ the thermal diffusivity.

7. Numerical temperature variation in Martian soil analogues

The first Martian soil analogue investigated corresponds to a dry soil, with the following characteristics: porosity, $p = 0.20$, thermal conductivity equal to $k = 0.144$ W/m/K, as evaluated by the theoretical model and $\alpha = 10^{-6}$ m^2/s, as suggested by the literature review. The temperature variation on the surface is assumed variable from the minimum value of 148 K to the maximum one of 298 K. The temperature variations with time are not reported because of space limits. At 0.30 m the temperature variation, during the day, is about 25 K and at 0.60 m is 4 K. At 1 m the temperature variation is 0.4 K, while at 1.30 m is 0.06 K (Gori and Corasaniti, 2003). The temperature variations are presented versus the depth of the soil analogue in Fig. 8 at several hours during the day, where 0 h is the beginning of the temperature variation, equal to 223 K. Fig. 8 shows that the temperature variations are detectable, inside the dry soil analogue, up to a depth of about 0.60 m.

The second Martian soil analogue investigated corresponds to a frozen soil analogue, with the following characteristics: porosity, $p = 0.20$, thermal conductivity equal to $k = 2.70$ W/m/K, according to the evaluation of the theoretical model and $\alpha = 10^{-5}$ m^2/s, approximately one order of magnitude higher than the dry soil. The temperature variation on the surface is assumed variable from the minimum value of 148 K to the maximum one of 298 K, with an average value of 223 K. At 0.30 m, the temperature variation, during the day, is about 85 K and at 0.60 m is about 48 K. At 1 m the temperature variation is about 22 K, at 1.30 m about 13 K and at 2 m about 3.3 K. The temperature variations are presented, versus the depth of the soil analogue, in Fig. 9,, at several hours during the day, where 0 h is the beginning of the temperature variation, equal to 223 K.

Fig. 9 shows that the temperature variations are detectable, inside the frozen soil analogue, up to a depth of about 1.5 –2 m. In other words, the temperature variation is more penetrating in the frozen soil, because the thermal diffusivity is one order of magnitude higher than in the dry soil. The temperature variations in the two soil analogues are reported, at several depths, in Table 3.

Table 3 shows the comparison between the temperature variations in the two soil analogues. At small depths the

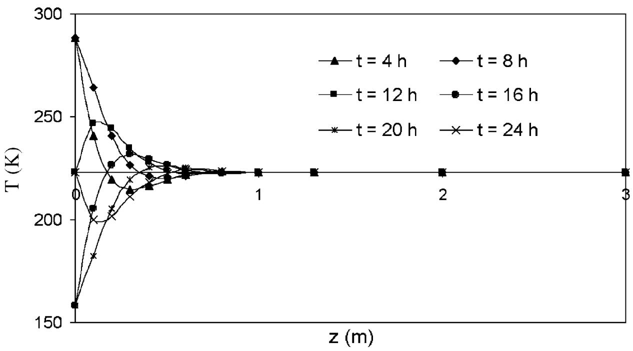

Fig. 8. Temperature distribution in a dry soil analogue versus depth at several times.

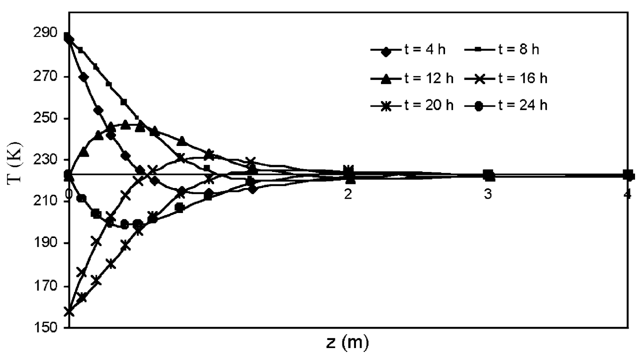

Fig. 9. Temperature distribution in a frozen soil analogue versus depth at several times.

Table 3
Temperature variation, $\Delta T = T_{max} - T_{min}$, in the soil analogues, at several depths

Depth z(m)	Frozen soil ΔT (K)	Dry soil ΔT (K)
0.10	123.70	82.00
0.20	102.40	44.90
0.30	84.60	24.60
0.40	69.90	13.40
0.50	57.80	7.40
0.60	47.70	4.00
0.80	32.60	1.20
1.00	22.20	0.40
1.30	12.60	0.06
2.00	3.30	9×10^{-4}
3.00	0.50	0.00
4.00	0.07	0.00
5.00	0.01	0.00

temperature variation is of the same order of magnitude, although higher in the frozen soil. On the other hand, Table 3 shows that the temperature variations are higher, of one or two orders of magnitudes, in the frozen soil, at large depths, like 1 or 2 m.

The data of Table 3 have been used to calculate the variation of $\Delta T = (T_{max} - T_{min})$ with the depth or, in other words, the derivative of ΔT with z, $\partial T/\partial z = \Delta T/\Delta z$ (K/m). The results are reported in Table 4.

Table 4
Derivative of the temperature variation, $\Delta T/\Delta z$ (K/m), at several depths

Depth z(m)	Frozen soil $\Delta T/\Delta z$ (K/m)	Dry soil $\Delta T/\Delta z$ (K/m)
0.15	213.00	371.00
0.25	178.00	203.00
0.35	147.00	112.00
0.45	121.00	60.00
0.55	101.00	34.00
0.70	75.50	14.00
0.90	52.00	4.00
1.15	32.00	1.10
1.65	13.30	84×10^{-3}
2.50	2.80	0.00
3.50	0.43	0.00
4.50	0.06	0.00

Table 4 shows the comparison between the derivative of the temperature variation, $\Delta T/\Delta z$, in the two soil analogues. At small depths $\Delta T/\Delta z$ is higher for the dry soil than for a frozen soil. On the other hand, at depths like 1 or 2 m, Table 4 shows that $\Delta T/\Delta z$ are higher, in the frozen soil, of one or two orders of magnitude.

8. Conclusions

The temperature variation during the Martian day, along the depth of a soil analogue, can be employed to evaluate its physical conditions, either in dry or frozen state. The temperature variations in a frozen Martian soil analogue, at about 0.40 m of depth, are five times higher than in a dry soil analogue. The derivative of the temperature variation with the depth, at a soil depth of 0.70 m is five times higher in a frozen soil than in a dry soil.

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