# High Altitude Infrared Spectroscopic Evidence for Bound Water on Mars

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The reflectivity of the Martian surface has been measured between 2.0 and 4.0  $\mu$ m. A broad absorption band is observed which has a minimum at 2.85  $\mu$ m. The position and shape of the band is compared to features in numerous terrestrial minerals. Bound water in the Martian surface material is the most likely cause of the observed band. The strength of the observed band suggests a bound water content of approx. 1% by weight. The band is inconsistent with > a few % limonito or hydrated polymeric carbon suboxide. A mean particle radius ~100 $\mu$ m is deduced. The quality of bound water derived has a range of geological, climatic, and exobiological implications.

#### INTRODUCTION

The possibility of bound water in the Martian surface material is of interest for the mineralogy of the planet, its past history of outgassing of volatiles, the possibility of more element environmental conditions at earlier epochs, and the possibility of life on the planet (Sagan

<sup>1</sup> Alfred P. Sloan Research Fellow. Copyright © 1973 by Academic Press, Inc. All rights of reproduction in any form reserved. et al., 1965; Pollack and Sagan, 1969). The first indication of bound water was provided by Sinton (1965), who found, from groundbased infrared spectra, evidence of a broad absorption band in the vicinity of  $3.0\,\mu$ . This feature is to be distinguished from the so-called Sinton bands at longer wavelengths now known to be primarily due to deuterated water in the terrestrial atmosphere (Rea, O'Leary and Sinton, 1965). A further indication of a bound water absorption feature was obtained by Beer *et al.* (1971) with a groundbased Michelson interferometer.

However, the above observers were unable to view Mars between 2.5 and  $3.0\,\mu\text{m}$ , because of the large opacity of telluric water vapor and carbon dioxide. By obtaining observations from aircraft altitude, where the water vapor above the observer is several orders of magnitude less than for a groundbased observer, we were able to view this region. As described below, this ability was most important because the band center lies in this previously inaccessible spectral region. Observations from an aircraft offer the added advantage of being able to record and meaningfully use the solar spectrum to obtain directly the desired reflectivity curve. Because of the great time variability of telluric water vapor, it is difficult to use directly the solar spectrum when observing from the ground and one often resorts to indirect means, such as recording stars of similar spectral type to the sun. Because at aircraft altitudes the atmosphere has close to

100% transmission over almost the entire spectral range of interest,  $2-4\,\mu\text{m}$ , the time variability of water vapor is far less of a problem.

#### INSTRUMENTATION

Observations were made from a NASA Convair 990 aircraft during the period August 4–10, 1971. During the observations, the aircraft altitude varied between 38,000 and 41,000ft (11.6–12.5km). The precipitable water vapor in the line of sight was always less than  $15\mu$ m. The elevation angle of Mars was  $55 \pm 2^{\circ}$ .

The light from Mars entered the aircraft through a 0.9in. (22.8mm) thick Corning Cortran No. 9753 window. A gyroscopically stabilized coelostat directed the radiation into a 12 in. (30.5cm) f/15 Cassegrain telescope and from there to the spectrometer. The optical layout of the spectrometer is shown in Fig. 1. The beam splitter passed 0.5% of the light to the guiding eyepiece. The remainder of the light was reflected by the beam splitter and 34 Hz chopper mirror to a small



FIG. 1. A schematic drawing of the spectrometer is shown. The Hadamard mask at the exit slit position allows approximately ten times more light to reach the detector compared to a conventional monochromator. The signal-to-noise ratio is therefore enhanced.



FIG. 2. The solar spectrum is shown. The large dip is due to an absorption band in the 22.8mm thick Cortran No. 9753 window.

antireflection coated Irtran II lens which matched the f/15 light cone to the f/5spectrometer. The entrance slit of the spectrometer was  $0.33 \times 1.00 \,\text{mm}$ , corresponding to  $54 \times 161$  arc seconds on the sky. A movable 19 slot Hadamard mask was mounted in the exit plane of the  $\frac{1}{4}$  m Ebert-Fastie spectrometer. The spectrometer had a fixed grating with 11.58 lines/ mm blazed at  $4.0\mu$ m. By shifting the position of the mask, it was possible to scan the entire spectrum from 2.0 to  $4.0\,\mu\text{m}$  in  $0.112\,\mu\text{m}$  steps. Since the mask allowed approximately 50% of the light in the 2-4 $\mu$ m band to reach the detector at any given instant, the resulting signalto-noise ratio was larger than for an equivalent monochromator. A complete analysis of the singly multiplexed Hadamard spectrometer has been given by Sloane (1969) and by Decker and Harwit (1968). After leaving the exit plane, the radiation was focused on the liquid nitrogen cooled PbS detector by an antireflection coated Si field lens. The output of the detector was amplified, demodulated, and recorded on punched paper tape.

Figure 2 shows the observed solar spectrum. The wavelength calibration was accomplished by fitting the dip in the solar spectrum to the measured dip in the spectrum of the window. An order separating filter  $(2.1 \,\mu\text{m} \text{ long wavelength pass})$ mounted just before the Hadamard mask was also used to check the calibration. We believe the calibration is accurate to  $\pm 0.04 \,\mu\text{m}$ . The solar spectrum was obtained by stopping down the telescope aperture by a perforated metal mask.

Figure 3 shows the lunar spectrum normalized by the solar spectrum of Fig. 1. The strong absorption due to the window is almost completely compensated by the normalization procedure. In particular, the lunar reflectivity at  $2.78\mu m$ , which is the wavelength position most strongly affected by the window absorption (see Fig. 2), lies along a line defined by the shorter wavelength points, which are not so affected. This leads us to believe that the flattening of the spectrum between 2.8 and  $3.0\,\mu\text{m}$  is mostly due to a similar behavior of the lunar spectrum. Beyond  $3\mu m$  the sharp rise in the lunar spectrum is due to the dominance of thermal emission over reflected sunlight.

Figure 4 shows the normalized spectra of Mars. The normalized spectrum has been adjusted to have a value of 1 at  $2.53 \,\mu$ m. In an attempt to better define the spectrum, two sets of spectra of Mars and the Sun were obtained, which were displaced from one another by approximately  $0.03 \,\mu$ m. The individual points shown in Fig. 4 are the result of averaging of



Fig. 3. The spectrum of the Moon divided by the solar spectrum is shown. The large dip shown in Fig. 1 is almost completely removed by the normalization procedure. The solid line is a mean curve drawn through the data points.

approximately 40 spectra of each type. The 80% depth of the Martian absorption feature is much larger than any residual absorption that can be imagined for the normalized lunar spectrum in Fig. 3. The wavelength position of the minimum is about  $2.85\mu$ m. To shift this position to significantly longer wavelengths would require a preferential absorption by the aircraft window of more than a factor of 2 in the Martian spectrum as compared with the solar spectrum, a possibility ruled out by the normalized lunar spectrum shown in Fig. 3. At most there is a 20%



FIG. 4. The normalized spectrum of Mars is shown. Unlike the lunar spectrum of Fig. 3, the normalized spectrum of Mars shows a large dip at about  $2.8 \mu m$ . This dip is identified with bound water in the Martian soil. The solid line is a mean curve drawn through the data points.

residual absorption in this latter spectrum, and as argued above most of this is probably due to the behavior of the Moon. In addition, spectra of Mars obtained on separate days agree quite well with one another.

For the purpose of subsequent analysis, we have converted the normalized Martian spectrum into one having the units of geometric albedo at  $2.5 \mu$ m. McCord *et al.* (1971) found that the reflectivity of Mars was approximately the same as 2.5 as at  $0.8 \mu$ m, while Irvine *et al.* (1968) determined a geometric albedo of 0.28 for Mars at  $0.8 \mu$ m.

### FACTORS INFLUENCING THE OBSERVED SPECTRA

The observed spectra are the sum of the radiation emanating from the entire disk of Mars near opposition. Besides containing light reflected from the dark and bright areas of Mars, the observed radiation contains contributions from the frostcovered southern polar cap, the aerosol and gaseous components of the atmosphere, and thermal emission. The discussion of the influence of these factors follows a similar discussion given by Pollack and Sagan (1969). Since the bright areas cover a large fraction of the disk of Mars and have a substantially larger reflectivity than the dark areas over most of the near infrared, the spectrum, when corrected for the above factors, will refer chiefly to the reflectivity of the bright areas.

At the time of the observations, the frost-covered area near the south polar cap subtended only about 1% of the solid angle subtended by Mars at the Earth. Furthermore, in the infrared the reflectivity of the cap is comparable to frost-free areas. Hence the frost-covered ground influenced the observations only in a very minor way.

At the time of the observations, the 1971 major dust storm on Mars had not yet begun. An upper limit to the contribution of aerosols to the observed radiation may be inferred from Mariner 6 and 7 results, which were taken at a relatively dust-free

time. Morphological details, such as craters, were equally visible in pictures taken with the blue, green, and orange filters (Leighton et al., 1969). Since in the blue the geometrical albedo of Mars is about 10% (Irvine et al., 1968), the geometrical albedo of the aerosols must be at most a few percent. Furthermore, if the particle diameters are comparable to a wavelength, the optical depth and hence the reflectivity will decrease toward longer wavelengths. As we have observed a change of the geometric albedo from about 28% outside the band to below 10%near the center of the band, we conclude that this band is not primarily the result of aerosols in the atmosphere.

Gases in the Martian atmosphere can influence the spectra by imposing absorption features at wavelengths where they have strong bands. Because the total pressure on Mars (several millibars) and the amount of water vapor (at most a few tens of precipitable micrometers) is so low, absorption by water vapor is at most a few percent even in its strongest absorption band. However, since carbon dioxide is a major constituent it can be a more significant factor. Over the spectral range under consideration carbon dioxide has strong absorption features only at 2.69 and  $2.77 \mu m$ . These bands are relatively narrow and for gas amounts and pressures typical of the Martian atmosphere, carbon dioxide can be expected to be highly absorbing only between about 2.67 and 2.81  $\mu$ m with the region centered around  $2.73\,\mu m$  being relatively transparent.

Particularly at the longer wavelengths, thermal radiation may make a large contribution to the observed radiation. One might even argue that the upswing in the ratio spectra toward longer wavelength is due solely to this effect. Sample calculations now described show that this is not the case. Pollack and Sagan (1969) found that, at the Martian season and wavelength region of interest. the temperature of an equivalent isothermal disk emitting the same amount of radiation as the actual Martian disk is about 273°K. One can readily show that such a blackbody will contribute an amount  $P_{\rm th}$  to the observed geometrical albedo:

$$P_{\rm th} = (1 - P) \pi B / F_{\rm s} \approx \pi B / F_{\rm s} \qquad (1)$$

where **B** is the Planck function and  $F_s$ is the solar flux at Mars' orbit and where P is the actual surface reflectivity. The spectral behavior of  $P_{\rm th}$  is displayed in Fig. 4, where it is compared with the observed values. The solid line in Fig. 4 is a mean line through the data points. This comparison implies that thermal emission is not responsible for the rise in the reflectivity from 2.8 to  $3.5\,\mu{\rm m}$ ; however, thermal emission is important at longer wavelengths. We conclude that the nonfrost-covered surface areas of Mars have a strong absorption feature in their reflectivity centered near  $2.85\,\mu{\rm m}$ .

### COMPOSITIONAL IMPLICATIONS

The spectra presented in Fig. 4 show a strong absorption band, centered at about  $2.85 \,\mu\text{m}$ . The uncertainty in this location is about  $0.1 \,\mu\text{m}$ . The band is further characterized by a sharp rise to shorter

wavelengths and a more gradual increase toward longer wavelengths from the band center. We surveyed the standard literature to determine which materials absorb with a minimum near  $2.85\,\mu$ m and with a contour comparable to that displayed in Fig. 4. With almost no exceptions, the compilations of Hunt et al. (1950), Miller and Wilkins (1952), and Hovis and Callahan (1966) show such features present only for bound water, a term referring to water present in the mineral assembly of the rocks, either as physically or as chemically bound water. The wavelength is a standard assignment of the O-H stretching vibrational fundamental.

The only apparent exceptions in the mineralogical literature are certain nitrates. However, in the case of the nitrates, their absorption feature near  $2.85\mu$ m is much narrower than the Martian feature and they have a stronger band centered near  $3.5\mu$ m, which is not present in the Martian spectrum. Nitrogenous organic compounds, such as imines, amines, and amides, display an absorption feature near the location of the Martian feature,

Sample	Size (µm)	Wavelength position (in $\mu m$ )		
		Minimum	Short	Long
NAA std granite	<38	2.78	2.60	3.8
Dunite-Merenfelder	<38	2.85	2.55	4.0
NAA std ore basalt	<38	2.84	2.60	4.0
NAA std ore serpentine	<38	2.70	2.60	3.8
Yellowstone tuff No. 6	<38	2.70	2.60	3.8
Tektite SE Asia	38 - 53	2.84	2.64	3.3
Atlantic city beach sand	38 - 53	2.94	2.60	3.4
Mohave Desert soil	38 - 53	2.85	2.60	3.85
Salt Pool, Death Valley	38 - 53	2.95	2.60	3.72
Pawnee Grassland soil	38 - 53	2.90	2.60	3.92
Gray clay		2.78	2.60	3.70
Rosamond Dry Lake soil		2.85	2.60	3.76
Goethite		3.20	2.55	4.6
Hematite		2.95	2.60	3.8
FeCl, (partially hydrated)		3.05	2.10	~3.8
Hydrated C <sub>3</sub> O <sub>2</sub> polymer		3.50		
NHACI		3.10		

TABLE I<sup>a</sup>

<sup>a</sup> References: Hovis (1965, 1966); Smith et al. (1963); Kuiper (1969).

due to the N-H stretch 'fundamental. For such compounds in the gas phase the location of the band center is about  $3.0\,\mu\text{m}$  and at somewhat longer wavelengths in the liquid and solid phases. Thus, while those compounds have a feature close to the observed location, it may not be close enough. In addition, if the observed feature were due to organic N-H, we would also expect a stronger C-H feature at  $3.3-3.5\,\mu\text{m}$ , which is not present in the Martian spectrum.

The above discussion indicates that minerals containing bound water, nitrates, and certain nitrogen-containing organic compounds show absorption features near the feature observed for Mars. On closer examination, we are able to rule out the nitrates and some arguments against the organic compounds are proposed. Bound water is at least consistent with the Mariner 9 observation of the silicate infrared reststrahlen features (Hanel *et al.*, 1972). In a later section, we will present additional evidence that the absorption feature is due to bound water.

As shown in Table I, the water band of most minerals is characterized by a reflectivity minimum within  $0.1 \mu m$  of  $2.85 \mu m$  and so is consistent with the observed band. However, there are several interesting exceptions. Goethite has been widely mentioned as being a major constituent on Mars (see, e.g., Dollfus, 1957; Pollack and Sagan, 1969). However, it exhibits an absorption band centered at  $3.2\,\mu m$  (Hovis, 1965), inconsistent with our  $2.85 \mu m$  feature. Plummer and Carson (1969) have suggested that the red color of Mars is due to polymers of carbon suboxide, which cover the surface. While polymeric carbon suboxide does not exhibit any strong features near  $3\mu m$ , its hvdrated form shows a band centered near  $3.5\mu m$  (Smith *et al.*, 1963), a result again inconsistent with our observations.

### ESTIMATE OF THE WATER CONTENT

Using the observed profile of the  $2.85 \,\mu$ m absorption band, we can make an estimate of the amount of water present in the surface material. A key theoretical para-

meter for this purpose is the single scattering albedo,  $\tilde{\omega}_0$ , which we determine by employing a simple multiple scattering theory;  $\tilde{\omega}_0$  in turn is related to the product of the molar abundance of water, f, the absorption coefficient k of the hydration water, and the mean particle radius a. Since we have some knowledge of k and a, we can derive f.

We first obtain estimates of  $\tilde{\omega}_0$ . In so doing, we implicitly assume that Mars is covered by a top layer of dust, an assumption which is strongly supported by a variety of observations (Pollack and Sagan, 1969) and confirmed by Mariner 9 data. An estimate of the light reflected from the surface may be obtained from multiple scattering theories designed for optically thick clouds. The major difference between a powder and a cloud layer is that for the powder the particles are close together and shadowing effects are of some importance. As we will apply the theory to results at opposition, this difference will be less of a problem. Sample Mie scattering calculations indicate that for a wide range of plausible indices of refraction the ratio of forward to backward scattering is not very different from unity; the use of isotropic scattering is reasonable. Additional justifications are given below.

One can readily show that the geometric albedo, p, of an optically thick medium characterized by isotropic single scattering is given by:

$$p = \frac{1}{4} \tilde{\omega}_0 \int H^2 \, \mu d\mu, \qquad (2)$$

where  $\mu$  is the cosine of the angle of view, and H the well-known H function of radiative transfer. The integral in Eq. (2) is evaluated for a variety of values of  $\tilde{\omega}_0$  in tables given in Chandrasekhar (1950), using a six-point Gaussian quadrature. This integral is a function only of  $\tilde{\omega}_0$ , and hence  $\tilde{\omega}_0$  may be derived from Eq. (2) using the observed value of p.

We determined the value of  $\tilde{\omega}_0$  at three wavelengths, 2.5, 3.2, and  $0.35\,\mu\text{m}$ . As the reflectivity at  $2.5\,\mu\text{m}$  is essentially unaffected by the water band, the value of  $\tilde{\omega}_0$  at this wavelength provides information on the continuum absorption. On the other hand the value of  $\tilde{\omega}_0$  at  $3.2\,\mu\text{m}$  is affected by the absorption of the bound water. The reflectivity of Mars falls from the visible to the blue and approaches a nearly constant, low value in the near ultraviolet. Using the value of p at  $0.35 \mu m$  (Irvine *et al.*, 1969) we derive a value of  $\tilde{\omega}_0$ , which is close to the lowest value achievable for the Martian surface material.

Generalizing a formula given by Sagan and Pollack (1967), we can express  $\tilde{\omega}_0$  as:

$$\tilde{\omega}_0 = r + (1 - r) \exp\left(-2ka\right) = r + (1 - r) \exp\left[-2a(k_c + fk_H)\right], \quad (3)$$

where r is a term due to external reflections and the second term refers to transmission through the particle. The term r is simply equal to the value of  $\tilde{\omega}_0$  at  $0.35 \mu$ m. Diffraction effects are negligible because we are dealing with particles considerably larger than the wavelengths of interest. The factor k is the sum of a continuum value,  $k_c$ , and the value due to the water absorption,  $fk_H$ . We assume that the continuum value is independent of wavelength over the observed portion of the spectrum, an assumption consistent with a similar value of the reflectivity on each side of the absorption band. Solving Eq. (3) for (2ka), we obtain  $(2k_ca)$  from the value of  $\tilde{\omega}_0$  at  $2.5 \mu m$  and  $2(k_c + fk_H)a$ from the values of  $\tilde{\omega}_0$  at  $3.2 \mu m$ . Since  $(2k_Haf)$  depends upon the ratio of albedos in a logarithmic manner, the assumption of isotropic scattering should not seriously affect the result.

Finally, we determine f from the above information. Subtracting the continuum value of (2ka) from the value obtained within the band, we obtain  $(2k_Haf)$ . Analyses of the variation of temperature over a Martian day indicate that  $a \sim 100 \,\mu\text{m}$  (Morrison *et al.*, 1969; Neugebauer et al., 1971). Pollack et al. (1972) have recently determined the value of  $k_{H}$  as a function of wavelength for several common terrestrial rock samples, whose water content had been evaluated. The values of  $k_{H}$  for the separate samples agree quite well with one another, indicating that  $k_{H}$  is not a sensitive function of rock type. Using the value of  $k_{H}$  at  $3.2\mu m$  for their Lake County obsidian sample and the above value of a, we find that the fraction of bound water in the



FIG 5. The normalized spectrum is shown with a theoretical curve for  $100\mu$ m particles and 1% fraction by weight to bound water. The rather poor fit at 2.7 and 2.8 $\mu$ m is due to the fact that no allowance has been made for CO<sub>2</sub> absorption.

surface material is 1% by weight. The uncertainty in this value is about a factor of 3, due principally to the uncertainty in a, It is important to recognize that the large depth of the absorption feature implies a rather large fraction of the surface must contain bound water.

As a check on our above procedures, particularly the application of multiple scattering theory, we have calculated a theoretical reflectivity curve using Pollack et al. value of  $k_{\mu}$ . The theoretical calculation includes the effect of thermal emission, but does not allow for absorption by carbon dioxide, which is important between 2.7 and  $2.8\,\mu\text{m}$ . The theoretical curve is compared with the observed values in Fig. 5. The very good agreement between the two not only vindicates the application of the multiple scattering theory, but also shows the quantitative agreement between the absorption band we observed on Mars and the band expected for bound water. This agreement strengthensour identification of the Martian feature as being due to bound water.

We can obtain an independent estimate of the mean particle size from the value of the continuum albedo at  $2.5\,\mu$ m. As discussed above the value of  $\tilde{\omega}_0$  at  $2.5\,\mu$ m leads to an estimate of the product  $k_c a$ . For the six rock specimens studied by Pollack *et al.* (1972), which range in composition from extreme acidic to extreme basic, there is a factor of 30 spread in the value of  $k_c$  at  $2.5\,\mu$ m. Using these values, we obtain a value for the mean particle radius which ranges from 10 to  $300\,\mu$ m.

#### DISCUSSION

In this section we consider some of the biological and geological implications of our estimates of the bound water content and mean particle size of the Martian surface material.

We begin by estimating the amount of bound water present near the surface. The top 2mm of the surface experiences significant diurnal temperature changes (Morrison *et al.*, 1969). This top layer is the most interesting biological locale, as it can reach much higher temperatures near noon than lower layers. Using the above estimate of f, we can estimate that this top layer contains the equivalent of about  $20\mu$ m of bound water, an amount comparable to the largest amounts found in the atmosphere (Schorn *et al.*, 1967) Thus the bound water may be an important source of water for any hypothetical Martian organisms (cf. Sagan *et al.*, 1965).

The dust layer has been estimated to be between 1 m and 3km deep (Pollack et al., 1970a). If f remains constant with depth, there would be between 1 and  $3000 \text{ g/cm}^2$ of bound water in the total dust layer. In fact, if the bound water is present mostly as adsorbed water, f may be larger in the layers below the top few millimeters (Pollack et al., 1970b), so that the total amount of water may be even larger. These estimates may be compared with a value of about  $10^5 \text{g/cm}^2$  of water present in the oceans of the earth, which is believed to be the result of outgassing of water from the interior. Thus, there probably has been a substantial amount of outgassing of water from the interior of Mars, though perhaps not as much as for the Earth. This inference is consistent with the presence of volcanic features on Mars, which have been photographed by the Mariner 9 spacecraft (Masursky et al., 1972).

Among the volcanic features photographed by Mariner 9 are an array of large (~100 km diam) calderas. The production of such a caldera will not only release water vapor derived from the interior into the atmosphere, but during its incipient phases, it should also heat the local soil and release significant quantities of bound water into the atmosphere. The amount of bound water released by one of the large calderas could be as high as  $10^{18}$ g. In time. such water will become adsorbed in the surface or frozen out as permafrost in the subsurface or as ice in the poles. But the shorter term consequences may be significant. The amount of bound water released. if it were all temporarily in the Martian atmosphere. could be as large as the amount of water now in the terrestrial atmosphere. Moreover the release of water is localized and it is barely possible that such volcanic water may contribute to the formation of the sinuous dendritic channels found on Mars by Mariner 9, which are so difficult to understand in terms of the present Martian environment.

The large quantity of bound water estimated above means that the soil is a principal sink of the outgassed water, though of course some of the outgassed water may be deposited in permafrost layers beneath the surface or in the polar regions as ice. Thus, it is possible that the large reservoir of bound water is responsible for the seasonal changes in the atmospheric water content as well as for the long-term regulations of the content of atmospheric water vapor (Pollack et al., 1970a and b). While the above authors considered those effects for a limonite soil, their arguments should also apply to other types of mineralogy.

Above we have estimated that the surface soil layer contains about 1% bound water by weight. We have also argued that our spectra are inconsistent with that expected from limonite, which typically contains several tens of percent of water by weight. Thus, limonite can at most constitute several percent of the surface material of Mars. This is not inconsistent with the appearance of our spectrum (Fig. 4) near the  $3.2\mu$ m limonite bound water feature.

On the basis of our multiple scattering model, we estimated that the mean particle radius of the surface material lies between 10 and  $300\,\mu\text{m}$ , a value consistent with the determination of  $100\,\mu\text{m}$  made from the temperature measurements. As discussed by Sagan and Pollack (1969), this is in the size range which is most easily set into motion and transported by winds on Mars. and therefore the particle size expected to be ingreatest abundance in the surface layer.

#### SUMMARY

The major conclusions of this paper are

1. Within a factor of 3, the abundance of bound water is 1% by weight.

2. While most common minerals have water bands consistent with our observed spectra, limonite and hydrated polymeric carbon suboxide do not.

3. The mean particle radius of the surface soil lies between 10 and  $300\mu m$ .

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