Spatial Variations in the Spectral Properties of Bright Regions on Mars

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Much of the surface of Mars is covered by material that is bright and reddish at visible wavelengths and which is interpreted as chemically altered soil. High spatial resolution data from the ISM near-infrared imaging spectrometer on the Phobos 2 spacecraft have provided new information about the composition and compositional heterogeneity of this soil. Most bright soil exhibits a weak, narrow absorption feature near 2.2 μm, suggesting the presence of poorly crystalline phyllosilicate. The strength of the 3.0-μm H₂O absorption exhibits spatially coherent variations, with stronger absorptions occurring in specific geologic units. These variations may result from differences in the content of molecular water in soils developed on different deposits. Additionally, the position and shape of the 0.9-μm Fe²⁺ absorption indicate hematite to be present in bright soils covering many areas, including Tharsis. However, other large areas covered by bright soil, including much of Arabia, exhibit a deeper absorption at a longer wavelength, suggesting occurrence of ferric minerals other than hematite. These spectral heterogeneities indicate that bright martian soil is compositionally diverse and not completely homogenized by eolian activity. The compositional differences may result from different histories of chemical alteration of parent materials and their resulting soils, possibly including local effects of liquid water.

INTRODUCTION

The albedo and color of Mars at visible wavelengths define two fundamental surface units, bright red material and darker, relatively gray-colored material. The dark, relatively gray material is generally interpreted to contain relatively unaltered “rock,” but the bright material is interpreted as chemically altered “soil” (Toulin et al. 1977, Singer et al. 1979, Arvidson et al. 1989). The mineralogy and lithology of such altered soil are a record of interaction of the atmosphere and hydrosphere with crustal rocks. Thus, understanding the evolution of Mars’s surface environment and volatile inventory requires an accurate understanding of the mineralogy, lithology, and compositional variability of the bright reddish soil.

Compositional information for bright soil is available primarily from two sources, in situ measurements by the Viking landers and spectroscopic measurements acquired telescopically and from spacecraft. The X-ray fluorescence (XRF) experiments on the two Viking landers found nearly identical Fe-rich elemental abundances, comparable to weathered basalts, at the widely separated landing sites (Toulin et al. 1977). The gas chromatograph and mass spectrometer (GCMS) experiments indicated a total content of molecular H₂O in the range of 1–3% (Anderson and Tice 1979).

Near-infrared spectroscopic studies of bright soils have provided evidence for the presence of specific phases. A strong absorption at 3.0 μm is due to molecular H₂O (Houck et al. 1973, Pimentel et al. 1974). The major attributes of spectra covering the extended visible wavelength range include a positive slope between 0.35 and 0.75 μm, a reflectance peak near 0.75 μm, and shallow absorptions near 0.63 and 0.86 μm. These features can be explained if Fe in the soil is present largely as pigmentary (nanophase) hematite with a small amount of more crystalline (bulk) hematite (Singer et al. 1979, Morris et al. 1989, 1990, Morris and Lauer 1990, Bell et al. 1990, Bell 1992). No clear spectral evidence for ferric minerals other than he-
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Matite has yet been identified (Bell 1992). Also, no strong absorption features attributable to crystalline silicates have been observed in bright soils, suggesting that their silicate fraction may be largely amorphous (Singer 1982). Nevertheless, several investigators have reported very weak absorptions at 2.2–2.4 μm that may indicate hydroxylated minerals (Singer et al. 1985, Erard 1992, Murchie et al. 1992a, Bell and Crisp 1993) or bicarbonate- or bisulfate-bearing minerals such as scapolite (Clark et al. 1990b) occurring in small amounts or in poorly crystalline form. At the spatial resolution and precision of telescopic data bright soils appear relatively homogeneous, and the major reported heterogeneities not possibly related to atmospheric effects may be variations in the abundance of bulk hematite (Bell 1992).

In contrast, the dark relatively gray regions exhibit 1- and 2-μm absorptions due to Fe-containing pyroxene (e.g., Singer et al. 1979, Erard et al. 1991, Mustard et al. 1993b). A 3.0-μm H₂O absorption is also present in these regions, but is generally weaker than in bright soils (Pimentel et al. 1974, Erard et al. 1991). These results are consistent with darker, grayer regions typically being composed of a relatively anhydrous basaltic lithology (Singer et al. 1979, Erard et al. 1991, Mustard et al. 1993a,b).

The similarity in spectral properties of bright soils at visible wavelengths, the nearly identical elemental abundances at the two Viking lander sites, and observations of global dust storms have led to the paradigm that bright soils represent a surface layer of chemically altered "dust" that is globally homogenized by wind and unrelated to underlying geologic units (e.g., Toublon et al. 1977, Arvidson et al. 1989). It has even been suggested that compositional variations over large areas can be explained by the mixing of only two components, the darker, grayer material and homogenized bright soil (Arvidson et al. 1989). We have tested these hypotheses using data from the near-infrared imaging spectrometer ISM on the Phobos 2 spacecraft, by identifying and mapping differences in absorptions associated with H₂O, Fe, and hydroxyl in the bright soils. ISM data are ideally suited to this analysis because of their high spectral and spatial resolution, high signal-to-noise ratio, and image format. These characteristics permit identification of subtle absorption features, and mapping of their positions and strengths for comparison with surface geology. We have found that bright soils are compositionally heterogeneous, that some of these heterogeneities are correlated spatially with surface geologic units, and that "globally homogenized dust" is an overly simplistic and largely inaccurate description of bright soils.

Analytical Procedures

ISM returned nine images or "windows," each approximately 24 × 120 pixels, with a pixel size of 22 km. These

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**FIG. 1.** ISM spectra of the bright region Sinai Planum, for comparison with and without atmospheric absorptions removed from the data. Downward-pointing arrowheads denote locations of major absorptions due to Fe, H₂O, and metal–OH bonds in major rock-forming minerals that are outside strong atmospheric absorptions; upward-pointing arrowheads denote locations of the strong atmospheric absorptions.

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cover ~20% of the martian equatorial region, including parts of Syrtis Major Planum, Isidis Planitia, Arabia, Valles Marineris, the Tharsis volcanoes and plateau, and Amazonis Planitia (Bibring et al. 1990). For each pixel a spectrum was measured by PbS detectors, composed of 64 channels each from the first grating order (1.64–3.16 μm) and the second order (0.76–1.51 μm) (Fig. 1). Each window has an internal signal-to-noise ratio ranging from 400:1 to 1300:1 in most channels, and overlapping windows demonstrate stability of different channels generally within ~0.3%. Consequently significant near-infrared spectral features can be analyzed across the whole data set.

Data Reduction and Calibration

The procedures used here for reduction and calibration of ISM data have been described previously (Bibring et al. 1990, Erard et al. 1991, Mustard et al. 1993b). The first- and second-order spectra each consist of 32 odd and 32 even channels, of which only the even channels were used because of their greater stability. The data were corrected for dark current measured in-flight, gain measured on-ground, response of the detectors to temperature, and order overlap, and were divided by the solar spectrum. The signal was converted to radiance factors by scaling DN levels measured on Phobos to a spectral model of that satellite (Erard et al. 1991, Mustard et al. 1993b). Subsequently the calibration was refined using additive and multiplicative corrections based on a spectral
model of Mars (Mustard et al. 1993b). Comparison of calibrated spectra of a variety of regions of Mars with corresponding telescopic spectra demonstrates both a high degree of correlation and a greater precision of the ISM spectra (Bell and Mustard 1993). In a single restricted region near Ascreaus Mons, several ISM channels near 1 \( \mu m \) became saturated due to high surface reflectance; these data were identified and not included in this study.

The ISM data include spectral effects of both atmospheric and surface components. Minerals in the soil containing \( \text{H}_2\text{O} \) and structural \( \text{OH} \) could potentially exhibit absorptions due to \( \text{H}_2\text{O} \) at \( \sim1.9 \) and \( \sim3.0 \mu m \); \( \text{H}_2\text{O} \) and metal-\( \text{OH} \) at \( \sim1.4 \mu m \); and metal-\( \text{OH} \) at 2.2–2.3 and 2.7–2.8 \( \mu m \) (Farmer 1974). Ferric oxides and oxyhydroxides exhibit absorptions centered at 0.85–0.98 \( \mu m \), depending on structure of the \( \text{Fe}^{3+} \) site (Sherman et al. 1982, Morris et al. 1985). These ferric phases can be discriminated from ferrous iron-containing pyroxenes and olivines, which also exhibit absorptions at \( \sim1 \mu m \), by the shape and position of the absorption, by the pyroxenes’ accompanying absorption near 2 \( \mu m \), and in the case of high-Ca pyroxene by an accompanying absorption at 1.15 \( \mu m \) (Hunt and Salisbury 1970, Cloutis and Gaffey 1991).

Figure 1 illustrates that several of these wavelength ranges also include absorptions due to atmospheric components, primarily absorptions due to \( \text{CO}_2 \) at 2.0 and 2.7 \( \mu m \) and closely spaced \( \text{CO}_2 \) and \( \text{H}_2\text{O} \) absorptions near 1.4 \( \mu m \).

In order to estimate the spectral properties of surface materials alone, the atmospheric absorptions were “removed” using previously described techniques (Erard et al. 1991, Rosenqvist et al. 1992). A spectral model of the atmosphere was constructed based on measurements from the summit and base of Pavonis Mons, which appear homogeneous at wavelengths outside important atmospheric absorptions. This model spectrum was then removed from the data, with strengths of the atmospheric absorptions scaled to the 2.0-\( \mu m \) \( \text{CO}_2 \) absorption. Accuracy of this procedure depends on accurate modeling of the shapes and relative strengths of the atmospheric absorptions; inadequacies in either can create “artifacts” on the surface. For this reason we have concentrated analysis on the \( \sim0.9-\mu m \) Fe absorption, the 3.0-\( \mu m \) \( \text{H}_2\text{O} \) absorption, and 2.2 to 2.3-\( \mu m \) metal-\( \text{OH} \) absorption, which lie outside the strongest atmospheric absorptions.

### Study Areas

Bright soils occupy two major geologic settings in the regions observed by ISM. First, they occur on cratered and volcanic terrains as a surface layer thin enough not to obscure hundreds-of-meters-scale topography. This mode of occurrence at the Viking Lander sites is characterized by a patchy cover of fine-grained sediment (Binder et al. 1977, Mutch et al. 1977). Second, they occur as eroded, kilometers-thick deposits which unconformably overlie and obscure preexisting surfaces. One such unconformable deposit partly infills the Candor Chasma region of Valles Marineris (arrows, Fig. 2a), forming deeply eroded plateaus of “layered materials” (Scott and Tanaka 1986). Another major unconformable deposit is located in the western Arabia–Oxia Palus region (Schultz and Lutz 1988). This deposit consists of a mantling at least several hundred meters thick, whose deep erosion is evidenced by inverted topography of large craters and by irregular depressions interpreted as deflation hollows (arrows, Fig. 2b).

In our analysis we used seven of the nine ISM windows, including the regions covering and adjacent to the two unconformable deposits described above. These windows sample a representative variety of environments in which bright soils occur, including both the unconformable deposits and the thin layers on Noachian-, Hesperian-, and Amazonian-aged cratered and volcanic terrains. They also include sites discussed in previous analyses of ISM data, specifically Valles Marineris and eastern Tharsis (Erard et al. 1991, Murchie et al. 1992b), the Syrtis Major Planum–Isidis Planitia region (Erard et al. 1991, Mustard et al. 1993b), and the Olympus Mons region (Fischer and Pieters 1993). All of the windows were observed at low phase angles (<20°) that vary by <5° in any given window, and at a local time of 0900–1500. Also, none of these windows exhibits evidence for significant water–ice clouds. Due to the backscattering and absorption of light by ice aerosols, such evidence would include patches having negative (“blue”) spectral slope and an enhanced 3.0-\( \mu m \) \( \text{H}_2\text{O} \) absorption, occurring especially in geographic settings where thin clouds are common, such as Tharsis (cf. McEwen 1992).

### Data Analysis

In this investigation we have utilized a two-pronged approach. First, we surveyed the bright soils observed by ISM to determine the presence of absorptions due to Fe, metal-\( \text{OH} \), and \( \text{H}_2\text{O} \) and to assess variability in their positions and strengths. This was done initially by examining individual spectra. Typically each spectrum was measured over a uniform region consisting of 6–9 ISM pixels. Uncertainty in absolute reflectance is of the order of \( \pm10\% \). However, relative reflectances are more germane to the identification and validity of spatial variations in spectral properties. The formal uncertainty in relative reflectances at different wavelengths derives from three sources, random noise in the data, real spectral variability, and systematic errors between windows which are typically \( \pm0.3\% \). The real variability is very small because of the uniformity of the measured sample areas, and effects of random noise were minimized through the averag-
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FIG. 2. (a) Candor Chasma and Ophir Chasma regions of Valles Marineris. Arrowheads denote the high-albedo layered deposits partially infilling Candor Chasma (Viking image 608A74). (b) Partly eroded unconformable mantling deposit in the Oxia Palus-western Arabia region. Arrowheads denote landforms indicative of deep erosion, including inverted topography of a large crater and depressions interpreted as deflation hollows (Viking image 615A36).

ing over multiple pixels. Consequently, outside of wavelengths at which important atmospheric absorptions have been removed, formal uncertainties in relative reflectances are of the order of ±0.3–0.4%, which is similar or smaller in scale than the symbols used in plotting spectra in this paper.

Spatial variations in spectral properties were then mapped, using "parameterized" images depicting the major sources of spectral variability. Reflectance was calculated as the average at three wavelengths near 1 μm. The center of the Fe absorption was estimated using a cubic spline fit to the data. Absorption depths were calculated by comparing reflectances at absorption centers to the continua estimated from reflectances at the wings of the absorptions. The metal–OH absorption was found by examination of individual spectra to be centered near 2.20 μm, so its depth was calculated assuming a constant absorption center and fixing the wings of the absorption at 2.15 and 2.30 μm. These latter wavelengths approximate the edges of the 2.2-μm metal–OH absorption in a variety of minerals (Clark et al. 1990a), yet lie outside the strongest portions of atmospheric absorptions due to CO₂ and CO. In calculating depth of the 3.0-μm H₂O absorption, a flat continuum was extrapolated from the reflectance at 2.5 μm. The assumption of a flat continuum in this wavelength region is warranted from analysis of a wide variety of ISM spectra covering bright and dark regions (Mustard et al. 1993b) as well as clouds observed in regions not included in this study (Erard et al. 1993). In each case, the negative spectral slope typifying many regions of Mars at shorter wavelengths becomes negligible at wavelengths longer than about 2.5 μm. Spectral parameterization was duplicated using data with and without atmosphere removed, to demonstrate reproducibility of measured properties.

Second, we undertook a detailed study of the two bright regions including and immediately surrounding the unconformable deposits, to investigate the relationship of spectral heterogeneities with highly varied surface physical geology. This was accomplished by comparing spatial variations in the spectral properties of bright materials with the spatial distribution of previously mapped geologic units (Scott and Tanaka 1986, Schultz and Lutz 1988, Murchie and Izenberg 1991, Witbeck et al. 1991). This latter part of our study does not replace a thorough investigation of the correspondence of spectral heterogeneities revealed in the whole ISM data set with surficial geology and visible-wavelength color properties. However, the scope of such an investigation is too broad to be covered adequately here and will be addressed in a future paper.
RESULTS

Figure 3a illustrates how “typical” bright soils differ spectrally from “typical” dark materials, by comparing selected wavelength regions of spectra from the dark region Syrtis Major Planum and the adjacent bright region Libya Montes. For clarity in illustrating spectral features due to surface materials, atmosphere-removed data are used here. As reflectance increases from Syrtis Major into the bright soil, several additional spectral variations are evident. First, the Fe absorption near 1 μm exhibits its well-documented shift to wavelengths shorter than 0.9 μm, as it is dominated more by ferric iron and less by ferrous iron-containing pyroxene (e.g., Singer et al. 1979, McCord et al. 1982, Mustard et al. 1993b). Second, the slope of the spectrum at longer wavelengths becomes increasingly less negative (reflectance decreases less toward longer wavelengths). Negative spectral slopes are common in martian dark regions and are generally attributed to a thin coating of bright ferric material occurring as either airfall or a stable coating (Singer and Roush 1983, Mustard et al. 1993, Fischer and Pieters 1993). Most bright regions, in contrast, exhibit relatively flat spectral slopes (Murchie and Mustard 1993). Third, the depth of the 3.0-μm H₂O absorption increases. This absorption is not shown directly in Fig. 3a because its depth is much larger than the scale of the plot. Finally, a very weak, narrow absorption centered between 2.20 and 2.25 μm appears in the bright soil. This latter feature, which is consistent with metal-OH, was documented for the first time by Erard (1992) and Murchie et al. (1992a). It has also been recognized recently using telescopic imaging spectroscopy (Bell and Crisp 1993).

The weak, narrow absorption at 2.2 μm is located approximately on the long-wavelength wing of the strong 2.0-μm absorption due to atmospheric CO₂. Hence it is important to demonstrate that this weak feature can be recognized in data from which no model atmosphere has been removed, as well as in data from which atmosphere has been removed. As explained in the preceding discussion of data analysis, the strength of this absorption exhibits the same spatial variations in both versions of the data. Figure 3b shows that the absorption itself is also discernable in individual “raw” spectra before atmospheric removal. In Fig. 3b, representative “bright” and “dark” spectra from Fig. 3a are illustrated, for comparison with and without the atmosphere removed. In the wavelength regions shown here, the largest difference between the two versions of the data is at the wing of the CO₂ absorption. In the raw spectrum of Syrtis Major Planum, reflectance in the 2.1-2.4-μm region peaks near 2.20 μm, beyond which the CO₂ absorption is negligible and spectral shape is dominated by the broad, shallow 2-μm pyroxene absorption and a superimposed negative
spectral slope. In contrast, in the raw spectrum of Libya Montes reflectance peaks at 2.36 μm; the wing of the CO₂ absorption exhibits a weak but distinct shoulder due to the superimposed 2.2-μm metal–OH absorption.

Clark et al. (1990b) reported an additional weak, narrow absorption in surface materials located at 2.35 μm, interpreted to be due to the mineral scapolite. Identification of this absorption is quite difficult, though, because this wavelength also exhibits an absorption of comparable magnitude due to atmospheric CO. We are unable to verify the occurrence of the “scapolite” absorption in most regions using atmosphere-removed data. Either the absorption is too weak to detect, or it is too narrow to resolve in ISM data, or it is attributable to atmospheric rather than surface components (cf. Encrenaz and LeLouch 1990; Rosenqvist et al. 1992). Erard et al. (1991) investigated this feature in more detail, and concluded that it is absent or unresolvable by ISM except possibly in the Lunae Planum region and in restricted parts of Valles Marineris.

Within bright regions, ISM data reveal spatial variations in properties of the Fe, H₂O, and metal–OH absorptions. These are discussed in detail below.

3.0-μm H₂O Absorption

The strength of the 3.0-μm H₂O absorption exhibits spatially coherent variations approaching 13% in some areas (Fig. 4). There is a number of possible explanations for these variations, which can be evaluated based on our knowledge of the martian surface and the behavior of the ISM data. Three-micrometer absorption strength is not correlated with elevation or atmosphere path length, so the variations are not an atmospheric effect or the result of inaccurate atmosphere removal. Bright soils in general have a stronger absorption than do dark regions, but there are several discrete areas including parts of Valles Marineris and Arabia (arrowhead Fig. 4A; left of arrowheads, Fig. 4B) where absorption strength is enhanced without an increase in albedo. Thus the variations cannot be explained simply as the result of either a calibration artifact or mixing of homogeneous bright and dark components (cf. Arvidson et al. 1989). Neither can the variations in absorption strength be realistically attributed to transient water–ice clouds. Enhancements in the 3.0-μm absorption do not have a consistently negative spectral slope, as would be expected if they were caused by ice clouds. Furthermore overlapping windows covering Valles Marineris, acquired on different days and at different local times, show the enhanced absorption in the same places. The variations in absorption strength cannot be attributed to outcrops of ground ice in an otherwise homogeneous soil, because ice is unstable at the equatorial latitudes observed by ISM (Clifford and Hillel 1983). The simplest explanation for the regions of enhanced 3.0-μm absorptions is differing soil lithology. The most obvious difference would be the content of molecular H₂O, although particle size differences could also contribute to variations in absorption strength.

Based on the strength of the 3.0-μm H₂O absorption, two general units of bright soils can be defined: “normal bright soils” with a typical 3.0-μm absorption, covering ~59% of the bright regions in Fig. 4, and “hydrated bright soils” with a consistently stronger 3.0-μm absorption, covering ~15% of the bright regions in Fig. 4. In the regions shown in Fig. 4, occurrences of these two spectral groups correlate spatially with independently identified surface features. Normal bright soils cover areas with surface morphologies indicative of only a thin covering of bright soil on volcanic and cratered terrains, including much of Tharsis and Arabia. The same regions are typified by low thermal inertias suggestive of loose, unconsolidated “dust” at least several centimeters thick (Palluconi and Kieffer 1981; Christensen 1986, Christensen and Moore 1992). In contrast, hydrated bright soils correspond with the deeply eroded unconformable deposits in Candor Chasma and western Arabia. The deposit in western Arabia has also been distinguished from normal bright soil in eastern Arabia by its high thermal inertia and by its darker red color (Presley and Arvidson 1988, Soderblom 1992, Christensen and Moore 1992). This correlation between independently identified spectral variations and surface physical attributes indicates real differences between bright soils in different geologic settings.

2.2-μm Metal–OH Absorption

The depth of the weak absorption at 2.20–2.25 μm also exhibits spatially coherent variations of up to ~1.1% (Fig. 4). This weak spectral feature is present in most bright soils, and is subdued or absent in all dark regions that we have investigated. The systematics of this absorption provide evidence for its origination as a surface feature. As in the case of the 3.0-μm absorption, brighter areas generally have stronger absorptions. However there are significant enough departures from this relationship (e.g., in Candor Chasma, Fig. 2a) that variations in absorption strength cannot be the result solely of either a calibration artifact or mixing of homogeneous bright and dark components. Variations in absorption strength are not correlated with atmospheric path length or elevation and are therefore not attributable to a weakly absorbing atmospheric species. The general positive correlation between absorption strength and brightness is inconsistent with the effects of a narrow, saturated atmospheric absorption. Rather, this correlation is a strong indicator of origination due to a mineral phase (or phases) typical of bright soils.
FIG. 4. Parameterized ISM images calculated using data from which the atmosphere has not been removed. (A) The Valles Marineris region; the arrowhead denotes Candor Chasma. (B) The Arabia region; the arrowheads denote the approximate eastern limit of the partly eroded mantling shown in Fig. 2b, which fills the high-albedo western part of the window. Curved lines are topographic contours spaced at 1-km elevation intervals. Top: Reflectance at 1 μm. Blue denotes lower reflectances and red higher reflectances. Middle: Depth of the 3.0-μm H₂O absorption, relative to a continuum extrapolated from 2.5 μm. Blue denotes deeper absorptions and red weaker absorptions. Bottom: Depth of the 2.2-μm absorption. Blue denotes deeper absorptions and red weaker absorptions.
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FIG. 5. Depths of the weak 2.2-μm absorption in materials having different 3.0-μm H₂O absorption strengths, measured from atmosphere-removed data. Three regions, where different spectral units are intermixed are shown separately: “typical” dark and bright materials in Syrtis Major Planum and Libya Montes, hydrated and normal bright soils in Arabia, and hydrated and normal bright soils in eastern Tharsis–Candor Chasma. For each increment of 0.01 in strength of the 3.0-μm absorption, this plot shows the mean and standard deviation of depth of the 2.2-μm absorption for pixels having a 3.0-μm absorption strength within that increment.

Two relationships are particularly critical to characterizing the origin of the 2.2-μm absorption: its strength exhibits a regionally variable relationship with strength of the 3.0-μm absorption, and its strength varies between adjacent geologic settings having soils with comparable reflectances. In comparing typical bright soils and dark regions, for example at the boundary of Isidis Planitia and Syrtis Major Planum, strengths of the 2.2- and 3.0-μm absorptions are positively correlated (Fig. 5). In contrast, within and around the hydrated bright soils in Candor Chasma and western Arabia, the strengths of the 2.2- and 3.0-μm absorptions are inversely correlated (r ~ -0.7) (Figs. 4, 5). This inverse relationship occurs because the 2.2-μm absorption is significantly weaker in hydrated bright soils covering the unconformable deposits than in surrounding normal bright soils. The dependence of systematics of these absorptions on physical geology demonstrates that the 2.2-μm absorption is real and characteristic of many bright soils, but that optical surfaces of hydrated bright soils are lacking in whatever phase(s) cause the absorption.

0.9-μm Fe Absorption

Bright soils also exhibit variability in the position and shape of their Fe absorptions near 0.9 μm. We surveyed the data set for relatively “pure” normal bright soils, i.e., those lacking “contamination” by dark material. These were recognized based on high albedo, lack of an absorption at 2 μm characteristic of pyroxene, and lack of broadening of the Fe absorption beyond 1.15 μm which could be indicative of olivine or high-Ca pyroxene. Representative spectra fitting these criteria are shown in Fig. 6a. The absorption center for each was estimated by fitting the data with a cubic spline. The position of the Fe absorption varies between 0.85 and 0.92 μm, but all the spectra exhibit the weak, narrow 2.2-μm absorption typical of normal bright soil. The latter observation suggests that variations in the Fe absorption do not result from a component which masks the very weak 2.2-μm feature.

Among normal bright soils judged to be free of dark “contaminants,” there are three geographic regions which exhibit distinct absorptions near 0.9 μm (Fig. 6b). Previous studies have provided some evidence for these differences, but did not fully resolve the distinct Fe absorptions. Bright soils throughout the Tharsis plateau plains and westward into Amazonis exhibit a spatially uniform ferric iron absorption 2–4% deep and centered at 0.84–0.86 μm. This region has been distinguished previously based on its very low thermal inertia, and interpreted as being covered by a widespread layer of dust-storm fallout (Pallucioni and Kieffer 1981, Christensen 1986, Christensen and Moore 1992). In contrast, soils at high elevations on some of the Tharsis Montes volcanoes, particularly Ascaraeus Mons, exhibit an absorption center shifted longward to 0.87–0.88 μm. In earlier analysis of ISM data, Bibbring et al. (1990) recognized that spectral properties near 1 μm distinguish Ascaraeus Mons from the surrounding region, although they did not identify position of the Fe absorption as being responsible for this difference.

In contrast, the absorption exhibited by bright soils throughout Arabia is spatially uniform but radically different from that in Tharsis, 5–7% deep and centered near 0.92 μm. As in the case of Tharsis, Arabia is also covered by material having a very low thermal inertia, interpreted previously to consist of fallout from dust storms. The distinctiveness of the 0.9-μm absorption in Arabia had been recognized by McCord et al. (1982) and attributed to “mafic minerals” in the soil.

DISCUSSION

Ferric Mineralogy of Bright Soils

Variations in the 0.9-μm Fe absorption may be attributable to particle size, lithologic, or mineralogic differences between bright soils. Bright soils throughout the low thermal inertia, presumably “airfall-covered” region of Tharsis exhibit a spatially homogeneous absorption centered near 0.85 μm. This band center is consistent with previous interpretations that hematite (α-Fe₂O₃) is the major Fe phase in these soils. The homogeneity of this
region also supports the idea of well-mixed soil such as dust-storm fallout, at least regionally throughout Tharsis.

A smaller particle size or substitution of cations such as Al$^{3+}$ into the hematite crystal lattice could shift this absorption to longer wavelengths by as much as 0.03 $\mu$m (Morris et al. 1985, 1992), possibly accounting for the longer-wavelength absorption high on Ascreaus Mons. There is no obvious reason to expect hematite of different composition on Ascreaus Mons, but the particle size of hematite there may be different. For example, bright soils throughout Tharsis may consist of well-mixed airfall, but with particles retained at the highest elevations on Ascreaus Mons having a finer size than particles which settled out at lower elevations. In this case the difference between the Fe absorptions on Ascreaus Mons and the surrounding plains might result from simple physical differences in their soils, and not from mineralogical differences.

The absorption center in Arabia near 0.92 $\mu$m cannot be reconciled with a purely hematitic Fe mineralogy by invoking particle size differences or cation substitution. McCord et al. (1982) attributed the distinctiveness of bright soil in Arabia to intermixture of mafic minerals. Presumably this refers to intermixture of pyroxene like that occurring in darker, grayer regions (Singer et al. 1979, Erard et al. 1991, Mustard et al. 1993a,b). In Fig. 7, the spectrum of Arabia is compared with a transsect of spectra from parts of Tharsis having different reflectances. Based on the homogeneity of bright soils in Tharsis, these latter spectra can be assumed to represent mixtures of a nearly uniform bright component with the underlying darker, grayer, pyroxene-bearing component. Compared to portions of Tharsis having a comparable brightness, Arabia exhibits a deeper, more symmetric Fe absorption at a longer wavelength, as well as a less negative spectral slope. This comparison does not support the hypothesis

![Graphical representation of the text content](image-url)
that bright soils in Arabia consist of a mixture of bright and pyroxene-bearing dark soils like those occurring in Tharsis.

Alternatively, we interpret the Fe absorption in Arabia to originate from a lithology not represented in Tharsis. In this case, bright soil in Arabia would have evolved along a different geochemical pathway than bright soil in Tharsis. It clearly would not represent dust-storm fallout from a single, well-mixed reservoir that also supplied airfall to Tharsis, and it may even have originated from a different parent material.

One possibility is that bright soils in Arabia represent a mixture of a Tharsis-like bright, hematitic component plus lower-reflectance pyroxene, but with textural properties resulting in a less negative spectral slope and a different bandshape than in intermediate-albedo parts of Tharsis. For example, Morris et al. (1993) have identified natural mixtures of hematite and pyroxene whose spectra exhibit a 0.9-μm Fe feature like that in Arabia, without a negative spectral slope or a significant 2-μm absorption. Bell and Morris (unpublished data) have found that a laboratory mixture of subequal amounts of hematitic volcanic cinders with pigeonite pyroxene will reproduce the same spectral attributes. However, we do not favor the hypothesis that bright soil in Arabia consists of such a mixture, because of the distinctive reflectance properties of the laboratory mixtures and bright soil in Arabia. In the laboratory, intermixtures of a sufficient amount of pyroxene to shift hematite’s absorption center from 0.85 to 0.92 μm also decreases reflectance at wavelengths outside the absorption (e.g., at 1.8 μm) by a factor of ~30%. In contrast, the 1.8-μm reflectance of bright soil in Arabia is within several percent of that in the presumably hematitic soil covering the brightest parts of Tharsis.

We suggest instead the presence of an additional high-reflectance ferric mineral in the bright soil in Arabia, which is either absent or present in lesser amounts in bright soils in Tharsis. Figure 8 compares representative spectra of Tharsis and Arabia with a variety of ferric minerals. The Fe absorption in Tharsis is well fit by a weak hematite absorption centered at 0.85–0.86 μm. Arabia exhibits an absorption center near 0.92 μm, but also an inflection near 0.86 μm. This configuration could result from intermixtures of hematite with another ferric phase having an absorption centered near 0.92 μm. Minerals exhibiting an absorption at this wavelength include goethite (α-FeOOH), ferrihydrite (Fe₁₀O₂·H₁₈), and jarosite ((K, Na, H₂O)Fe₃(SO₄)₂(OH)₆) (Fig. 8) (Hunt et al. 1971a, Sherman et al. 1982, Morris et al. 1985, Clark et al. 1990a). Hematite can form by gas–solid weathering of mafic minerals under ambient martian conditions; all else being equal, the later minerals would have formed on Mars in more water-rich environments (Goodying 1978, Burns and Fisher 1990, Burns 1992, Goodying et al. 1992). Goethite and ferrihydrite both age to form hematite by dehydration and hence are attractive possibilities to produce the absorption observed in Arabia. Occurrence of either mineral would be evidence for a regional differences in the style of chemical weathering of bright soils between Tharsis and Arabia.

**Water and Hydroxyl in Bright Soil**

The absorption centered at 2.20–2.25 μm constitutes important evidence for the mineralogy of bright soil. This feature is attributed to metal–OH bonds in hydroxides, oxyhydroxides, or hydroxylated silicates. A narrow absorption centered at ~2.2 μm indicates Al–OH bonds, whereas one at ~2.3 μm generally indicates Mg–OH or Fe–OH (Hunt et al. 1971a, Clark et al. 1990a, Mustard 1992). Among the minerals exhibiting absorptions centered near 2.20 μm, only a subset are geochemically plausible major constituents of altered martian soil (cf. Goodying et al. 1992). Of this subset, aluminous phyllosilicate best explains the position of the weak martian absorption. The observed broadening toward longer wavelengths (Figs. 3 and 6a) would be consistent with an additional component of a hydroxylated ferromagnesian mineral. The weakness of the absorption could result from poor crystallinity (cf. Singer 1982) or dilution by other phases (Orenberg and Handy 1992, Bishop et al. 1993a).

The inverse correlation of strengths of the H₂O and metal–OH absorptions within and around hydrated bright soils is important evidence for the origin of these soils. Assuming that our interpretation of the composition of

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**FIG. 8.** ISM spectra of bright soils in Tharsis and Arabia, compared with laboratory spectra of ferric minerals whose 0.9-μm absorptions are located at the same wavelengths as those exhibited by bright soils (J. Bishop, unpublished data).
normal bright soils is correct (phylllosilicates + hematite ± another ferric mineral), at least three models can potentially explain this relationship (Fig. 9). Each envisions modification of parent materials by different processes, but all have in common a role of liquid water. First, the hydrated bright soils may consist of “palagonite,” silicate glass particulates with rims that have been oxidized and hydrated by interaction with aqueous solutions. Comparison of the spectral properties of crystalline phylllosilicates and palagonites have shown that palagonites typically have weaker 2.2-μm absorptions (e.g., Singer 1982, Orenberg and Handy 1992). The weaker absorption has been attributed to poorer crystallinity of phylllosilicate in the palagonite (Singer 1982), but may in some cases result from a mineralogical difference. However, under dehydrating conditions, such as the ambient martian surface environment, palagonite retains more water and has a stronger 3.0-μm H₂O absorption than does crystalline phylllosilicate (Rous 1987, Murchie et al. 1992a). This model is attractive for explaining hydrated bright soil in Candor Chasma, where there is geomorphic evidence for contemporaneous volcanism and standing water (Witbeck et al. 1981, Lucchitta et al. 1992). Alternatively, the hydrated bright soils may consist of lithic particles that were oxidized and hydrated by hydrothermal processes (cf. Morris et al. 1993), yet lack crystalline phylllosilicates. A third possibility is motivated by observation of induced soils or “duricrust” at the Viking Lander sites (Binder et al. 1977, Mutch et al. 1977). The Viking Lander XRF experiments found that, compared with unconsolidated soil, the duricrust is enriched in sulfur presumably occurring as sulfate (Toulin et al. 1977). Water-bearing sulfate salts such as gypsum exhibit strong H₂O absorptions (Hunt et al. 1971b, Clark et al. 1990a), and phylllosilicates containing sulfate salts in interlayer sites exhibit stronger H₂O absorptions than do the same materials lacking these inclusions (Bishop et al. 1993b). In this model, hydrated bright soil consists of normal bright soil plus lithic particles, cemented into duricrust by hydrated sulfate salts mobilized in thin films of liquid water located below the soil surface. The duricrust was later exposed by erosion of overlying, unconsolidated sediment. In this case the 2.2-μm absorption would be masked by dilution of normal bright soil by additional components in the duricrust.

An origin as duricrust has previously been proposed for soil in western Arabia that is “dark red” at visible wavelengths, to explain the high thermal inertia and relative lack of eolian redistribution of this material compared with normal bright “dust” in adjacent terrains (Arvidson et al. 1982, Presley and Arvidson 1988, Soderblom 1992, Christensen and Moore 1992). Spatially, this dark red soil corresponds with what we term here as “hydrated bright soil” covering the unconformable deposits. Exposure of
duricrust in this region is consistent with the deep erosion of the unconformable deposits, which would have preferentially removed more poorly consolidated soils (Fig. 2b).

Comparison with Chemical Measurements of Altered Martian Surface Material

Our interpretations about the mineralogy of bright martian soils are supported by laboratory studies of SNC meteorites, thought to be derived from Mars. Chemically altered pockets in these meteorites contain aluminous phyllosilicate and Fe-rich smectite with very poor crystal ordering (Gooding and Muenow 1986, Gooding et al. 1991, Treiman and Gooding 1991, Gooding 1992). This combination of phases is consistent with that inferred from the position and strength of the 2.2-µm absorption, i.e., poorly crystalline aluminous phyllosilicate and a hydroxylated ferromagnesian mineral. Furthermore, these altered pockets also contain as accessory phases several minerals suggested as possible explanations for spectral heterogeneities in bright soils. Gypsum (CaSO₄· nH₂O) has been identified as the major salt phase in SNC meteorites, and hematite and ferrihydrite have both been identified as ferric-bearing phases (Gooding et al. 1991, Treiman and Gooding 1991). Gooding (1992) has also found that an admixture of these minerals can account for the elemental abundances in bright soil measured by the XRF experiments on the Viking landers.

Our results are also consistent with, although not uniquely indicative of, the presence of phases suggested to account for results from the Viking biology experiments. The labeled release (LR) experiment detected ¹⁴C evolved from decomposition of a labeled nutrient broth added to bright martian soil; decomposition activity was reduced considerably by heating the soils to 160°C. These behaviors are well simulated in the laboratory by Fe-rich montmorillonite clays (Banin and Rishpon 1979). The pyrolytic release (PR) experiment detected fixation of ¹⁴CO or ¹⁴CO₂ by dry bright soil, which was also simulated in the laboratory using Fe-rich montmorillonites (Hubbard 1979, Banin and Rishpon 1979). The weak, narrow absorption detected in bright soils at 2.2 µm is indicative of some aluminous phyllosilicate, and Fe-rich montmorillonite is a strong candidate based on its spectral resemblance to bright soil in the extended visible wavelength region (Banin et al. 1985, Bishop et al. 1993a). However, the ISM data could also be fit by a different aluminous phyllosilicate such as illite.

Comparison to Previous Compositional Models of Martian Soil

Two major lithologies have been proposed as compositional analogs to bright martian soil. Ferric-rich montmorillonite, proposed by Banin and Rishpon (1979), Banin et al. (1985), and Bishop et al. (1993a), consists of a crystalline phyllosilicate with nanophase inclusions of ferric oxide (Bishop et al. 1993a). Palagonite, and in particular the finest-grained fraction of Hawaiian palagonite proposed by Singer (1982). Morris et al. (1990), and others, consists of amorphous to poorly crystalline silicate with inclusions of nanophase ferric oxide (Morris et al. 1990). Both proposed lithologies have spectral properties in the extended visible wavelength range consistent with bright soil. However, when measured under ambient terrestrial conditions, palagonite more closely resembles Mars at near-infrared wavelengths than does crystalline phyllosilicate, because of the relative weakness of its absorptions at 1.4 and 1.9 µm. These features are either extremely weak or absent in the spectrum of bright martian soil.

Our results are partly consistent with both models, but uniquely support neither because the difference between them is largely semantic. Hawaiian palagonites are largely X-ray amorphous, but in detail typically exhibit a weak, clay-like absorption at 2.2 µm indicative of some ordering in an aluminous hydroxylated silicate (cf. Singer 1982, Orenberg and Handy 1992). Mineralogic analysis of the finest-grained fraction of Hawaiian palagonite by Golden et al. (1993) shows that it consists of ferric oxide and an aluminous phyllosilicate, such that the difference between the two proposed lithologies would involve only degree of crystallinity and small chemical variations. Although bright martian soil does not exhibit strong absorptions at 1.4 and 1.9 µm typical of clay minerals measured under ambient terrestrial conditions, it must be remembered that the martian surface environment is highly desiccating compared to terrestrial conditions. Laboratory studies have shown that desiccation of clays decreases the strengths of both of these absorptions (Rough 1987, Bishop et al. 1993a); absorption strength could be further weakened by coordination of interlayer water with Fe³⁺ or by dilution with opaque phases (Bishop et al. 1993a). Thus, while the ISM data do suggest the presence of phyllosilicate in bright martian soil, they also suggest that it is desiccated, poorly crystalline as in "palagonite," and/or partially masked by additional phases present in the soil.

CONCLUSIONS

Bright reddish materials on Mars are agreed to represent a surface layer of chemically altered soil. The lithology of this soil has previously been described as dust that has been largely homogenized by wind, that is unrelated to underlying geologic units, and that contains hematite and water but lacks strong evidence for crystalline silicate. High spatial resolution data from the near-infrared imaging spectrometer ISM on the Phobos 2 spacecraft
have shown that this description is overly simplistic and partially inaccurate. In this study we have demonstrated significant lithologic heterogeneity in bright soil, and we have provided evidence for its mineral makeup. A weak, narrow 2.2-μm absorption suggests the occurrence of poorly crystalline phyllosilicates comparable to those identified in altered pockets of the SNC meteorites and in Hawaiian palagonites. Spatial variations in strength of the 3.0-μm H₂O absorption suggest differences in the content of water in soils developed on specific geologic units. The position and shape of the 0.9-μm Fe³⁺ absorption indicate hematite to be present in bright soils covering some areas such as Tharsis, but spatial heterogeneities in this absorption also suggest that bright soils covering other areas including Arabia contain ferric minerals other than hematite. The bright soils do not constitute a single lithology, and their heterogeneities cannot be simply explained by mixing with “basaltic” material covering dark regions. Instead, the heterogeneities probably result from different histories of chemical alteration of parent materials and their resulting soils, possibly including local effects of liquid water. Additionally, eolian mixing has not been sufficient to erase differences in soil composition arising from local and regional geology.

At this stage in analysis of ISM data, it is not possible to draw unique mineralogic or lithologic interpretations of the spectral heterogeneities of bright soil. However, this situation may be improved upon in several ways. More detailed comparisons of ISM data covering specific regions with surface geology and visible-wavelength color will allow better assessment of competing hypotheses to explain the spectral heterogeneities. Laboratory studies of Mars soil analog materials with differing contents of salts and histories of environmental exposure will provide better “ground truth” for interpretation of spectral heterogeneities than does the current inventory of analog spectra, many of which are measured in pure form and under ambient terrestrial conditions. Finally, results from the Omega near-infrared imaging spectrometer and the SVET UV-visible spectrometer on Mars-94 will provide a more global assessment of the heterogeneous spectral properties of bright martian soil.

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