Martian Aerosols: Near-Infrared Spectral Properties and Effects on the Observation of the Surface

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INTRODUCTION

Understanding of the physical and chemical properties of aerosols in the martian atmosphere was greatly enhanced by space missions of the 1970s, namely Mariner 9 and Viking. Aerosols are known always to be present in the atmosphere, although the minimum opacity is uncertain. These particles consist mostly of suspended dust as well as H₂O and/or CO₂ ice crystals. Specific studies of the aerosols have concerned their composition, size distribution, vertical distribution, and seasonal behavior; most of them were made by using data from instruments that measured the visible wavelength range and sometimes the infrared (IRS/Mariner 7, IRIS/Mariner 9, IRTM/Viking). Several experiments showed that most optical scatterers are micrometer-sized particles. The first estimate of the particle size was proposed by Toon et al. (1977) from IRIS observations, and confirmed by Pollock et al. (1979) with the Viking Lander camera. These works inferred an effective (cross-section weighted) radius on the order of 2.4 μm (see Clancy and Lee 1991, and references therein). During the Phobos-2 mission, the solar-occultation instrument “Auguste” inferred an effective radius on the order of 1.2 μm (Korablev et al. 1993). The smaller size was ascribed to the absence of...
global dust storm in the previous few years. Optical scattering by particles of this size is expected to reach a maximum in the near-infrared, between 0.7 and 3 \( \mu \text{m} \). A recent treatment of photometric data from IRTM showed that between 0.3 and 3 \( \mu \text{m} \), even under conditions of low atmospheric opacity the magnitude of this contribution may reach 25\% of the surface albedos (Clancy and Lee 1991).

Optical scattering of this magnitude in the martian atmosphere is clearly a major problem for spectroscopic observations of the surface at infrared wavelengths. The aim of these observations is to identify the main mineralogical components of the materials by looking at spectral features having a typical depth of 1\%, and these are easily masked, subdued, or modified by the atmosphere. Nevertheless, this atmospheric contribution to the radiance of the planet has often been neglected in studies of the surface because of its small magnitude at 0.5 to 0.7 \( \mu \text{m} \) (e.g., Arvidson et al. 1989). In practice it is impossible to separate, in a rigorous manner, surface and aerosol contributions to existing infrared spectral data sets for Mars; this work would require multiple observations of the same regions under varying viewing geometry within a short lapse of time (minutes to hours).

In this paper we attempt a twofold investigation of martian aerosols. First, we derive estimates of these particles’ mineralogical composition, size distribution, and spatial and vertical distribution. Second, we identify the aerosols’ effects on infrared spectra of the surface and hence on mineralogic interpretations. The data used for this purpose were acquired by the ISM experiment on the Phobos-2 spacecraft in early 1989. They consist in sets of \( \approx 3000 \) contiguous spectra acquired within 20 min during a period of low opacity. The ISM data have been used previously to study both the martian atmosphere and the surface mineralogical composition of the equatorial regions. In the first section of the paper, we summarize the calibration and accuracy of ISM data. In the second section, we model the contributions from the surface and aerosols to the total radiance measured by ISM. In the third section we estimate the spectra of the aerosols and infer some of their physical and compositional properties. In the last section we discuss how the aerosols scattering affects the key spectral properties of the surface and to what degree it modifies compositional inferences drawn from these properties.

1. CALIBRATION AND ACCURACY

The instrument and the data set have been described in details in previous papers (Bibring et al. 1989, 1990, Erard et al. 1991) and are summarized here to introduce the discussion of the accuracy of the data.

1.1. Instrument Principle

The spectrometer spans the spectral range from 0.76 to 3.16 \( \mu \text{m} \) in 128 spectral channels (hereafter called spectrals). The measurements in all spectrals were performed simultaneously for a spot on the surface (hereafter called pixel) corresponding to the 12\% IFOV. Due to the altitude of the spacecraft, the pixel size is about 25 \( \times \) 25 km\(^2\) (medium-resolution) for most of the observations of Mars. The image format of the data resulted from both the orbital motion of the spacecraft and the scanning motion of an entrance mirror. The dispersing system was a grating, whose first and second orders were both used to minimize the size of the instrument. These two wavelength ranges (1.64 to 3.16 \( \mu \text{m} \) and 0.76 to 1.51 \( \mu \text{m} \), respectively) were separated by a beam splitter and focused onto two independent 64-element PbS arrays. Each array consisted of two rows of 32 detectors, looking in directions that differed by \( \approx 23\% \) (about two pixels width). Due to the difficulty in coregistration of measurements from the two rows, in this work we use for each order the row of 32 detectors whose signal-to-noise ratio proved to be higher (Erard et al. 1991). The resulting spectral steps between contiguous spectrals are 0.05 in the longer-wavelength first spectral order and 0.02 \( \mu \text{m} \) in the shorter-wavelength second order.

Although nominally separated by the beam splitter, some of the photons scattered in the first and third orders actually reached the detector array intended to measure photons from the second order. The main contributions are located below 0.85 \( \mu \text{m} \), from 1.0 to 1.3 \( \mu \text{m} \) (from the first order, proportional to the flux at wavelength 2\( \lambda \)) and from 1.15 to 1.45 \( \mu \text{m} \) (from the third order, proportional to the flux at wavelength 2\( \lambda \)). Conversely, only photons from the first order should have reached the long wavelength-range detector array, but a similar effect possibly occurred at its short-wavelength edge. As part of the calibration procedure, these order-overlap signals were estimated and removed.

1.2. Data Collected

The data used in this paper are from the nine medium-resolution image cubes acquired when the spacecraft was in three-axis stabilized mode. Their dimensions are approximately 25\( \times \)25 pixels, and they cover latitudinally oriented strips 500\( \times \)3000 km\(^2\) in size, located equatorward of 20\degree\ latitude. The integration time was 0.5 sec, and the analogic gain was 1, 2, or 3 depending on the image cube. All the observations were made at small phase angles (\( \phi \leq 10\degree \) in general) and close to the surface normal (at incidence and emergence angles smaller than 40\degree). Other observations were also acquired, but are not the primary focus of this paper. Those include higher resolution Mars
measurements from a lower-altitude orbit, two image cubes on Phobos, and strips covering the limb of Mars (see Bibring et al. 1990). Calibration sessions were also performed before and after each observation session to measure dark currents and verify instrument gain.

1.3. Data Calibration

The calibration procedure for ISM data has been described in detail in previous papers (Erard et al. 1991, Mustard et al. 1993a). In summary, the successive steps are:

- Subtraction of the average of dark currents measured before and after each observation session.
- Correction for misregistration of the spectrums. The misregistration between spectrums amounted to a few tenths of a pixel across the scanning direction. This problem is most significant near sharp albedo boundaries, where two spectrums look at areas having slightly different coverage by bright and dark materials. After correction, the relative brightnesses of the spectrums may be in error by up to 2% (see Erard et al. 1991).
- Division by the analogic gain chosen for the session.
- Correction for the variation of detector sensitivity with temperature.
- Removal of spectral order overlaps. The first order contribution of signal to the second-order detectors was measured on-ground using a blackbody, but the third order contribution could not be measured reliably with the available light sources. It was instead derived empirically from the in-flight data using spectral models for both Phobos and Mars and then refined by analyzing the statistical variance within the image cubes. A comparison with the semiempirical values inferred from the measurements of the optical parts of the instrument (before integration) proved to be satisfactory.
- Division by the solar spectrum scaled to the Sun–Mars distance at the time of observation.
- Absolute calibration, including division by the transfer coefficients and subtraction of an additional offset.

Ground-based calibrations were intended to provide the response function for each spectrel and the spectrels' relative sensitivity. However, this latter measure could not be performed confidently for the entire spectral range of the instrument; plus it was conducted at detector temperatures differing from those encountered in orbit (radiative cooling in space proved more effective than predicted). The absolute calibration of the data was instead derived empirically out to 2.6 μm, using overlapping ISM and telescopic (from McCord et al. 1982) observations of large, uniform dark and bright standard regions. This procedure is described in detail by Mustard et al. (1993a).

It provides a gain and an offset for each spectrel that are then applied to the whole data set. Above 2.6 μm, the spectra were calibrated by using the previous spectral model for Phobos (L6-Pervomaiski plus a thermal contribution, cf. Erard et al. 1991, Murchie et al. 1991) and scaled to match the telescopic model at 2.6 μm. The computed offsets represent less than 2% of the average signal from Mars in the long-wavelength range and up to 5% at the shorter wavelength edge.

In this procedure, the gains represent the transfer function of the instrument, while the offsets mainly account for a small amount of light coming from outside the nominal field of view. This latter component was estimated by Bibring et al. (1990) to be less than 2% of the signal at 2.0 μm, when looking at a uniform field. The offsets also depend on aerosol opacity during the telescopic observations that were used as reference spectra, as demonstrated by applying the model described in section 2.1 to the reference telescopic spectra. Altogether, the stray light acts like a wavelength-dependent, additive background, reducing the spectral contrast of the martian surface and producing systematic features in the uncorrected spectra.

This calibration procedure was developed to treat the observations of the martian surface, i.e., the nine medium-resolution and the two high-resolution image cubes that constitute the main body of the Mars observation sessions. In the cases of limb observations of Mars and the two image cubes of Phobos, the amount of light coming from outside the field of view is expected to be different. Consequently, the offsets are also different and the calibration procedure cannot be applied without modification.

1.4. Accuracy

Absolute reflectance. The radiometric accuracy of the absolute calibration is limited by that of the reference spectra. The radiance measured by a given spectrel is uncertain at the 15–20% level, due to uncertainties in the solar flux, changes with viewing geometry, and atmospheric variability of Mars that limits the comparisons between observations that were made at different times. These limitations are well illustrated by comparison of the 1978 reference spectra from McCord et al. with the most recent (1986) photometric observations of Mars, which are about 10% brighter in the infrared (see discussion in Roush et al. 1992). However, such a systematic error in the absolute values of the reference spectra would result only in a scalar difference from the albedo measured by ISM and would have no effect on compositional interpretations, provided that the spectral contrast between different regions of the surface remained similar.

Relative spectral accuracy. The spectrel-to-spectrel relative accuracy of the data depends most critically on
the absolute calibration, correct removal of order overlap, variation in the strengths of atmospheric gas absorptions between the times of acquisition of the reference spectra and the ISM data, and linearity of detector response. A critical test of the absolute calibration was performed for wavelengths less than 1 \( \mu m \), using telescopic observations acquired during the 1988 opposition. The comparison proved to be excellent and the calibration procedure adequate (Mustard and Bell 1994).

The first- and third-order contributions to the measured second order radiance induce wavelength-dependent additive errors. Uncertainties in their removal may thus result in systematic absorption-like features. Careful examination of spectral variations as a function of reflectance shows that, in general, these features cannot be separated from random noise and hence are of the same order. However, there are three wavelength ranges where these corrections may be significantly less accurate: the first two spectrals below 0.78 \( \mu m \), three spectrals between 1.07 and 1.14 \( \mu m \), and the first two spectrals of the first spectral order at 1.63 and 1.70 \( \mu m \).

The reference spectra used in the absolute calibration were compensated for atmospheric absorption features. Consequently, the calibration procedure decreases accuracy of the ISM data within the atmospheric bands, especially the wide \( CO_2 \) bands from 1.95 to 2.09 \( \mu m \) and from 2.58 to 2.96 \( \mu m \). Between 1.70 and 1.85 \( \mu m \) and in two spectrals around 1.40 \( \mu m \), the accuracy also depends on the precise content of water vapor in the reference spectra.

Finally, detector response was determined to be linear to within 1% for almost all observations. For a small number of pixels covering the very bright region around Ascreaus Mons, acquired at an analogic gain of 3, four spectrals in the wavelength range 0.9--1.1 \( \mu m \) show evidence for saturation. Fortunately, the saturation is observed only in every second spectrel, such that meaningful spectral information can still be obtained from the intervening channels.

All other possible errors in the data are smaller in magnitude than these four, so that in most of the spectral range of ISM spectrel-to-spectrel relative accuracy is comparable to or better than that of the most accurate telescopic spectra, typically 3-5%. Beyond 2.6 \( \mu m \), where no reference spectra were used, the actual relative accuracy may be lower.

**Accuracy of ratioed spectra.** The uncertainties discussed above limit the interpretation of an absorption feature in an individual spectrum, but not the meaningfulness of variations between spectra. In other words the limitations in accuracy do not limit the precision of the data. Within a single image cube, the precision is limited only by the instrumental noise and the accuracy of removal of additive components, discussed above. For these reasons variations within an image cube are significant statistically at much smaller magnitudes than the 3-5% limit.

The random noise was estimated directly from the flight data (observations of deep space and large uniform regions of Mars). At analogic gains 2 and 3, the sensitivity of the experiment was limited only by the electronic noise. The digitization noise was slightly more important for observations at gain 1.

For observations of regions having average martian spectral properties, at gain 2 the signal-to-noise ratio varies from 200 to 1500 at most wavelengths and becomes lower than 100 only beyond 2.7 \( \mu m \). These values slightly exceed those derived from the ground-based calibrations (Erard et al. 1991), due to the lower temperature of the detectors during observations. Since the fluctuations of the signal were estimated by comparisons of superficially similar spots which may in fact exhibit subtle spectral differences, these estimated signal-to-noise ratios are conservative.

The spectral variability of the martian surface is still small, however. Table 1 provides a comparison of the rms noise to the observed dynamic range of the main spectral attributes of the surface. The pixel-to-pixel variability is generally limited by instrumental noise and represents about 0.3% of the signal in most of the spectral range. Below 0.78 \( \mu m \), from 1.07 to 1.26 \( \mu m \), and from 1.63 to 1.70 \( \mu m \), variations of less than 0.5% may be meaningless. Above 2.8 \( \mu m \), a limit of 3% is imposed by the noise. These estimates are valid for individual spectra. This relative accuracy, or precision of the data, represents the real limit on the compositional interpretations made from ISM data.

**SPECTRAL VARIABILITY AND ITS ORIGIN**

2.1. Contributions to the Signal

The calibrated data are scaled as radiances factors, i.e., as the radiance relative that of a Lambertian surface observed under normal incidence. The measured flux is affected not only by the reflectance properties of the surface, but also by the properties of atmospheric gases and aerosols and possibly a thermal component. In the case of Mars, the temperature is low enough that the thermal component is negligible in the spectral range of ISM (lesser than 1% of the signal at 3.2 \( \mu m \)). At red and near-infrared wavelengths, gas molecules contribute only through absorption at given spectral locations. Outside these absorption features, scattering by micrometer-sized aerosols modifies the continuum of incident and emergent radiation.

**Modeling.** The observations from the Phobos-2 spacecraft took place around the northern spring equinox, with a solar longitude \( L_s \) ranging from 350° to 18°. Ac-
TABLE I
Range of Variation of the Spectral Criteria in the Syris–Isidis Image Cube

<table>
<thead>
<tr>
<th></th>
<th>Surface + scattering (measured)</th>
<th>Surface alone (Modelized)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Min</td>
<td>Max</td>
</tr>
<tr>
<td>Reflectance at 1.09 μm</td>
<td>0.115</td>
<td>0.33</td>
</tr>
<tr>
<td>Spectral slope (refl. units/μm)</td>
<td>−0.025</td>
<td>0.0002</td>
</tr>
<tr>
<td>3 μm hydration (band depth)</td>
<td>0.41</td>
<td>0.60</td>
</tr>
<tr>
<td>2.2 μm pyroxenes (band depth)</td>
<td>0.97</td>
<td>1.00</td>
</tr>
<tr>
<td>0.9 μm Fe²⁺ (band depth)</td>
<td>0.965</td>
<td>1.055</td>
</tr>
<tr>
<td>1 μm Fe²⁺ (band area)</td>
<td>0.0</td>
<td>0.40</td>
</tr>
</tbody>
</table>

Note. The rms noises on the criteria are computed from those on the individual spectrals.

According to long-term observations, this is usually a season of very low atmospheric opacity (Zurek and Martin 1993). Several instruments on-board Phobos-2 showed that only a minor amount of dust was present in the atmosphere by this time. A total vertical opacity of 0.3 was measured at 0.55 μm by the KRFM spectrophotometer (Moroz et al. 1991), consistent with ground-based observations made during the same period (e.g., Lecacheux et al. 1991). Opacity was evaluated as 0.25 at 1.9 μm from the solar-occultation measurements of the instrument “Auguste” (Korablev et al. 1993). Under conditions of such a low opacity, combined with the small incidence and emergence angles of ISM observations, the single-scattering approximation is valid: photons arriving at the instrument were scattered at most once in the atmosphere.

Under the single-scattering assumption, the measured radiance is the sum of three contributions: the light reflected on the surface and attenuated by the atmosphere, the light backscattered by the atmosphere before reaching the surface (atmospheric path radiance), and light scattered by the aerosols and reflected by the surface (indirect illumination). This formulation can be formalized as detailed in the Appendix. Since the phase angles are always small (φ ≤ 10°), the cosines of the incidence and emergence angles have similar values (μ0 = μ). Hence, the flux reaching ISM detectors can be approximated as

$$r_{tot} = r_s + \frac{\tau_{eff}(φ)}{4\mu} + \frac{\tau_f}{\mu} (aw - 2),$$  \(1\)

where the three terms are, respectively, the surface contribution, the atmospheric path radiance, and the difference between indirect illumination and extinction (only the latter term is affected by the approximation μ0 = μ). τ, f(φ), and w represent vertical opacity, the phase function and the single-scattering albedo of the aerosols, respectively. a describes the angular partitioning of light interacting with the surface and the aerosols (see Appendix). All quantities depend on wavelength, except the cosine of the emergence angle μ; a has a small dependence on the aerosols’ phase function and surface reflectance.

Aerosol contribution. Equation (1) implies that the relative contribution of light interacting with aerosols will depend on the reflectance of the underlying surface. This is consistent with multiple-scattering simulations developed to treat the photometric sequences from the IRTM radiometer (Clancy and Lee 1991). That study showed that, when observed under small phase angles through a thin aerosol layer, the dark terrains appear brightened, whereas the bright ones appear darkened. This implies that extinction is slightly larger in magnitude than indirect illumination (aw ≤ 2). The situation may be slightly different in the case of ISM observations: the measurements of IRTM are integrated over a broad band ranging from 0.3 to 3.0 μm and are most sensitive to visible wavelengths (λeff = 0.67 μm) with only a small contribution from IR wavelengths at which most of the scattering occurs. Atmospheric conditions were also different during Viking and Phobos-2 observations. In a previous study of the aerosols using ISM data Drossart et al. (1991) showed that, at least up to 2.6 μm, the reflectance of a bright region can be fitted by the sum of a surface component and a positive term proportional to 1/μ, independent of the surface reflectance. This is equivalent to taking into account only the backscattered contribution in Eq. (1), which means that extinction is nearly compensated by...
indirect illumination over most of the spectral range of ISM. The implicit approximation reads:

$$r_{\text{tot}} = r_\text{s} + \frac{\pi \mu f(\varphi)}{4\mu} .$$ (2)

In this context, the contribution of light scattered by aerosols is proportional to the aerosols' scattering cross section ($\sigma_w$). The factor $1/\mu$ accounts for the geometric path length between the surface and the spacecraft, along which are located the backscattering particles.

The last formulation is consistent with the fact that the ISM spectra almost always have a negative slope, as the scattering continuum is expected to exhibit. It is also consistent with numerous observations of Mars showing that the Minnaert exponent describing limb darkening increases with wavelength and normal albedo (also evidenced by ISM, as explained in the next section). Furthermore, this empirical model explains the almost constant difference between "Lambert albedos" measured by IRTM under clear atmospheric conditions (Pleskot and Miner 1981) and the Clancy and Lee estimate for the surface alone (see their Fig. 18. The difference is ≈0.05 whatever the surface albedo). The approximation of Drossart et al. described above is therefore not contradicted by NIR observations and is reasonably followed at least within areas with no major albedo contrast.

According to Eq. (2), the aerosols' additive contribution to the overall reflectance of Mars must bear some similarity with the spectra of the limb of the planet. Figure 1 shows two such spectra acquired at gain 3 at different altitudes over Xanthe Terra. Their accuracy is limited by three factors: a low signal; misregistration causing spectels to look at altitudes differing by several kilometers, at which the amount of backscattering is probably not uniform; and a larger amount of stray light than for observations acquired near the center of the disk, such that the offset correction described in Section 1.3 is inappropriate. In order to remove the stray light from these two spectra, we subtracted a third spectrum acquired several seconds later for a slightly higher altitude. The general shape of these spectra is a close approximation to the scattering continuum.

Where the wavelength of incident light is comparable to the size of the scattering particle, the envelope of the aerosol contribution is given by Mie's theory using the refraction index in the continuum. In the general case of scattering centers with a continuous particle size distribution, the scattered light component reaches a maximum at a wavelength governed by the average particle size and then smoothly decreases with increasing wavelength. Both spectra in Fig. 1 exhibit this negative spectral slope at longer wavelengths, with a maximum in the darkest spectrum near 0.9 μm. This infrared spectral slope, together with the brown-red color of the martian sky, imply that maximum scattering occurs at 0.6–1.0 μm. Apart from this continuum, the spectrum of scattered light should also exhibit the absorption bands characteristic of the material composing the aerosol particles. Although the uncertainty in the limb spectra of Fig. 1 is very large, the 2.9-μm features in the darkest one (acquired at higher altitude) may be indicative of water-ice.

**Atmospheric absorptions.** The major atmospheric absorptions are due to CO$_2$, H$_2$O, and CO and are easily identified (Fig. 8a). To a first approximation they can be modeled by multiplying Eq. (1) by an exponential function of wavelength-dependent absorptivity and the atmospheric path length (which depends on both the elevation and the viewing geometry). This approximation makes it possible to remove most of the atmospheric gas absorption features, using an absorption spectrum derived from the data as described by Erard et al. (1991). However, this procedure is inadequate for absorptions approaching saturation or where atmospheric scattering is large. More accurate altimetry measurements were made using an absorption lines library (Bibring et al. 1991). However that method did not allow a satisfying fit of the 2.7-μm CO$_2$ band. This indicates that scattering is not negligible at this high wavelength.

**Surface reflectance.** The radiance factor of the surface $r_s$ exhibits wavelength dependence governed in part by absorption features caused by minerals on the surface. However, this term also depends on the physical state of the surface, such as particle size or porosity (e.g., Mustard and Pieters 1989), and on viewing geometry (e.g., Gradie et al. 1980). Thus the magnitude and the shape of the continuum, and to a lesser extent the depths of absorption bands due to surface materials, are affected by these intrinsic and observational quantities. In order to compare properties of surfaces measured at different geometries,
it is therefore necessary to correct reflectance measurements for limb darkening induced by variation in the viewing geometry.

2.2. Continuum Variations

Among the most prominent spectral features in the ISM data are variations in reflectance and spectral slope, as described by Erard et al. (1991). The spatial variations in continuum reflectance can reach a factor of 3 and depend on surface albedo, on the relative contribution of the aerosols, and on the viewing geometry. From Eq. (2), the effect of greater aerosol opacity will be to increase the reflectance at each wavelength by a value independent of surface reflectance and thus to reduce the spectral contrast between various regions. According to photometric models (e.g., McEwen 1991) and lunar observations (e.g., Helfenstein and Veverka 1987), changing viewing geometry can result in variations in surface brightness of up to 25%, with the maximum reflectance at the subsolar point.

Since the expected aerosol spectrum decreases in reflectance with increasing wavelength, the aerosol contribution will produce more steeply negative spectral slopes, i.e., bluer spectra, at higher opacities and longer lines of sight (1/μ). Changing viewing geometry is also known to result in changing the spectral slope of mineral samples, but the systematics are opposite in the near-infrared to those resulting from aerosol effects: mineral spectra generally become redder with increasing emergence or phase angles (e.g., Gradie et al. 1980). The same effect is observed on atmosphereless bodies at the scale of tens of kilometers (e.g., Jaumann et al. 1992).

Photometric behavior. To correct the data for limb darkening, we decided to use the phenomenological approach of Minnaert rather than a more sophisticated model based on the properties of surface materials, which would employ more free parameters and would not be easily applicable to observations through an atmosphere. Minnaert's law describes the geometrical variations of reflectance as

$$ r = B \frac{\mu^k}{\mu - k}, $$

where $B$ is equivalent to normal albedo, and $k$ describes limb darkening. Both parameters, $B$ and $k$, are functions of phase angle and wavelength (e.g., Binder and Jones 1972).

Two major problems with Minnaert's functions have been identified: a singularity either at the limb or at the terminator and multiple solutions at phase angles in excess of about 30°. Although the meaning of the two parameters is not straightforward, Minnaert's law is well-adapted to the reduction of ISM data, because the variations in incidence and emergence angles are much larger than the variation in phase angle. Furthermore, the ISM observations on which this paper is focused are not located near the edges of the disk and the phase angle is always small ($\leq 15°$), so that the problems described above are minimized.

We first fitted Minnaert exponents using profiles across several regions near the photometric equator. Although care was taken to select profiles within areas that did not exhibit abrupt albedo changes, possible small departures of the surface from spectral homogeneity represent the greatest source of uncertainty in the exponents. For example, one image cube covers bright regions encompassing the Medusae Fossae formation and bright plains of the Tharsis plateau. The signal decreases smoothly from west to east as the photometric angles increase; however, the rate of this variation is different in the eastern and western halves of the image cube, even though no abrupt albedo boundary is visible. The difference between the exponents fitted using these two image-cube halves ($\pm 0.04$) is assumed to be the uncertainty of the estimation.

All the observed regions can be fit using a Minnaert's law, although the fit is in general not as good in the dark regions, which is not surprising considering their stronger intrinsic variability. Consistent with the systematics of previous measurements at comparable wavelengths (Binder and Jones 1972, Pleskot and Kieffer 1977) the Minnaert exponents fitted using ISM data slightly increase with phase angle and wavelength and generally also with albedo.

The variation with phase angle is shown on Fig. 2, as measured at 0.91 μm in the most homogeneous regions, together with values from Binder and Jones (1972). At low phase angles, ISM and telescopic exponents are of
the same order, but at larger phase angles only in Lunae Planum does the ISM exponent follow the trend from telescopic spectra. In contrast, it exhibits no phase dependence in southern Amazonis and east of Ascreaus Mons. This difference in the phase dependence of \( k \) in different regions could result from a difference in spatial resolution, from differing surface textures, or from different aerosol opacities. Figure 3 depicts the spectral variation of \( k \) for ISM data covering southern Amazonis Planitia (dashed line) and for the telescopic observations of Binder and Jones at similar phase angles. Both sets of exponents exhibit the same increase with wavelength, consistent with a decreasing additive contribution from the aerosols. We also derived an additional estimate of the Minnaert exponents by using the overlap between image cubes over Tharsis, as described in Section 3.1. Three hundred fifty-four pixels are observed here under two different incidences and emergences, but with a similar phase angle. The Minnaert exponents are computed here in every spectral, and the values are averaged over two groups of \( \sim 175 \) pixels corresponding to phase angles of about 6° and 8°. The error bar represents statistical dispersion. Figure 3 shows the results outside the atmospheric bands, up to 2.6 \( \mu m \). The exponents computed this way are significantly larger than those derived using the first method, but have a similar wavelength dependence.

Most of the values of \( k \) from ISM data range from 0.62 to 0.82, an interval which is only five times larger than the estimated uncertainty. An average value of \( k = 0.70 \) is therefore used to correct albedos on every terrain, at every phase angle, and at every wavelength. Recent measurements at comparable wavelengths yield consistent values, which range from 0.64 to 0.77 (e.g., de Grenier et al. 1991, Pierazzo and Singer 1993).

**Estimates of albedo and spectral slope.** Albedo can therefore be retrieved from the ISM reflectance measurements using Eq. (3), with \( k = 0.7 \). This exponent is found to yield good compensation for limb darkening and to improve the match in albedo between adjacent image cubes. The result is an estimate of the normal albedo of the surface plus the atmosphere. The maximum local contrast is observed in the Syrtis Major–Isidis Planitia image cube, where the albedo at 1.09 \( \mu m \) varies between 0.12 and 0.32.

The spectral slope has commonly been estimated as a ratio of reflectances at different wavelengths (e.g., Erard et al. 1991, Bell 1992). This parameter is widely used because it does not require an accurate calibration of observational data. However, it results from Eq. (2) that this estimate is severely biased in the case of Mars: scattering under conditions of uniform opacity results in a uniform, additive contribution with a given spectral slope. Neglecting the variations in spectral slope of surface materials and assuming that they vary only in brightness, the relative contribution of the aerosols is in this case greater over dark regions. Then the reflectance ratio described above depends systematically on the surface reflectance, and a spurious correlation between low albedos and steep negative slopes is the result. It is therefore necessary instead to estimate the spectral slope as the derivative of reflectance between two spectrals. We use the spectrals at 1.84 and 2.35 \( \mu m \), which are outside strong martian absorptions features and are contained in a single spectral order. This estimate is not correlated closely with albedo (see Mustard et al. 1993a), but as a result of the backscattered contribution from aerosols, there remains a correlation between steep negative slopes and large emergence angles. Important variations in spectral slope unrelated to viewing geometry are recognized over different surface features and imply that different surface materials may have different spectral slopes (Erard et al. 1991, Murchie et al. 1992, Mustard et al. 1993a, Fischer and Pieters 1993). As this estimate of spectral slope is given in units of reflectance, the Minnaert correction should be applied to it.

**AEROSOLS CONTRIBUTION**

A first estimate of the aerosol spectrum was given by Drossart et al. (1991), based on data that were incompletely calibrated. Using the model given by Eq. (2) and assuming for the surface a Minnaert photometric function with an exponent comprised between 0.7 and 0.9, their study of the photometric variations within a single rela-
tively homogeneous image cube west of Pavonis Mons yielded a range of estimates of the backscattered light. A comparison with telescopic observations of September 1988 indicates that the actual contribution is most likely within the lower half of their stated range (Erard 1991, Erard et al. 1992). Another method of estimating the aerosol spectrum is used here, which exploits the spatial overlap between pairs of image cubes measured at different geometries.

3.1. Derivation

ISM image-cubes overlap only in two locations, in Tharsis around Pavonis Mons and at the boundary of Melas Chasma and Ophir Planum. Both image cubes over Tharsis were acquired at gain 2 (see Bibring et al. 1990 for references). More than 600 overlapping pixels were coregistered for study. The easternmost ones were rejected to avoid observations under very high emergence angles; the pixels covering the volcano were also rejected to minimize possible effects of orographic clouds, and also because small registration errors there could result in major differences in average altitudes sampled. We eventually used 354 pairs of pixels located east of Pavonis Mons (from 102°W to 112°W and from latitudes −5° to 0°) and acquired at low phase angles (4° to 10.5°). Within each pair one spectrum was acquired at a high emergence angle (36°–49°) and the other at a more nearly normal geometry (26°–36° emergence angle). A difference in atmospheric opacity between the two dates has been suspected from inconsistent estimates of altimetric measurements in the two image cubes (Combes et al. 1991, Bibring et al. 1991). It is therefore possible that a thin ice cloud was present during acquisition of the higher emergence angle observations, in which case the aerosol spectrum derived from this location may be partly indicative of cloud properties.

The other two overlapping image cubes were acquired with different analog gains. The region of overlap includes northern Melas Chasma and parts of Ophir Planum on the plateau plains. We selected 83 pixels in the more uniform area on the plateau, located at 59°–67°W, 11.0°–8.5°S. The phase angles range from 1.4° to 5°; the emergence angles range from 34° to 42° in one image cube and from 22.5° to 26.5° in the other.

Outside atmospheric absorption bands, the relationship between two spectra covering essentially the same pixel can be rewritten from Eq. (2) as

\[
r_{\text{low}} = r_s(\mu_{\text{low}}, \mu_{\text{low}, \lambda}) + \frac{\tau_{\text{wf}}(\varphi)}{4\mu_{\text{low}}} \]

\[
r_{\text{high}} = r_s(\mu_{\text{high}}, \mu_{\text{high}, \lambda}) + \frac{\tau_{\text{wf}}(\varphi)}{4\mu_{\text{high}}},
\]

where the subscripts low and high refer to the cosines of the emergence angles \(\mu\); \(r_s\) is the reflectance of the surface alone, including its photometric function; and \(\tau_{\text{wf}}(\varphi)/4\) is the vertical contribution of the aerosols in reflectance units.

Assuming that the ratio \(K = r_s(\mu_{\text{high}})/r_s(\mu_{\text{low}})\) (variation of the surface photometric function) can be measured at some wavelength and remains constant over the spectral range considered, an estimate of the vertical aerosols contribution can be derived from each pair of spectra:

\[
\frac{\tau_{\text{wf}}(\varphi)}{4} = (K r_{\text{low}} - r_{\text{high}}) \left(\frac{K}{\mu_{\text{low}}} - \frac{1}{\mu_{\text{high}}}\right).
\] (4)

This procedure was applied to all the selected pairs of spectra. Our final estimate of the aerosol spectrum over each overlapped region is the average of the spectra computed for each pixel. The results are shown in Figs. 4a and 4b for spectrums outside the main atmospheric absorption bands. The statistical error (standard deviation of the estimated mean aerosol reflectance at a given wavelength) includes instrumental noise, the spatial variability of the aerosols, and possible registration errors. It should be stressed that this method is independent of that used in Drossart et al. (1991), since no hypothesis is made concerning the shape of the surface photometric function. However, the validity of these evaluations is dependent on five assumptions:

• We used the simplified Eq. (2) which assumes that the difference between extinction and indirect illumination is small compared with backscattering.
• Vertical aerosol opacity is assumed to be identical within the two overlapping image cubes.
• The variation of the aerosol phase function \(f(\varphi)\) is neglected.
• The difference between two spectra of the same pixel is attributed entirely to the aerosol contribution, i.e., we assumed that the photometric function of the surface is independent of the wavelength.
• To compute the normalizing factor \(K\) in Eq. (4), we have to assume the magnitude of backscattering \((\tau_{\text{wf}}(\varphi)/4)\) at some wavelength. The spectra in Figs. 4a and 4b are computed assuming no scattering at 2.6 \(\mu\mbox{m}\).

A defect in one of the first three hypotheses would simply affect the magnitude of aerosol reflectance and would increase the statistical dispersion. On the contrary, the shape of the estimated aerosol spectra may be affected by the last two assumptions.

We estimated the effect of the surface photometric function by using spectra of basalt measured by Gradie and Veerka (1982) at a low phase angle (4°) and varying
emergence. When applied to these spectra, the procedure described by Eq. (4) yields, up to 2.7 μm, a spectral slope five times smaller in magnitude than observed in ISM data and opposite in sign. However, photometric effects also result from different geometries of subpixel scale patches of materials having different spectral attributes. The area used on Tharsis is extraordinarily uniform and exhibits only minute macro-scale variations in spectral properties, so that the subpixel variability is probably also small. In these conditions, the photometric effects would be negligible. However, the lesser spatial uniformity of spectral properties on Ophir Planum may contribute to the larger uncertainty in Fig. 4b.

We estimate the magnitude of backscattered light at 2.6 μm, where its relative contribution is expected to be near minimum. The study of Drossart et al. indicates a vertical contribution of backscattered light between 0.00 and 0.01 at this wavelength; these values are confirmed by studies of the effects of scattering on the 2.75-μm CO₂ band. Figure 4c compares aerosol spectral reflectance computed assuming that its value at 2.6 μm is 0.00 or 0.01. Neglecting the contribution at 2.6 μm results in a small systematic error in the estimates, equal to the surface spectrum, scaled to the magnitude of scattering at this wavelength. Outside the atmospheric bands, this error is almost independent of wavelength up to 2.8 μm.

In summary, the assumptions we had to make to derive estimates of spectrum of scattered light are reasonable. At wavelengths shorter than ~2.6 μm, they mainly result in increased statistical dispersion (our estimate of the error bar) and in uncertainty in magnitude of scattered light by a factor of 2 or less. The shapes of the spectra (spectral slope and possible absorption features) should be unaffected. Conversely, beyond ~2.6 μm, the shape greatly depends on the level of scattering assumed at this wavelength and on the photometric function of the surface materials. On this basis the estimated spectrum of aerosols derived over Tharsis, a more uniform area, should be more reliable than the estimate over Ophir Planum.

3.2. Analysis of the Aerosol Spectrum

The overall shape of the aerosol spectra derived above is similar to that of the limb spectra, with a negative spectral slope beyond 1 μm. Differences may be related to the fact that low altitude dust does not contribute to the limb spectra. The estimate derived over Tharsis is also consistent with the lower half of the range derived by Drossart et al.: a very steep slope at the beginning of the spectral range, with a ratio of about 4 between the magnitudes of scattered light at 1.0 and 1.5 μm; a decreasing slope with increasing wavelength; and no maximum at the short wavelength end of the spectrum. Although noisier, the spectrum derived over Ophir Planum shows a flattening from 0.95 to 1.4 μm and some evidence of a positive spectral slope at shorter wavelengths.
The shapes of the spectra of scattered light contain information on the physical properties of scatterers. We fitted local reflectance maxima within the two spectra with a cubic spline (Fig. 4). The resulting envelopes are representative of the scattering continua from 1.0 to 2.4 μm, and are related to the size distribution of the scattering particles. The shapes of the two envelopes are consistent with the size distribution derived from Phobos-2 Auguste measurements (effective radius ≈1.2 μm and effective variance ≈0.2; see Chassefière et al. 1992, Korablyev et al. 1993, Drossart et al. 1991). The fit is actually better if a backscattering magnitude of 0.01 is assumed at 2.6 μm in the ISM spectra (solid line in Fig. 4c). Over Ophir Planum, the occurrence of a scattering maximum within the wavelength range of ISM and the lesser spectral slope may indicate larger average and minimum sizes than over Tharsis. Alternatively, the maximum can result from deeper Fe⁺⁺⁺ absorption below 0.95 μm. Adopting optical parameters derived from IRTM (asymmetry parameter of the phase function g = 0.6, and w = 0.92) leads to opacities of 0.23 (Tharsis) and 0.62 (Ophir) at 1.9 μm, compared with values of 0.16 to 0.35 at 1.9 μm derived using data from Auguste (Korablyev et al. 1993).

The ratios of the estimated aerosol spectra to their continua derived above are shown in Fig. 5. This representation highlights absorption features due to the airborne particles. In the Tharsis spectrum, all features more than two spectrals in width are statistically significant. Most are similar in strength and position with those observed in the water-ice spectrum plotted in the same figure: weak absorptions from 1.09 through 1.39 μm, a much stronger one below 1.69 μm, the strong absorption from 1.83 to 2.20 μm, and finally the drop-off beyond 2.20 μm. The drop-off of the continuum itself makes it impossible to obtain a correct fit of the absorption above 2.3 μm. The spectrum from Ophir Planum also exhibits ice features, in particular the absorptions from 1.4 to 1.7 μm and above 1.83 μm. The absence of an absorption beyond 2.3 μm is meaningless because of the problem with continuum drop-off.

These absorptions remain even if the spectra are recalculated assuming that the magnitude of backscattering at 2.6 μm is 0.01 instead of 0.0 (solid line spectrum in Fig. 4c). In this case, an additional absorption above 2.92 μm is also apparent and is consistent with the strong water-ice band at that wavelength. However, this feature cannot be interpreted in detail, because a change of the magnitude of backscattering assumed at 2.6 μm results in a variation in the aerosol spectrum proportional to the surface spectrum (dash-dotted line in Fig. 4c), which itself contains the very similar band due to hydrated minerals.

The absorptions apparent in both spectra show departures from those expected for water-ice. The Ophir spectrum is too noisy at short wavelengths to infer anything conclusive from the apparent absorption from 1.0 to 1.2 μm, but it must be noted that its general shape (Fig. 4b) could be reconciled with a scattering maximum located at a shorter wavelength as in Tharsis, with a superimposed absorption below ≈0.9 μm. Such an absorption would be indicative of Fe⁺⁺⁺ in ferric oxides. However, this absorption is quite weak even in pure bright soils, so that failure to identify it conclusively in aerosol spectra having the uncertainties shown in Fig. 4 is not surprising.

A significant inflection also occurs in the Tharsis spectrum between 1.74 to 1.83 μm (Fig. 5c). This feature is absent from water-ice spectra, but is found in the spectra of several common hydrated minerals. Clark (1981) attributed this feature to a combination overtone of water and suggested that it is indicative of water in clay minerals (e.g., montmorillonite or kaolinite). To evaluate its possible origins in the aerosol spectrum, we undertook a systematic search through laboratory spectra of minerals that have been proposed as major components of the Martian soils and dust, on the basis of the Viking Landers analyses and telescopic or orbital spectral observations. These include ferric oxides (mainly hematite or goethite), silicates (either plagioclases, montmorillonite, or micas), sulfates, and carbonates (see Soderblom 1992, Bell 1992, Pollack et al. 1990). The best fits of the 1.8 μm feature are provided by several clay minerals (montmorillonite, nontronite, saponite) and one sulfate (gypsum). No convincing similarity is found with spectra of carbonates, micas, plagioclases or iron oxides. The spectra of clays and gypsum also contain other, deeper, absorptions in the spectral range of ISM, but most of them are related to OH or H₂O vibrations and thus are nearly indistinguishable from water-ice bands. A notable exception is the metal–OH absorption occurring at 2.2 μm, which occurs in clay spectra but is absent from the water-ice spectrum. Such an absorption could actually be present in the aerosol spectra, if the continua at this wavelength have been underestimated. However, no single nonice analog exhibits all the ice-like absorptions in aerosol spectra. Figure 5c demonstrates that the 1.8 μm feature can be understood as a combination of water-ice and hydrated clay absorptions.

In summary, aerosol spectra exhibit relatively deep absorption bands that can be explained by water-ice. The Tharsis spectrum also exhibits a very small feature at 1.8 μm that, if real, is consistent with clay or possibly gypsum in atmospheric dust. Additional absorptions, at 0.8–0.9 μm due to ferric oxide and 2.2 μm due to clay, are not clearly demonstrated but may exist.

3.3. Variations in Opacity with Altitude

In general it is impossible to measure the contribution of aerosols to a single spectrum, and hence variations of opacity within an image cube cannot accurately be de-
FIG. 5. Estimates of the aerosols spectra divided by their computed continua, (a) from Tharsis and (b) from Ophir Planum. The error bars show statistical dispersion between spectra derived from individual pixels and do not take into account uncertainty in the continuum. A spectrum of water-ice (W. Calvin, personal communication 1993) is plotted for comparison (dashed line). It was computed to restore laboratory measurements to a viewing geometry of \( \theta = 0^\circ \) and \( \epsilon = 30^\circ \) and particle size of 10 \( \mu \)m, judged as appropriate for ISM's viewing geometry of martian ice haze. The result was then convolved with the ISM response function and finally divided by a fitted continuum. (c) Details of the 1.8-\( \mu \)m region of (a), compared with the ice spectrum and a laboratory sample of kaolinite mixed with 25 wt% water (from Clark 1981, normalized at 1.68 \( \mu \)m).

derived. However, it still may be possible to derive information on variations in aerosol density with altitude, by examining the spectral slopes observed on a uniform material present at a variety of elevations.

On the flanks of the Tharsis volcanoes, there are large elevation variations within a small area. If the surface material is uniform, the spectral slope would only be a function of aerosol opacity and may depend on surface elevation. In their analysis of spectral slope variations on Olympus Mons, Fischer and Pieters (1993) defined three concentric annuli on the flanks of the volcano. Steeper spectral slopes were found to be correlated with higher albedos, with deeper Fe\(^{2+}\) features, and to some degree with morphology, but not with elevation. In this case spectral slope is thus controlled by variations in surface properties.

Pavonis Mons was observed from both the west and east by ISM at medium resolution; another western view was acquired at higher resolution (5-km pixel size). Although significant variations in spectral slope occur on the flanks of the volcano, they do not bear a clear relationship with surface altimetry for the two medium-resolution observations (Fig. 6). Instead, the base of the eastern flank displays a steeper spectral slope in both image cubes; the direction of solar illumination is opposite in the two cases. This area of steeper spectral slope more or less follows the limits of the youngest lava flows. A similar lack of altimetric variation was reported for the high-resolution image cube by Drossart et al., but a correlation between altimetry and spectral slope was observed within the caldera.

Only on Ascreaus Mons does the spectral slope appear to decrease with increasing elevation. Thus, apart from this last case, over the flanks of the Tharsis volcanoes the column density of aerosols seems to be independent of that of CO\(_2\). This result is consistent with an observation of Olympus Mons by IRTM, that showed an increasing dust-to-CO\(_2\) mixing ratio above 10 km (Clancy and Lee 1991). One of the proposed interpretations was that the volcano flanks provide an additional source of dust to the atmosphere.

A slightly different approach to this problem is to examine homogeneous materials present over a broader area with a regional topographic slope providing a large elevation range. The western flank of the Tharsis plateau is such an area, and it is covered by the image cube ex-
tending west from Pavonis Mons. The emergence angle is larger toward the east, where terrains of higher elevation are observed. The variation in elevation is actually so important that it controls the airmass between the surface and the instrument, which decreases toward the east. On homogeneous regions like this one, most of the spectral slope variation should be due to the backscattered light component. According to Eq. (2) this is proportional to $-\tau/\mu$. Over the whole window, spectral slope is observed to be negatively correlated with the secante $1/\mu$, but positively correlated with the airmass as measured from the depth of the 2 $\mu$m CO$_2$ absorption (Fig. 7). These observations are consistent with Eq. (2), provided that the vertical opacity of the aerosols is independent of elevation and remains nearly constant across the region. This situation could be explained in either of two manners: most scattering is due to a high-altitude layer of particles, as it would be natural for a thin haze of ice crystals; or alternatively, most scattering is caused by a spatially uniform layer of uplifted dust, likely to be concentrated in the first few meters above the surface. This vertical structure is consistent with that inferred by Jaquin et al. (1986) from Viking limb observations.

In conclusion, on the volcano flanks and on the western flank of the Tharsis plateau, aerosols opacity is largely independent of elevation, implying an increase of the aerosol-to-CO$_2$ mixing ratio with altitude. It is interesting to compare this result with those obtained for the minor atmospheric constituents. Combes et al. (1991) and Rosenqvist et al. (1992) found evidence for a decrease in the CO-to-CO$_2$ mixing ratio over the volcanoes based on strength of the 2.35 $\mu$m CO absorption in the ISM measurements. These observations could be related if recombination of CO into CO$_2$ is catalyzed by airborne dust or ice crystals, as proposed by Atreya and Blamont (1990).

4. INFLUENCE OF SCATTERING ON THE OBSERVATIONS OF THE SURFACE

In this section, we try to eliminate the aerosol contribution to ISM spectral measurements in order to obtain spectra of the surface alone. The Syrtis–Isidis image cube is used for this purpose because it exhibits the largest spectral contrast. The six spectra in Fig. 8a are representative of the diversity observed in the whole area. To correct them for the aerosol contribution, an estimate of the average opacity must first be derived.

4.1. Derivation of Spectra of Surface Materials

Based on the above analysis of Tharsis we assume that the variations are small, and we estimate an average amount of scattering throughout the whole image cube using the spectra derived in Section 3.1. This is accomplished through comparison of the spectral slopes at shorter (second order) and longer (first order) wavelengths. If the continuum spectral slope of surface materials is exponential, then the slopes at first- and second-order wavelengths should be proportional regardless of the surface albedo. Departure from this correlation is then
due to the additive component of the aerosols' backscattering. Conversely, once an accurate estimate of the aerosols' spectrum is removed from the data, correlation between the two slopes should be maximized.

Using the estimate from Tharsis, the highest correlation is found for an aerosol opacity 0.6–0.9× that present over Tharsis. A value of 0.75 also results in a limb darkening independent of wavelength (in other words, k = 0.72 at $\phi = 5^\circ$). Adopting the Clancy and Lee optical parameters for airborne dust ($w \gtrsim 0.85$ and $g = 0.6$), this corresponds to an opacity of $\tau = 0.17$ at 1.9 $\mu$m. The main effect of neglecting the magnitude of scattering at 2.6 $\mu$m is to overestimate surface reflectance by a few percent.

Our working hypothesis for this section is that, in this particular image cube, scattering increases reflectance uniformly by 0.75/µ times the continuum shown in Fig. 4a. Consequently, this quantity was subtracted from the representative spectra (Fig. 8b). The implicit assumption is that vertical opacity is independent of albedo. Of course this correction is not completely accurate, but we believe that the spectra in Fig. 8b are a much closer approximation to spectra of the surface alone than are the uncorrected data.

The darkest spectra are probably overcorrected, however, since uplifted dust is likely to remain preferentially over bright dusty areas. It must be clear that in dark areas, this procedure is not equivalent to “unmixing” a fraction of bright material: the continuum we subtract results from scattering by separated particles and is not expected from bright surface materials, even if the latter have the same composition as airborne dust.

4.2. Effects of Scattering on Surface Spectral Features

Comparison of spectra that are corrected and uncorrected for the aerosols' contribution of backscattered light makes it possible to study the influence of scattering on interpretations of surface composition made from ISM data. For this purpose, the spectral properties of surface materials are parameterized here by albedo, spectral slope, and five spectral parameters related to the main absorption bands of surface origin. These parameters have been selected from among various methods to measure the absorptions so as to maximize the signal-to-noise ratio and also because they were found to offer minimum sensitivity to subtraction of a scattered light spectrum. The occurrence of these spectral features in ISM data and their interrelationships have been described in previous papers (Bibring et al. 1990, Erard et al. 1991, Mustard et al. 1993, Murchie et al. 1993). Table I gives the range of parameterized spectral properties of the Syrtis–Isidis image cube, computed both before and after removing light backscattered by aerosols.

Reflectance and Photometric Behavior. By hypothesis, the spectra corrected for aerosol scattering are unchanged in shape beyond 2.6 $\mu$m, but reflectance is lowered and spectral contrast increased at shorter wavelengths. Reflectance measured at 1.09 $\mu$m without photometric correction ranges from 0.115 to 0.33; after subtraction of the aerosol spectrum, it ranges from 0.08 to 0.30. Thus albedo contrasts are increased from a factor of 3 to a factor of almost 4. The calculated reflectance of the aerosol spectra in Fig. 4, 0.02 to 0.05 below 2 $\mu$m, represents 5–15% of the albedo (Minnaert-corrected reflectance) of bright areas and 13–30% of that of dark regions. This estimate of the contribution of scattered light to total reflectance of Mars is similar to that inferred for “clear atmosphere” from IRTM observations (Clancy and Lee 1991). Although this value may seem large, it is still much smaller than differences in reflectance between
surface units, so that variations in albedo are dominated by surface materials.

Under our assumptions that photometric behavior of the surface is independent of wavelength, and that scattering is very small or negligible at 2.6 μm, the photometric properties of the surface alone (free of atmospheric effects) can be determined using measured reflectance at that wavelength. This yields a Minnaert exponent probably between 0.65 and 0.75 at phase angles of 5°–10° (Fig. 3). This exponent is much larger than that of the Moon (0.5–0.55 at similar phase angles) and indicates a less rough and porous surface, consistent with the presence of an atmosphere. These values of the Minnaert exponent would be underestimated if a notable amount of scattering has been neglected at 2.6 μm. However, the Martian surface, and especially the dark regions, appears significantly different from a Lambertian reflector (which Minnaert exponent is 1).

Spectral slope. The spectral slope at 1.84 to 2.35 μm, computed after subtraction of the aerosol spectrum, ranges from −0.01 to 0.02 so that not all the materials have a negative spectral slope in this wavelength range (Figs. 8a and 8b). The main effect of the correction is to reduce the slope of spectra acquired under high emergence angles. Contrast between dark regions such as the eastern and western parts of Syrtis Major is enhanced, and the values in west Syrtis are now very similar to those observed in Isidis.

The steep negative slope in eastern Syrtis has previously been interpreted as the effect of a thin ferric coating on basalt (Mustard et al. 1993a), which is a reasonable configuration on Mars and is known to modify the continuum of laboratory samples (Singer and Roush 1983, Fischer and Pieters 1993). Other possible origins of spectral slope variations include variations in particle size and induration of the soils, and may be responsible for a good part of the differences between slopes of spectra in Fig. 8b. Table II summarizes the influences of likely surface processes known to affect spectral slope. This comparison demonstrates that small spatial variations in the backscattering of light by aerosols, expected under clear atmospheric conditions, could result in variations of measured spectral slope as large as those induced by differences in surface properties. Hence, near-infrared spectral slope should only be used to characterize surface materials where variations correlate very strongly with morphologic features and/or spatial variations in mineralogic absorptions. In particular, spectral slope should not be incorporated into the set of parameters used to define surface units by means of automatic classification of spectral measurements.

Band strengths. From the model used to remove the aerosol contribution, below 2.6 μm mineralogic absorp-

| TABLE II | Influence of Some Processes on the NIR Spectral Slope |
|-----------------|-----------------|-----------------|-----------------|
|                | Δε(1.2 μm) = 0.2° | Δε(2.5 μm) = 0.06 |
| Aerosols backscattering | 0.024 | 0.025 |
| decreasing particle size | 0.03 |
| Photometry of minerals | 0.005 | 0.010 |
| Increasing surface granularity | 0.002 | 0.009 |
| Ferric coatings | 0.02 | 0.004 |

* From Fig. 4a.
* From Fig. 4a, with e = 20° to 60°.
* Difference of spectra in Fig. 4a and 4b.
* Basalt from Gradie and Veverka 1982; e = 0° to 30°.
* Basalt from Gradie et al. 1980, extrapolated at 2.5 μm, ϕ = 4° to 30°.
* Reflectance computed for particle diameters of 20 and 100 μm, from Hapke's model, 1981.

Iron-bearing minerals located near 1 μm is diagnostic of surface composition. The short-wavelength component of this absorption is measured at 0.88 μm, which is an average of the band centers observed in different bright regions. The continuum must be extrapolated from longer wavelengths (at 1.07–1.39 μm) so this estimate is not precisely a band depth, explaining the values higher than 1 in Table I. Although this parameter is dominated by Fe³⁺ in ferric oxides, it may also contain a component of an absorption due to Fe⁺⁺ in ortho-pyroxenes (e.g., Hunt and Salisbury 1970, Townsend 1987). The dynamic of this parameter is increased after subtraction of the aerosols, with bright regions having deeper absorptions. Syrtis Major appears more symmetric with respect to its central axis. Although this parameter is based on an assumed shape of the spectral continuum of surface materials, it appears to be very insensitive to small variations of aerosol opacity.

The long-wavelength component of the 1.0-μm band is estimated here as a band area, evaluated between the spectrum and a base-line between the long wavelength shoulder (at 1.09 μm) and the average center of the Fe⁺⁺
band (at 0.86 μm). This band area is a first order estimate of the absorption due to Fe²⁺ in mafic minerals, but it would be also affected by some ferric oxides absorbing at very high wavelengths, like lepidocrocite. This parameter is much more sensitive than the 0.86-μm band depth to the removal of the aerosol contribution. This transformation is almost a linear stretch in the scale of differences in the band area, however, and the uncertainty in the magnitude of aerosol opacity does not prevent discrimination of areas having shallow or deep absorptions.

The strength of the pyroxene band at 2 μm is measured assuming the band center to occur at 2.17 μm, which is representative of band centers in a variety of dark regions. The band has been attributed to pyroxenes of different composition, generally dominated by augite (Mustard et al. 1993b). This band occurs at longer wavelengths where scattering is much less important, so that removal of the aerosol contribution leaves the dynamic range of band strength nearly unchanged. The spatial distribution of variations in band strength once again also remains unchanged.

Since the continuum on the long-wavelength side of the 3 μm water absorption is far outside the spectral range of ISM, the absorption is evaluated as the ratio of intensities within the band (from 2.96 to 3.06 μm) and on the short-wavelength shoulder (from 2.40 to 2.45 μm). This band is almost systematically deeper in brighter regions (Erard et al. 1991). Under either of the two possible assumptions about the magnitude of backscattering at 2.6 μm (0.00 or 0.01), scattering by aerosol becomes negligible above 2.9 μm (Fig. 4c). Consequently, spatial contrast in the 3-μm water band depth is affected only by the magnitude of scattering on the band shoulder below 2.6 μm. The effect of the aerosols, if any, is therefore to deepen this band in dark regions. We conclude that airborne particles are not responsible for the observed correlation between albedo and strength of the 3 μm hydron band. This correlation may reflect actual compositional differences; alternatively, it may in part be an artifact of the band approaching saturation, in which case differences in albedo on the shoulder of the absorption would produce apparent variations in band depth.

**Band centers.** The wavelength positions of the broad, shallow bands due to Fe-containing minerals are critical to compositional characterization of the surface. For example they indicate that hematite is a major phase in bright materials (e.g., Morris et al. 1989, Bell et al. 1990, Bell 1992, Murchie et al. 1993) and that calcic pyroxenes are a dominant component of dark regions including Syrtis Major (Pinet and Chevrel 1990, Mustard et al. 1993a,b). In the spectra shown in Fig. 8a, these absorption bands are shifted to longer wavelengths by the addition of the scattered light component. The shift is more important at shorter wavelengths, but is also observed for the 2 μm pyroxene band. In contrast, the atmospheric bands remain centered at their nominal wavelengths. This different behavior is not inconsistent: no such effect is expected under conditions of moderate scattering for deep, narrow bands such as those of the atmospheric gases, under the multiplicative approximation discussed in Section 2.1. In the context of a classical radiative transfer model (e.g., Goody 1964) this conclusion remains valid at least for symmetric absorption bands.

We determined 1-μm band centers for the six typespectra both before and after removal of the aerosol contribution by fitting them with cubic splines and then interpolating the reflectance minima. The results are listed in Table III. This measure is more sensitive to the accuracy of absolute calibration perhaps than any of those listed above; however, the calibration is sufficient to resolve shifts in band centers by tens of nanometers. The largest shifts of the band centers are observed on the dark areas, for the absorption due to pyroxene (≈60 nm). On Syrtis Major, the shape of the spectra is greatly modified by the removal of the aerosols. The band itself, already visible in the corrected spectra, is further emphasized in ratioed spectra (Fig. 9).

The wavelength positions of the 1 and 2 μm pyroxene absorptions have been used by Mustard et al. (1993a,b) to infer the composition of Syrtis Major volcanics. The 1-μm band centers were measured on the spectra divided by their own continuum (fitted at 0.79 and 1.26 μm) and were found to range between 0.96 and 0.98 μm in this region. After we subtract the scattered light contribution with its highly negative spectral slope, band centers are located at only slightly shorter wavelengths, ranging from 0.93 to 0.97 μm (Table III). Thus, even with our extreme assumption about the effects of scattering, the shift in band center does not dramatically change the compositional interpretation by Mustard et al. The absorptions

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**TABLE III**

Radiance Factors at 1.09 μm and Center of the 1.0 μm Band in the Average Spectra of Fig. 8

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<th>Calibrated spectra</th>
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</tr>
<tr>
<td>Isidis Planitia</td>
<td>0.316</td>
<td>0.863</td>
</tr>
<tr>
<td>Isidis Fossae + Lybia</td>
<td>0.297</td>
<td>0.916</td>
</tr>
<tr>
<td>Syrtis E</td>
<td>0.223</td>
<td>0.980</td>
</tr>
<tr>
<td>Syrtis E-center</td>
<td>0.170</td>
<td>1.036</td>
</tr>
<tr>
<td>Syrtis center</td>
<td>0.143</td>
<td>1.031</td>
</tr>
<tr>
<td>Syrtis W</td>
<td>0.137</td>
<td>0.998</td>
</tr>
</tbody>
</table>
at 1 and ≈2.1 μm still lie in the domain of clinopyroxenes, although the pyroxenes would be slightly poorer in calcium (Ca/(Fe + Ca + Mg) ≈0.22 ± 0.07 instead of 0.275 ± 0.075) and richer in iron (Fe/(Fe + Ca + Mg) ≈ 0.35 ± 0.1 instead of 0.3 ± 0.1), and thus perhaps be enriched in pigeonite. The difference in band centers is small enough to validate the methods used by Mustard et al. under conditions of small opacity.

The shift of the band centers is smaller on bright areas (≈25 nm) because the relative contribution of scattered light is smaller. A global survey of bright soils using ISM data was performed by Murchie et al. (1993) using seven image cubes covering Amazonis, Tharsis, Valles Marineris, Isidis, and Arabia. The bright soils uncontaminated by dark materials proved spectrally heterogeneous. Three groups of regions were identified as having characteristic 0.9 μm bands: Tharsis, Amazonis, and portions of the Isidis region (band center at 0.84–0.86 μm); Ascræus Mons (0.86–0.88 μm); and Arabia (0.92 μm). The first two groups are very similar in all other respects, and were interpreted as altered materials spectrally dominated by hematite, with the difference in band centers tentatively ascribed to differences in particle size. It results from the present work that a slightly larger aerosol opacity over Ascræus Mons could also account for this difference. Conversely, the material exposed in Arabia was interpreted as a mixture of hematite with another ferric mineral absorbing at longer wavelength (e.g., goethite, ferricydrite, jarosite).

The data used by Murchie et al. were uncorrected for aerosol scattering, so it is instructive to determine whether the heterogeneities they detected can be explained by this effect. Subtracting our estimate of the scattered light contribution from bright region spectra shifts the band center by no more than ≈25 nm, such that the Arabia spectra are still not compatible with a ferric mineralogy dominated by hematite alone. Furthermore, a band center in Arabia located at a longer wavelength than in Tharsis due to the effects of aerosols would necessarily be accompanied by a more negative spectral slope. In contrast, Arabia exhibits a less negative slope than bright materials on Tharsis. We concur with the interpretation of Murchie et al.; there are different Fe mineralogies in Tharsis and Arabia, possibly resulting from a different style of chemical weathering.

In summary, backscattering by airborne particles appears to have a strong influence on the NIR continuum, and small changes in aerosol opacity or size distribution may result in large spectroscopic effects. However, it is possible to define a set of spectral parameters describing the surface absorptions, by taking into account the local continuum. Among the surface spectral properties examined in this study, only variations in the 1.0 μm Fe2+ band are significantly affected by the scattered light contribution, but they are dominated by surface materials. Variations in the spectral slope itself can be comparably influenced by scattering and surface properties.

4.3. Implications for Comparison of Martian Spectra with Laboratory Analogs

The smoothness of martian spectra in the wavelength region from 0.4 to 0.7 μm led to the idea that Fe2+ occurs in bright regions only in poorly crystalline form, such as palagonite (e.g., Singer 1982). Recent telescopic observations (Bell et al. 1990, Bell 1992) at higher spectral resolution have demonstrated evidence for small Fe3+ absorptions at 0.65 and 0.86 μm, attributed to hematite in somewhat coarser grains. These features could be subdued by the addition of a scattered light component, especially in dark areas. However, we do not feel that such an effect calls into question the long-accepted concept of low crystallinity of ferric components or the lesser abundance of ferric minerals in dark regions (Fig. 8b).

The Fe2+ feature at 1 μm was first recognized on Mars by Adams and McCord (1969). This feature, whose occurrence is generally localized in small areas, is difficult to observe at low spatial resolution in telescopic data. It appears moreover to be very efficiently masked by scattering, even under conditions of moderate aerosol opacity: we estimate it to be reduced in strength by almost a factor of two in our data covering Syrtis Major. Thus the effects of aerosols may help to explain why the mafic features are not always observed from the ground. It is also important to note that very calcic pyroxenes can have their bands shifted up to 1.05 μm by the addition of a highly negatively sloped scattered light component. In this case they could be mistaken for olivine, especially in data extending to only ≈1.1 μm in which the longer wavelength wing of the olivine band is not visible (e.g., Huguenin et al. 1977, Pinet and Chevrel 1990). No valid
detection of olivine can thus be accomplished using ground-based data covering that spectral range. It is also clear that simple ratios of intensities at two wavelengths (i.e., slope indices) actually measure a combination of band depth and spectral slope, partly due to atmospheric scattering. Around 1 μm where the scattered light spectrum is highly variable, such indices are poor indicators of surface mineralogy.

Singer and McCord (1979) attempted to isolate the spectrum of dark material unresolved in the telescopic data. They assumed that dark regions actually consist of two subpixel units, bright and dark materials. For such a “checkerboard model,” the resulting spectrum is approximately a linear combination of the end-members spectra, where the bright material spectrum is assumed to be identical to that of the most uniform bright regions. The conclusions of this work were that one single type of dark material cannot explain the spectral diversity of the martian surface and that the 1 μm Fe²⁺ feature and the negative infrared spectral slope are due to dark materials.

The first conclusion is validated by more recent, high-resolution data (Pinet and Chevrel 1990, Bell 1992, Mustard et al. 1993b), but the second one can be contested at least concerning spectral slope. From Eq. (2), and assuming a uniform aerosol opacity, the Singer and McCord estimates of “pure” dark material spectra (Fig. 4) would still contain the same amount of scattered light as the “unmixed” dark region spectra. In addition, their reference bright spectrum has a lesser spectral slope than their dark spectrum; if it represents areas having a lower aerosol opacity, then the aerosol contribution to unmixed dark material spectrum would be even greater. In any case, the spectral slope ascribed to dark materials is due at least in part to the unremoved effects of atmospheric scattering. Subsequent laboratory work has led to the reproduction of the spectral slope of those dark spectra, using thin oxidized coatings on dark substrates (Singer and Roush 1983, Fischer and Pieters 1993). However, Table II demonstrates that the aerosols have a comparable influence on this quantity. We conclude that, although surface mechanisms are still important in explaining the spectral behavior of dark areas, atmospheric backscattering is responsible for a large portion of their negative infrared spectral slope and may be partly responsible for its spatial variations.

CONCLUSIONS

NIR spectroscopy has provided major insights to the mineralogy of Mars. Airborne particles also contribute to the reflectance of the planet at these wavelengths, and this effect is often neglected. We used spectra from the ISM instrument on Phobos 2, covering the wavelength range 0.76–3.16 μm, to investigate the microscopic (composition, size distribution) and macroscopic (vertical and horizontal abundance) properties of these particles, and their effect on the spectra of surface materials (modification of the spectral continuum and absorption features). This work relies on the single-scattering approximation of Sobolev (1975). A further assumption is that under low opacity (τ = 0.2) and near-normal geometry (φ = 15°, μ ≥ 0.7), the dominant effect of aerosols from 0.7 to 2.6 μm is backscattering rather than extinction.

Study of two regions with overlapping coverage at different viewing geometries, composed of 83 and 354 pixels respectively, yielded two estimates of the aerosols’ spectra. These are consistent with independent results from ISM (Drossart et al. 1991), with optical parameters retrieved from IRTM (single-scattering albedo w = 0.9, asymmetry parameter of the phase function g = 0.6; Clancy and Lee 1991), and with the size distribution of aerosols derived from the Auguste experiment on Phobos 2 (effective radius = 1.2 μm and effective variance = 0.2; Koralev et al. 1993). These spectra contain absorption features due to water-ice and perhaps, in one case, hydrated clay minerals and/or gypsum. Additional absorptions due to ferric oxides and clay are not clearly identified, but neither can they be ruled out. In areas of nearly uniform surface properties observed at widely varying elevations, no large variation in aerosol opacity is observed. This implies that for a clear atmosphere, the scattering particles either form a high-altitude layer or consist mostly of uplifted dust likely to be concentrated in the first few meters above the surface.

The magnitude of the scattered light contribution from aerosols decreases with increasing wavelength. At the low atmospheric opacity at the time of ISM observations, between 0.7 and 2.0 μm, it represents 5–15% of the reflectance of bright regions and 15–30% of that of dark regions. Because of this large difference in the relative importance of scattered light, spectral slope must be measured as a derivative of reflectance. If measured as the ratio of intensities at two wavelengths, as is usually done, there results a spurious correlation of spectral slope with albedo. The spectral slope appears to be comparably influenced by aerosol scattering and surface properties; thus spatial differences in spectral slope can be ascribed to the surface only using data acquired under similar opacities and emergence angles. Similarly, because simple intensity ratios are very sensitive to aerosol opacity, they are poor indicators of the strengths of absorption features arising from surface minerals, especially the 1 μm absorption due to iron-containing minerals.

The major absorptions due to surface minerals can be represented using four parameters that are only minimally sensitive to the additive component of light scattered by aerosols under low-opacity atmospheric conditions. These absorptions are subdued and shifted to longer
wavelengths, but spatial differences in their strengths and positions are not strongly modified. The most sensitive parameter is area of the 0.9–1.05-μm Fe^{2+} absorption in dark regions, which is generally attributed to pyroxenes. Band strength is reduced by a factor of two even under conditions of low atmospheric opacity. The center of this band is also shifted to longer wavelengths, typically by ≈50 nm. Mineralogical interpretation of the surface is not dramatically changed: compared with a previous study of pyroxene absorptions in Syrtis Major, aerosol-corrected spectra are still compatible with calcic pyroxenes, although inferred abundance of Ca is probably smaller (Ca/(Fe + Ca + Mg) ≈0.22 ± 0.07). Conversely, the spectral effect of aerosols on bright regions is small and cannot explain spatial differences in the center of the 0.8–0.9 μm ferric band, which requires variability in ferric mineralogy. The effect on the 3 μm band due to water in hydrated minerals cannot be studied under the assumptions made in this study, but it seems to be very small.

Finally, limb-darkening of the planet at 0.8–2.6 μm at small phase angles can be fitted by a Minnaert exponent of 0.6–0.8, which increases with wavelength, albedo and phase angle. The surface itself does not appear to be a Lambertian reflector (κ = 0.8), especially in the dark regions.

Although these results are very important to determining the nature of martian aerosols and their effects on observations of the surface, they represent only a 2-month period of low atmospheric opacity. The OMEGA imaging spectrometer will remain in a high-inclination orbit on board Mars-96 for more than one martian year and is likely to experience more dusty conditions, especially when the planet approaches perihelion. Consequently, the effects described here will have to be taken into account. Analysis of the OMEGA data will benefit from the experience acquired with ISM, and the instrument is intended to observe scattering at the limb with better spatial resolution and more extensively than did ISM. Expected results include identification of the mineralogical composition and size distribution of airborne particles as functions of altitude, latitude, and season, not determined by Phobos-2 due to untimely loss of the spacecraft. These future observations should also yield further insight into the effect of atmospheric scattering on the observation of the surface.

**APPENDIX**

We adopt the expansion given by Sobolev (1975) for the single-scattering approximation. It was first applied to ISM observations by Combes et al. (1991) who discuss it in further detail. This approximation was checked against a multiple-scattering model by Drossart et al. (1991) and proved satisfactory under the conditions in which ISM observations were made. Under conditions of low opacity, and outside the atmospheric bands, this model gives the radiance factor as

\[
r_{\text{rot}} = \frac{L_D}{E_D} = r_1 r_s \left( \frac{\mu_0 + \mu}{\mu \mu_0} + \frac{w f(\phi)}{4\mu} \right) + \frac{r_1}{2} \left( \frac{\pi (\mu_0 - \mu)}{\mu} + \frac{2r_1}{\mu_0} \int_{-\mu}^{+\mu} f(-\mu) \, d\mu \right),
\]

where \( L \) is the observed radiance, \( E_D \) is the solar irradiance at the distance of Mars, \( \mu_0 \) and \( \mu \) are the cosines of the incidence and emergence angles, respectively (relative to the local vertical), \( r_1 \) \( f(\phi) \) and \( w \) are, respectively, the vertical opacity, phase function, and single-scattering albedo of the aerosols, and

\[
I(\mu_0) = \int_{-\mu_0}^{\mu_0} \rho(\mu, \mu_0) \, d\mu_0
\]

where \( \rho(\mu, \mu_0) \) is the phase integral.

The first term \( r_1 \) is the reflectance of the surface alone. The second term represents the light that was scattered and subtracted from the incident beam before reaching the instrument (extinction). The third term accounts for the light scattered toward the observer, that never reached the surface (atmospheric path radiance). The fourth term quantifies the light both scattered by the aerosols and reflected by the surface, plus the light scattered once and reflected twice (indirect illumination).

In sections 2.1 and 3.1, we note for conciseness \( a = I(\mu) + r_1 \int_{-\mu}^{+\mu} f(-\mu) \, d\mu \).

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