TABLE 1. Range of variation of the spectral criteria in the Syris-Isidis image cube.

	Surface + Scattering (Measured)			Surface Alone (Modeled)		
	Min	Max	Range/rms Noise	Min	Max	Range/rms Noise
Reflectance at 1.09 μm	0.115	0.33	1200	0.08	0.30	1200
Spectral slope reflectance units/μm	−0.025	0.0002	35	−0.010	0.020	30
2-μm pyroxenes (band depth)	0.97	1.00	20	0.98	1.01	20
0.9-μm Fe^{3+} (band depth)	0.965	1.055	50	0.955	1.055	60
1-μm Fe^{2+} (band depth)	0.0	0.40	85	0.0	0.70	150

[1]. The main hypotheses in this procedure are that (1) the aerosol contribution is assumed to be zero at 2.6 μm, (2) the photometric function of the surface is assumed independent of the wavelength, and (3) the difference between extinction and forward scattering is supposed to be small compared to backscattering. Although there is some uncertainty about the absolute level of the spectra, the result is reasonable in terms of opacity ($\tau \approx 0.23$ at 1.9 μm) and is consistent with the efficace radius of 1.2 μm derived from Phobos 2/Auguste [3].

The image cube covering Syrtis Major and Isidis Planitia (400 × 3000 km^2), which is the more contrasted one, was used to test the effect of the scattering on the spectra. Statistical analysis of these 3000 spectra showed that the vertical scattering contribution can be considered uniform on the whole region, with a value of 0.75× the estimate derived on Tharsis, and no altimetric dependence. This correction is probably overestimated on the darkest areas.

Surface Properties: The six main spectral criteria related to the surface materials are the reflectance at 1.09 μm; the spectral slope, estimated here as the derivative between 1.84 and 2.35 μm; the strength of the 2-μm pyroxene band; an equivalent of the depth of the 0.88-μm ferric oxide band; and the surface of the Fe^{2+} band, integrated from 0.86 to 1.09 μm. They were computed on the whole image cube for the calibrated spectra and for the spectra corrected from scattering (estimate of the surface alone); their values are given in Table 1. The last three criteria were also used to establish a classification of the calibrated spectra [4]; the main spectral types are given in Fig. 1a. The spectra of the same pixels were averaged after subtraction of the aerosols contribution, and are given in Fig. 1b.

Discussion: The level of the scattered spectrum below 2 μm ranges from 0.02 to 0.05, which represents 5–15% of the albedo of the bright areas and 15–30% of that of the dark regions. This large contribution of scattering is similar to that infered for "clear atmosphere" from IRTM observations [5] and independent ISM analysis [6]. The negative spectral slope often ascribed to dark materials appears to be largely due to the aerosol scattering continuum, although intrinsic variations are recognized on the data. The opposition between the eastern and western parts of Syrtis Major is enhanced by the correction.

The three criteria measuring the main surface absorptions are relatively insensitive to the addition of a low-opacity scattered component. Although they are subdued and slightly shifted to longer wavelengths, their spatial distribution and relative contrast are not deeply modified. The Fe^{3+} and 2-μm pyroxene bands appear to be very insensitive to small variations of opacity. Conversely, the Fe^{2+} band area is reduced by almost a factor of 2, although the transformation is almost a linear stretch of the scale. This reduction may help explain why the mafic features are not always observed from the ground.

We measured the centers of the 1-μm band on the spectra of Fig. 1. On the corrected spectra they are systematicaly shifted by some tens of nanometers toward the short wavelengths, but the shift is not large enough to change dramatically the mineralogical interpretation of the surface: The absorptions are still compatible with hematite on bright regions (center at 0.86 μm) and with calcic pyroxenes in Syrtis Major (center at 0.94 μm), though probably less rich in Ca than previously inferred from ISM [7] (Ca/(Fe + Ca + Mg) ≈ 0.20 ± 0.08 instead of 0.275 ± 0.075). Very calcic pyroxenes could have their bands shifted up to 1.05 μm by the addition of a steep scattered component. In this case they could be mistaken for olivine, so no detection of this mineral could be validated unless made under very low opacity.

References: [1] Erard et al. (1993) *LPS XXIV*, 445. [2] Drossart et al. (1991) *Annal. Geophys.*, *9*, 754. [3] Korablev et al. (1993) *Icarus*, *102*, 76. [4] Erard et al. (1993) *LPS XXIV*, 443. [5] Clancy and Lee (1991) *Icarus*, *93*, 135. [6] Erard et al. (1992) *LPS XXIII*, 335. [7] Mustard et al. (1993) *JGP* 98, 3387

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N94-33205

A MODEL FOR THE EVOLUTION OF CO_2 ON MARS.

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Our MSATT work has focused on the evolution of CO_2 on Mars. We have constructed a model that predicts the evolution of CO_2 on Mars from a specified initial amount at the end of the heavy bombardment to the present. The model draws on published estimates of the main processes believed to affect the fate of CO_2 during this period: chemical weathering, regolith uptake, polar cap formation, and atmospheric escape. Except for escape, the rate at which these processes act is controlled by surface temperatures that we calculate using a modified version of the Gierasch and Toon [1] energy balance model. The modifications account for the change in solar luminosity with time, the greenhouse effect, and an equatorial (as well as polar) energy budget. Using published estimates for the main parameters, we find no evolutionary scenario in which CO_2 is capable of producing a warm (global mean temperatures >250K) and wet (surface pressures >30 mbar) early climate, and then evolves to present

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conditions with ~7 mbar in the atmosphere, <300 mbar in the regolith, and <5 mbar in the caps.

Such scenarios only exist if the early Sun was brighter than standard solar models suggest, if greenhouse gases other than CO_2 were present in the early atmosphere, or if the polar albedo is significantly lower than 0.75. However, these scenarios generally require the storage of large amounts of CO_2 (>1 bar) in the carbonate reservoir. If the warm and wet early Mars constraint is relaxed, then we find best overall agreement with present-day reservoirs for initial CO_2 inventories of 0.5–1.0 bar. We also find that the polar caps can have a profound effect on how the system evolves. If the initial amount of CO_2 is less than some critical value, then there is not enough heating of the poles to prevent permanent caps from forming.

Once formed, these caps control how the system evolves because they set the surface pressure and hence the thermal environment. If the initial amount of CO_2 is greater than this critical value, then caps do not form initially, but can form later on when weathering and escape lower the surface pressure to a point where polar heating is no longer sufficient to prevent cap formation and the collapse of the climate system. Our modeling suggests this critical initial amount of CO_2 is between 1 and 2 bar, but its true value will depend on all factors affecting the polar heat budget.

References: [1] Gierasch P. J. and Toon O. B. (1973) *J. Atmos. Sci., 30,* 1502–1508.

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S16-91 ABS ONLY

POSSIBLE TEST OF ANCIENT DENSE MARTIAN ATMOSPHERE. W. K. Hartmann[1] and S. Engel[2], [1]Planetary Science Institute, Tucson AZ 85705, USA, [2]Lunar and Planetary Lab, University of Arizona, Tucson AZ 85721, USA.

We have completed preliminary calculations of the minimum sizes of bolides that would penetrate various hypothetical martian atmospheres with surface pressures ranging from 6 to 1000 mbar for projectiles of various strengths (weak icy comet, carbonaceous bodies, coherent chondrite, iron). The calculations are based on a program kindly provided by C. Chyba [1]. These numbers are used to estimate the diameter corresponding to the turndown in the crater diameter distribution due to the loss of these bodies, analogous to the dramatic turndown at larger size already discovered on Venus due to this effect.

We conclude that for an atmosphere greater than a few hundred millibars, a unique downward displacement in the diameter distribution would develop in the crater diameter distribution at D ~ 0.5–4 km, due to loss of all but Fe bolides.

Careful search for this displacement globally, as outlined here, would allow us to place upper limits on the pressure of the atmosphere contemporaneous with the oldest surfaces, and possibly to get direct confirmation of dense ancient atmospheres.

We are currently searching for support to refine the calculations and conduct the necessary careful search in the cratering records.

References: [1] Chyba C. (1993) *Nature, 361,* 40.

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GROUNDBASED MONITORING OF MARTIAN ATMOSPHERIC OPACITY. K. E. Herkenhoff[1] and L. J. Martin[2], [1]Mail Stop 183-501, Jet Propulsion Laboratory, California Institute of Technology, Pasadena CA 91109, USA, [2]Lowell Observatory, Flagstaff AZ 86001, USA.

The amount of dust in the martian atmosphere is variable in both space and time [1,2]. The presence of aerosols in Mars' atmosphere complicates quantitative analysis of martian surface properties [3–7]. Dust storms have been observed telescopically for almost 200 yr and are known to have major effects upon the structure and circulation of the martian atmosphere [8,9]. Great dust storms tend to occur during the southern spring and summer [2] and may be an important mechanism by which dust is transported into the polar regions [10]. It is widely believed that the martian polar layered deposits record climate variations over at least the last 10–100 m.y. [11–18], but the details of the processes involved and their relative roles in layer formation and evolution remain obscure [19]. The layered deposits are widely believed to be the result of variations in the proportions of dust and water ice deposited over many climate cycles [13–15]. However, the amount of dust currently transported into the polar regions is unknown, as are the effects of global climate changes on dust transport. In order to infer the climate history of Mars from geologic evidence including the polar layered deposits, the current cycling of dust through the martian atmosphere must be understood. In addition, future missions to Mars (including possible human exploration) will require better knowledge of the likelihood and severity of martian dust storms.

Zurek and Martin [2] found that "planet-encircling dust storms do not occur every Mars year, and . . . that there may have been periods of several successive years without such storms." The clarity of Mars images taken during recent oppositions suggests that the martian atmosphere has been less dusty recently than in previous years [20]. Hubble Space Telescope images of Mars show that the dust opacity was less than 0.06 in December 1990 [21]. Ingersoll and Lyons [22] proposed that martian great dust storms are chaotic phenomena, influenced by the amount of "background" dust in the atmosphere. However, their analysis was hindered by gaps in the historical record of martian dust opacity. Martian dust storms can be detected only when Mars is relatively close to Earth, so a complete seasonal or interannual history of dust storms is impossible to obtain from groundbased data alone. The optical depth of aerosols in the martian atmosphere between dust storms has been determined primarily from spacecraft data [23,24], but can also be inferred from groundbased observations [5]. Groundbased images of Mars show that atmospheric dust opacity significantly affects the photometric behavior of the planet. Lumme [5] modeled martian limb brightening using high-quality visible-light photographs taken on September 3, 1973. The optical thickness (0.16) and single-scattering albedo (0.55) at 435 nm that he derived are consistent with more recent results using Viking Orbiter violet-filter television data [4,6,25,26], indicating that groundbased data may be used to determine the scattering properties of the martian atmosphere. As shown below, Mars limb brightening data can be used to determine the opacity of aerosols in the martian atmosphere between dust storms.

We have developed a model for Mars surface and atmospheric scattering based on equations (1)–(6) in Hillier et al. [27]. This formulation was chosen for its speed of computation and because it accounts for the spherical geometry of atmospheric scattering at