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VERY HIGH ELEVATION WATER ICE CLOUDS ON MARS: THEIR MORPHOLOGY AND TEMPORAL BEHAVIOR; F. Jaquin, Cornell University

Quantitative analysis of Viking Orbiter images of the martian planetary limb has uncovered the existence and temporal behavior of water ice clouds that form between 50 and 90 km elevation.[1] These clouds form a thin haze layer over most of the planet during southern spring and summer when Mars is near perihelion. At other times of year this high elevation haze is absent. As well as seasonal control, the cloud has a strong diurnal dependence, being observed in the early morning, but not in the afternoon. A radiometric inversion indicates that the optical depth of this high elevation haze is less than 0.05 and may contain a few hundredths of a precipitable micron of water. Enhanced vertical mixing of the atmosphere as Mars nears perihelion is hypothesized as the cause of the seasonal dependence, and the diurnal dependence is most easily explained by the temporal behavior of the martian diurnal thermal tide. The small water content of this high elevation haze indicates that the haze layer is unimportant with regard to volatile storage or transport. However, the seasonal dependence of the haze provides preferential protection from sunlight to the southern polar cap during southern summer. This may be important in maintaining the cap through the summer.

Viking Orbiter images of the martian limb provide a seasonally and latitudinally complete data set regarding the vertical distribution of aerosols in the martian atmosphere. Sunlight reflected from aerosols above the planetary limb can be measured with a radiometric uncertainty of about 7%, and elevations above the 6.1 millibar pressure surface can be calculated to an accuracy of a few kilometers. Thus, the temporal and spatial distribution of aerosols can be characterized.

The geometric nature of limb viewing introduces a multiplicative constant of about 46 between the vertical optical depth to any level in the atmosphere and the corresponding line-of-sight optical depth. This large factor limits the depth to which information regarding the vertical distribution of aerosols can be measured to those levels above a vertical optical depth of about 0.1. Below this level no information can be retrieved concerning the vertical distribution of aerosols. During periods of very low atmospheric opacity, the aerosol distribution of the entire atmospheric column can be measured. However, in practice, atmospheric opacities are large enough that the line-of-sight optical depth limit is reached at about 30 km elevation. This limit is only a weak function of wavelength, because of the large multiplicative constant. Therefore, these and future orbital observations of the aerosol distribution are limited to elevations above three scale heights. Above this limit a radiometric inversion has been used to retrieve the true vertical aerosol distribution. The inversion assumes spherical symmetry, the aerosol single scattering albedo, asymmetry factor, and surface reflectance properties. Figure 1 illustrates the limb-viewing geometry.

Morning limb profiles from southern spring and summer display a characteristic morphology of an extended detached haze near 60 to 70 km, above a continuous haze that extends to the surface. Afternoon profiles show the detached haze to be absent or diminished in prominence. The

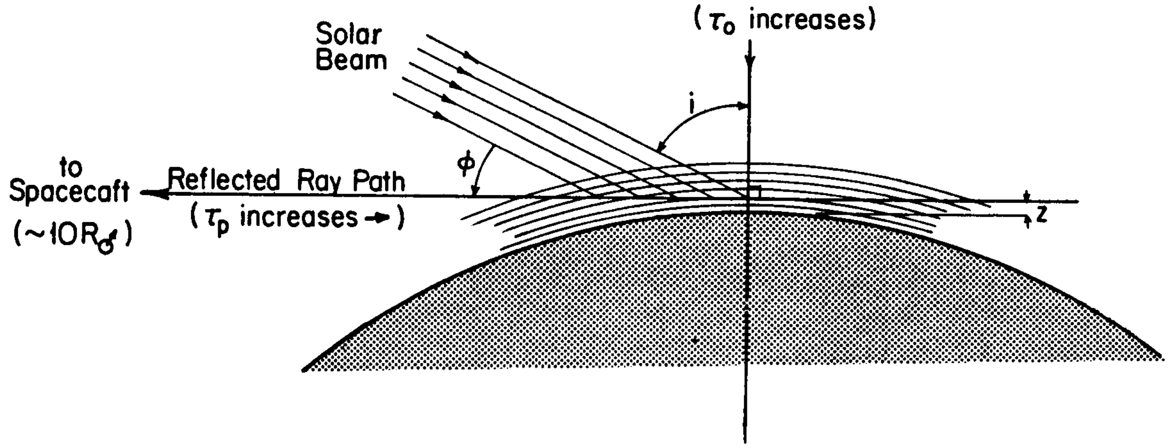

Figure 1. Geometry of limb reflectance measurements. Line-of-sight optical depth τ_p is approximately 46 times larger than the vertical optical depth τ_0 to the same level z.

detached haze often has a layered structure with roughly 5 to 10 km wavelength. Figure 2 illustrates a typical limb profile of the type discussed.

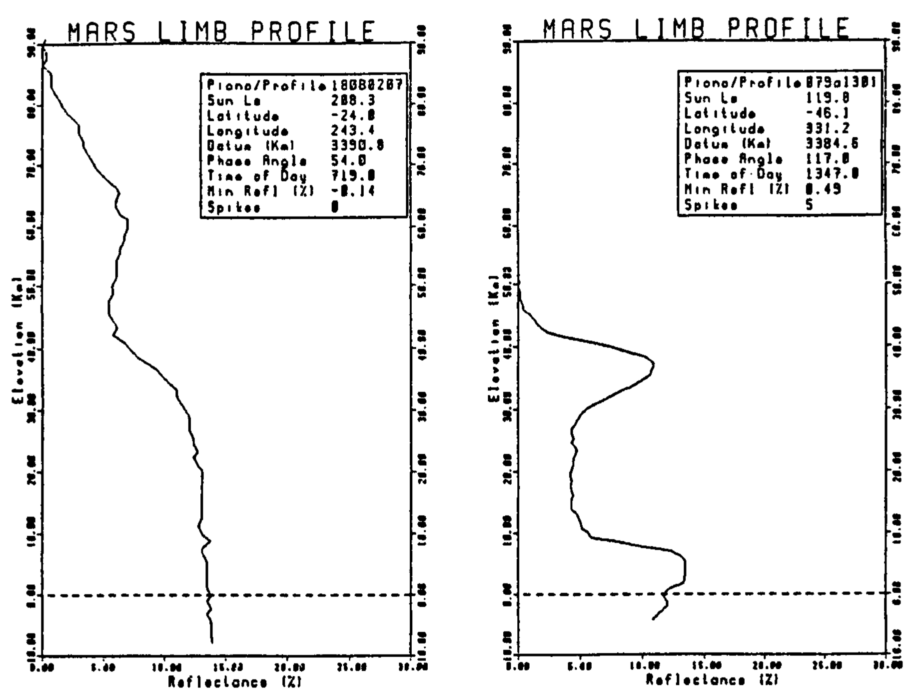

Figure 2. Typical limb reflectance profiles in the Mars atmosphere: (a) illustrating high elevation condensate haze between 85 and 60 km discussed in text, and (b) a lower elevation condensate haze typical of southern mid-latitudes during winter and fall.

The detached haze is inferred to be a water ice condensate based on morphology and atmospheric thermal structure. The detached morphology of the high elevation hazes is most easily explained as a natural result of a condensate haze that forms at some level in the atmosphere. The detached morphology is very difficult to explain if the aerosol is a non-volatile material (i.e., dust particles) that is being transported through the atmosphere in some narrow layer. The diurnal behavior of the detached haze supports the condensate origin of the aerosol. That the condensate is water rather than CO_2 is inferred from the prohibitively high temperatures for CO_2 condensation observed in the martian atmosphere. Temperature profiles from the Viking Landers during entry into the atmosphere, as well as derived temperature profiles from the ε Geminorum stellar occulation, indicate atmospheric temperatures in excess of the required 102 K required for CO_2 condensation at these levels.[2,3] These measurements were made during a different time of the year, but indications are that the upper atmosphere of Mars is warmer during southern spring and summer than during the rest of the year.[4,5] Measured water vapor abundance of near 10 precipitable microns, if uniformly mixed throughout the atmospheric column, is more than adequate to allow condensation at high elevation.[6]

As indicated in Figure 3, high elevation clouds are observed at most latitudes during southern spring and summer. The coincidence of the appearance of this haze around perihelion suggests that the increased insolation drives a more vigorous vertical mixing, that lifts water vapor to high elevations. Further analysis indicates that the detached haze is more often observed in morning profiles rather than afternoon profiles. This is most readily attributed to the action of the diurnal thermal tide that has a maximum amplitude near 60 km early in the morning.[7]

Results of the radiometric inversion indicate that the vertical optical depth of the high-elevation detached hazes is less than or equal to 0.05. Derived extinction coefficients are on the order of 1×10^{-3} inverse kilometers, similar to terrestrial stratospheric aerosols. Assuming a reasonable average radius of 0.1 μm and a reasonable number density of 100 per cubic centimeter, the detached haze contains about 0.01 precipitable microns of water.[8] It is clear that this order of magnitude estimate implies that these high-elevation hazes are unimportant in volatile storage and transport.

The seasonal dependence of this haze preferentially shields the south polar cap from sunlight during the southern summer, and provides no such shielding for the northern cap during its summer. Sunlight traversing the haze layer near normal incidence will suffer no appreciable attentuation due to the low optical depth of the haze. However, at high incidence angles, when a large air mass is traversed, the haze layer may provide significant extinction. The southern polar cap is illuminated by sunlight during southern summer at an incidence angle of near 65°. This high elevation haze layer alone attenuates incident solar flux by $EXP(-\tau/\mu_0) = 0.89$, reducing the surface flux by 11%. This shielding occurs during the peak heating of the south polar cap, and surely influences its final dimensions.

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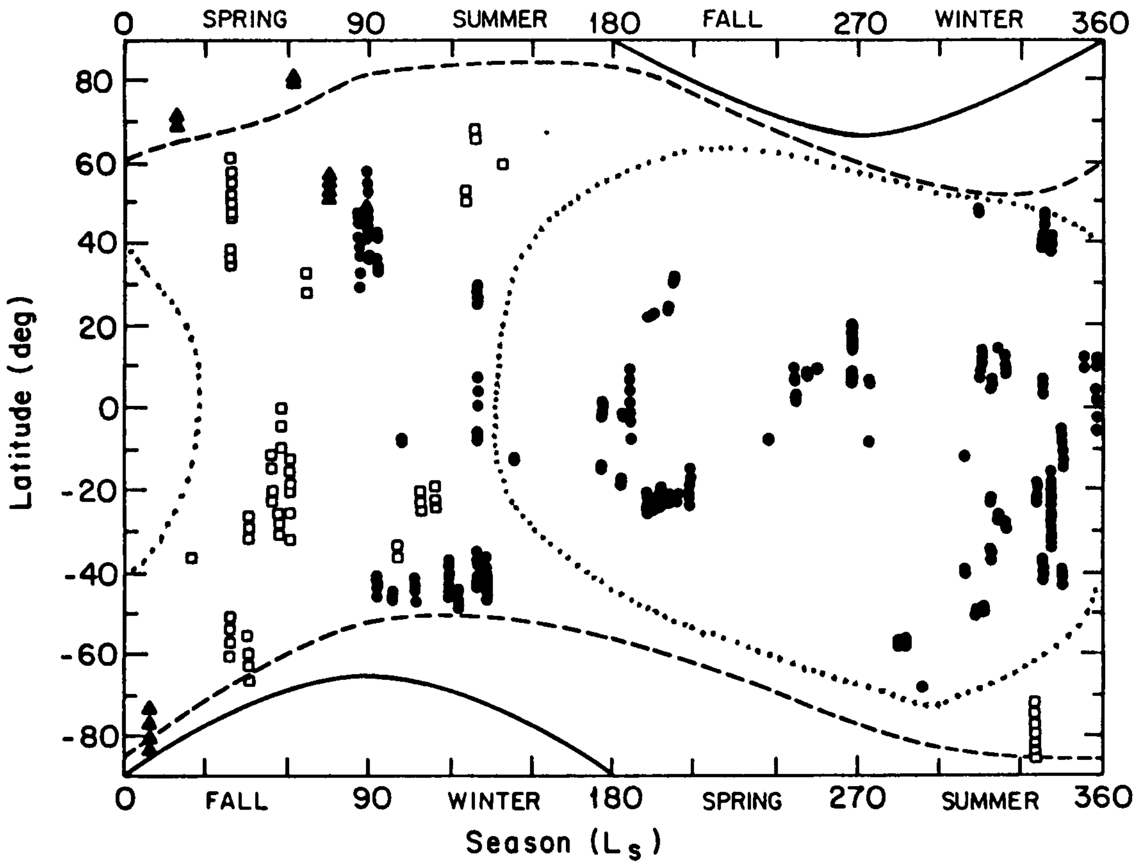

Figure 3. Seasonal and latitudinal distribution of limb profiles. The solid lines bound the area of seasonal night. The dashed line indicates the approximate polar cap edges. Limb profiles with aerosols above 50 km are enclosed by the dotted line.

References

Jaquin, F., Gierasch, P. J., and Kahn, R. (1986, in press) in Icarus.

Seiff, A., and Kirk, D. B. (1977) in J. Geophys. Res., 82, p. 4364-4378.

Elliot, J. L., French, R. G., Dunham, E., Gierasch, P. J., Veverka, J., Church, C., and Sagan, C. (1977) in Astrophysical J., 217, p. 661-679.

McElroy, M. B., Kong, T. Y., and Yung, Y. L. (1977), in J. Geophys. Res. 82, p. 4379-4388.

Martin, T. Z., and Kieffer, H. H. (1979) in J. Geophys. Res. 84, p. 2843-2852.

Jakosky, B. M., and Farmer, C. B. (1982) in J. Geophys. Res. 87, p. 2999-3019.

Zurek, R. W. (1976) in J. Atmos. Sci. 33, p. 321-337.

Toon, O. B., and Farlow, N. H. (1981) in Ann. Rev. Earth Planet. Sci., 9, p. 15-58.