MARS OZONE: MARINER 9 REVISITED

Bernhard Lee Lindner

Atmospheric and Environmental Research, Inc.
840 Memorial Drive
Cambridge, Massachusetts 02139 USA
Telephone: (617)349-2280
Fax: (617)661-6479

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ABSTRACT: The efficacy of the UV spectroscopy technique used by Mariner 9 to remotely measure ozone abundance at Mars is discussed. Previously-inferred ozone abundances could be underestimated by as much as a factor of 3, and much of the observed variability in the ozone abundance could be due to temporal and spatial variability in cloud and dust amount.
INTRODUCTION

Ozone is a key to understanding atmospheric chemistry on Mars. Over 20 photochemical models of the martian atmosphere have been published, and O₃ is often used as a benchmark for these models (e.g., Lindner, 1988; Shimazaki, 1989; Krasnopolsky, 1993). O₃ abundance has been inferred from instrumentation on several spacecraft, with the most complete coverage provided by Mariner 9 (Barth et al., 1973; Lane et al., 1973; Wehrbein et al., 1979). The Mariner 9 UV spectrometer scanned from 2100 to 3500 Angstroms with a spectral resolution of 15 Angstroms and an effective field-of-view of approximately 300 km². The only atmospheric absorption in the 2000 to 3000 Å wavelength region was previously assumed to come from the Hartley band system of ozone (Barth et al., 1973; Lane et al. 1973), which has an opacity of order unity at winter polar latitudes (Lindner, 1988). Therefore the amount of ozone was inferred by fitting this absorption feature with laboratory data of ozone absorption, as shown in Fig. 1 (Lane et al., 1973). Mars O₃ shows strong seasonal and latitudinal variation, with column abundances ranging from 0.2 μm-atm at equatorial latitudes to 60 μm-atm over northern winter polar latitudes (1 μm-atm is a column abundance of 2.689x10¹⁵ molecules cm⁻²). However, the O₃ abundance is never great enough to significantly affect atmospheric temperatures (Lindner, 1991; 1993a) or surface temperatures and frost amounts (Lindner, 1990; 1992; 1993b). Figure 2 (Barth, 1985) shows some of the previously-inferred O₃ abundances.

MODELING PROCEDURE

A radiative transfer computer model is used to re-examine the Mariner 9 UV spectra. The discrete ordinate method of Stamnes et al. (1988) is used to treat the scattering and absorption of monochromatic radiation through the martian atmosphere. O₃ absorption cross-sections from 2000 Å to 3200 Å are included (Daumont et al., 1983; Freeman et al, 1984). Ultraviolet absorption cross-sections for CO₂, O₂, H₂O, HO₂ and H₂O₂ are negligible (e.g., Lindner, 1988; 1991;1993b). The Rayleigh scattering optical depth is computed as in Hansen and Travis (1974), using parameters appropriate for Mars. Atmospheric composition is taken as 95% CO₂ (Owen et al., 1977). The altitude dependence of O₃ is based on model results (Lindner, 1988).
Atmospheric properties are zonally averaged and assumed azimuthally-independent. The region from the surface to 40 km altitude is broken into twenty 2-km-thick layers to account for vertical inhomogeneity. The Chapman function is used to approximate the slant path in place of the secant function (Smith and Smith, 1972), because the winter polar atmosphere always has large solar zenith angles, and the secant function is in error for large angles. This work examines 57°N latitude in late winter ($L_s = 343^\circ$), which is when the maximum O$_3$ column abundance was observed (Barth et al, 1973; Lane et al., 1973). [$L_s$, the solar longitude, is a seasonal index; $L_s$ of 0°, 90°, 180°, 270°, correspond to northern spring equinox, summer solstice, autumnal equinox, and winter solstice, respectively]. The surface is covered by ice with an albedo of 0.6 (Warren and Wiscombe, 1980; Hapke et al., 1981; James and Lumme, 1982; Warren et al., 1990; Lindner, 1993a). The altitude profile of temperature rises linearly from 150K at the surface to 160K at 10 km, and then falls linearly with altitude to 130 K at 40 km, typical for winter polar conditions (Lindal et al., 1979; Kieffer, 1979).

Dust opacities varied from 0.2 to 1.0 for conditions other than global dust storms during the Mariner 9 and Viking observations (Briggs and Leovy, 1974; Pollack et al., 1979; Lumme and James, 1984), although dust opacities over winter polar latitudes may be slightly less [e.g., Lindner, 1990]. Dust opacity varies with wavelength (Zurek, 1978; 1982). The vertical distribution of dust is well-mixed to 20 km altitude (Anderson and Leovy, 1978; Zurek, 1982; Korablev et al., 1993). The single scattering albedo of airborne dust has a solar average of 0.9 (Clancy and Lee, 1991), but is only 0.6 in the 2000 Å to 3000 Å range (Pang et al., 1976; Thorpe, 1977). Scattering of radiation by dust is represented by the Henyey-Greenstein phase function, with an asymmetry parameter of 0.55 (Clancy and Lee, 1991). Typical cloud and fog opacities of 1.0 are taken from theory (Kulikov and Rykhletskii, 1983) and observations (Moroz, 1976; Pollack et al., 1977; Clancy and Lee, 1991), although large seasonal and latitudinal variations exist (Briggs and Leovy, 1974). A single-scattering albedo for cloud particles of 1.0 is taken from observations (Clancy and Lee, 1991). The Henyey-Greenstein phase function with asymmetry parameter of 0.55 is also used to describe the scattering by clouds (Clancy and Lee,
Computational difficulties which accompany highly asymmetric phase functions are removed with the Delta-L method (Wiscombe, 1977).

RESULTS AND DISCUSSION

Using a constant mixing ratio for $O_3$ (e.g., Wehrbein, 1979; Lindner, 1988) and no chemical or radiative interaction between $O_3$ and clouds/dust, Fig. 3 shows that when typical amounts of dust and cloud are present that significant underestimation of $O_3$ abundance occurs. A factor of 3 times as much $O_3$ is needed to generate the same spectrum the spacecraft would measure for a cloudy, dusty atmosphere as for a clear atmosphere. If the scattering properties of martian clouds and dust were well known, then their appearance would not be a problem, as a model would be capable of retrieving the $O_3$ abundance. However, these properties are not well known, which raises doubts about the effectiveness of the UV reflectance spectroscopy technique for measuring $O_3$ abundance on Mars. The simulations shown in Fig. 3 are repeated for a range in solar zenith angle ($50^\circ$-$90^\circ$), ground albedo (0.3-0.8), altitude distribution of $O_3$, satellite viewing geometries, dust scattering properties, and cloud, dust and $O_3$ abundances. A factor of 3 underestimation is typical, with greater underestimation for high ground albedo or high dust opacities. Even if scattering by clouds is properly accounted for (as previously done with Mariner 9 data reduction in [4]), masking by dust can easily result in factor of 2 underestimation. Results are not strongly dependent on solar zenith angle.

Spatial and temporal variability in temperature and water vapor have been claimed to account for the scatter of the data points in Fig. 2 (Barth and Dick, 1974). A decrease in temperature results in a decrease in water vapor, assuming the atmosphere is saturated. A decreased water vapor abundance decreases the availability of odd hydrogen ($H$, $OH$, and $HO_2$), which converts $CO$ and $O$ into $CO_2$ catalytically, decreasing the abundance of $O$ needed to form $O_3$. However, water vapor is a small source of odd hydrogen in the winter polar atmosphere compared to $H_2$, and may not account for most of the variability in Fig. 2 (Lindner, 1988). Masking by clouds and dust may also account for some of the observed $O_3$ variability, because the nature and opacity of the clouds and dust at winter polar latitudes change significantly spatially and temporally. As the
maximum \( O_3 \) abundance resides near the surface (Lindner, 1988), spacecraft must be able to observe through the entire cloud and dust abundance in order to measure the total \( O_3 \) column abundance. If reflectance spectroscopy is used, as on Mariner 9, then the cloud and the airborne dust must be traversed twice; first by the incoming solar flux down to the surface, and then once again upon reflection from the surface out to the spacecraft. In addition, the large solar zenith angles at winter polar latitudes mean several times the vertical opacity of cloud and dust must be traversed. Indeed, part of the observed latitudinal variation in \( O_3 \) abundance in Fig. 2 may be due to the inability of the spacecraft to observe through the increasing effective optical depths as one goes poleward.

By using a photochemical model which included multiple scattering of solar radiation, Lindner (1988) showed that the absorption and scattering of solar radiation by clouds and dust should actually increase \( O_3 \) abundances at winter polar latitudes. Hence, regions with high dust and cloud abundance could contain high \( O_3 \) abundances (heterogeneous chemistry effects have yet to be fully understood [Atreya and Blamont, 1990; Krasnopolsky, 1993]). It is quite possible that the maximum \( O_3 \) column abundance observed by Mariner 9 of 60\( \mu \)m-atm is common. In fact, larger quantities may exist in some of the colder areas with optically thick clouds and dust. As the Viking period often had more atmospheric dust loading than did that of Mariner 9, the reflectance spectroscopy technique may even have been incapable of detecting the entire \( O_3 \) column abundance during much of the Mars year that Viking observed, particularly at high latitudes. The behavior of \( O_3 \) is virtually unknown during global dust storms, in polar night, and within the polar hood, leaving large gaps in our understanding.

Other possibilities for measuring \( O_3 \) abundance include solar occultation (Blamont et al., 1989), IR observations in the 9.6 \( \mu \)m \( O_3 \) absorption band (Espanek et al., 1991), and observations of the \( O_2 \) dayglow at 1.27 \( \mu \)m, produced by photolysis of \( O_3 \) (Traub et al., 1979). However, further studies of these other techniques are required, especially as regards the effects of clouds and dust.
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REFERENCES


FIGURE CAPTIONS

Figure 1. Ultraviolet spectrum measured by Mariner 9 at 57°N latitude on orbit 144 (taken from Lane et al., 1973). To enhance the O₃ absorption feature, this spectrum was divided by one obtained at 20°N latitude on orbit 144, where O₃ abundances are minimal.

Figure 2. Measurements of the O₃ column abundance previously inferred from the Mariner 9 UV spectrometer data during the northern winter, L₉ = 330-360°, in the northern hemisphere (taken from Barth, 1985).

Figure 3. Synthetic spectra as would be observed by spacecraft for atmospheres with no cloud or dust and 30 μm-atm O₃ (solid line), vertical opacities of dust and cloud of 0.3 and 1.0, respectively, and 30 μm-atm of O₃ (dashed line), and vertical opacities of dust and cloud of 0.3 and 1.0, respectively, and 100 μm-atm of O₃ (dotted line). All cases assume a solar zenith angle of 75° (typical for winter polar observations), and a polar cap albedo of 0.6.
UV SPECTRUM AT "B"
57°N

RATIO (57° N) / (20° N)

ORBIT 144

WAVELENGTH (Å)

RELATIVE INTENSITY