Martian North Polar Hazes and Surface Ice: 
Results From the Viking Survey/Completion Mission

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Viking infrared observations of the Martian north polar regions taken during the 1979–1980 Viking Survey/Completion mission have been used to determine the location and nature of the retreating polar cap edge and of an accompanying atmospheric ice haze. The haze is composed of water ice, and its visible opacity is approximately unity, equivalent to 1–2 precipitable microns of water ice. During the polar cap retreat (aerocentric longitude \( L_a \approx 340^\circ\)–60\(^\circ\)), the haze is nearly always present and extends from at most a few degrees south of the polar cap edge to an undetermined distance over the cap interior. The transition from ice-free ground to a predominantly ice-covered surface at the cap edge spans 5°–10° of latitude. The ability of the Viking infrared observations to distinguish haze from surface condensate yields a more accurate seasonal regression for the polar cap than can be determined from visual images. Generally, the infrared observations presented here yield a regression similar to those determined from earth-based and/or spacecraft visual data for the same and for previous northern springs. However, there are differences concerning details of the cap retreat: the thermal data indicate that the Martian seasonal polar cap did extend south of 60°N during this particular late winter and that the mid-spring resumption of the cap retreat occurs at \( L_a \approx 75\).

INTRODUCTION

Condensate hazes in the vicinity of the north polar cap edge around the time of the vernal equinox (\( L_a \approx 340^\circ\)–10°) are well known from both spacecraft [Leovy et al., 1972; Briggs and Leovy, 1974; James, 1979; French et al., 1981] and earth-based observations [Baum and Martin, 1973; Iwasaki et al., 1982]. \( L_a \) is the aerocentric longitude of the sun measured in Mars-centered coordinates such that \( L_a = 0^\circ \) corresponds to the vernal equinox and \( L_a = 90^\circ \) marks the beginning of northern summer on Mars. The dissipation of this north “polar hood” leads to a period during early spring when the polar cap can be clearly seen and clouds are generally thought to be absent. Observations by the Mariner 9 television cameras, however, revealed faint clouds of either CO\(_2\) or water ice over the cap and just south of its edge even during these relatively clear mid-spring periods [Leovy et al., 1973]. These faint clouds are of interest for their own sake as temporary reservoirs of condensate materials and for their potential for obscuring the true location of the polar cap edge beneath them.

The purpose of this paper is to determine the nature and location of these springtime arctic hazes and of the polar cap edge beneath them for the 1979–1980 Viking Survey/Completion mission. The data to be used are essentially north-south infrared scans of northern latitudes on Mars taken by the Viking Orbiter 1 infrared thermal mapper (IRTM) and Mars atmospheric water vapor detector (MAWD) instruments for the period \( L_a \approx 348^\circ–86^\circ \) (August 1979 to April 1980). The IRTM observations consist of thermal observations in four nominally surface-sensing wavelength bands centered at 7, 9, 11, and 20 \( \mu m \) and a single 15–\( \mu m \) channel which senses radiance from a broad atmospheric layer centered near 25 km altitude [Kieffer et al., 1977]. The MAWD instrument measures the differential absorption of reflected solar radiance in and out of water vapor absorption bands near 1.4 \( \mu m \) [Farmer et al., 1977]. The Viking Survey/Completion mission data covered the period of the north polar cap retreat, and we have used the 13 infrared scans which extended, each at nearly constant longitude, from low latitudes up to and beyond the north polar cap edge. As indicated by the small number of scans, infrared data were taken infrequently and at somewhat irregular intervals during this final phase of the Viking orbiter mission. The local times, longitudes, and spatial resolutions of these observations were nearly the same for each individual scan but varied greatly from scan to scan (see Table 1). As a whole, this series of infrared scans provided an excellent opportunity to study the latitudinal variation of atmospheric water vapor or ice and the location of the ice cap edge during the retreat phase. As a demonstration of the method by which these features are determined, we first consider in detail data from just one of the infrared scans.

INTERPRETATION OF THE ORBIT 1243 DATA

Figure 1 shows a theoretical spectrum for a water ice cloud, as computed by Curran et al. [1973]; this is compared with a high-resolution spectrum obtained by the Mariner 9 infrared interferometer spectrometer (IRIS), taken over the Tharsis Ridge on Mars during mid-afternoon in early summer (\( L_a = 90^\circ \)) in 1972. By comparison with the theoretical spectrum, the broad absorption feature observed between 550 and 950 cm\(^{-1}\) is characteristic of water ice clouds [Curran et al., 1973]. Figure 1 also shows the spectral response of the Viking orbiter IRTM bands. Comparison with the theoretical water ice spectrum shows that the 11-\( \mu m \) band is the most sensitive to the presence of water ice, that the 20-\( \mu m \) band is much less sensitive, and that water ice is nearly transparent in the 7- and 9-\( \mu m \) bands. Thus, despite their broadband nature, the IRTM
TABLE 1. Viking Survey/Completion Mission Infrared Scans: Geometry

<table>
<thead>
<tr>
<th>Orbit</th>
<th>Lₙ</th>
<th>Longitudes</th>
<th>θ_max</th>
<th>Slant Range</th>
<th>Emission Angles</th>
<th>Local Times</th>
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<td>1159</td>
<td>348</td>
<td>50°–115°</td>
<td>60°</td>
<td>25,000</td>
<td>20°–70°</td>
<td>1200–1500</td>
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<tr>
<td>1184</td>
<td>1</td>
<td>180°–300°</td>
<td>80°</td>
<td>19,000</td>
<td>5°–60°</td>
<td>0900–1500</td>
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<tr>
<td>1236</td>
<td>26</td>
<td>190°–230°</td>
<td>85°</td>
<td>8,000–12,000</td>
<td>13°–70°</td>
<td>1100–1500</td>
</tr>
<tr>
<td>1243</td>
<td>29</td>
<td>130°–145°</td>
<td>85°</td>
<td>11,000–14,000</td>
<td>30°–70°</td>
<td>1300–1400</td>
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<tr>
<td>1266</td>
<td>40</td>
<td>280°–310°</td>
<td>85°</td>
<td>9,000–13,000</td>
<td>6°–70°</td>
<td>1000–1200</td>
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<tr>
<td>1284</td>
<td>48</td>
<td>80°–120°</td>
<td>85°</td>
<td>10,000–14,000</td>
<td>10°–65°</td>
<td>1000–1200</td>
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<tr>
<td>1293</td>
<td>51</td>
<td>20°–70°</td>
<td>85°</td>
<td>11,000–13,000</td>
<td>6°–70°</td>
<td>1000–1200</td>
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<tr>
<td>1302</td>
<td>56</td>
<td>289°–295°</td>
<td>85°</td>
<td>12,000</td>
<td>12°–75°</td>
<td>1000–1100</td>
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<tr>
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<td>60</td>
<td>197°–211°</td>
<td>85°</td>
<td>12,000</td>
<td>18°–78°</td>
<td>0900–1100</td>
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<tr>
<td>1315</td>
<td>61</td>
<td>160°–181°</td>
<td>8°</td>
<td>12,000</td>
<td>15°–79°</td>
<td>0900–1000</td>
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<tr>
<td>1323</td>
<td>65</td>
<td>90°–105°</td>
<td>80°</td>
<td>12,000</td>
<td>2°–75°</td>
<td>0900–1000</td>
</tr>
<tr>
<td>1337</td>
<td>71</td>
<td>310°–316°</td>
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<td>12,000</td>
<td>5°–80°</td>
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<tr>
<td>1372</td>
<td>86</td>
<td>319°–330°</td>
<td>73°</td>
<td>12,500</td>
<td>8°–76°</td>
<td>0700–0900</td>
</tr>
</tbody>
</table>

θ_max is the approximate northermost latitude viewed on a given sequence. Slant range is the distance (in kilometers) from the spacecraft to the center of the instrument's footprint on the planet. Local times are given in Martian hours (1 = 1/24 Martian day ≈ 3700 s; 1200 = local noon).

Radiations can be used to spectrally detect the presence of water ice clouds given sufficient contrast between the ground and the cloud temperatures [Martin et al., 1979; Hunt, 1979].

Throughout this paper, differences in the IRTRM radiances will be discussed in terms of relative brightness temperatures. Brightness temperatures are defined for each IRTRM wavelength band as the temperature of a blackbody emitting the amount of flux observed by the IRTRM instrument in that wavelength band. Differences in brightness temperatures, referenced to the 20-μm brightness temperature (T₂₀), will be referred to as spectral differences (e.g., T₁₁–T₂₀). Figure 2 shows the T₁₁–T₂₀ spectral difference for a north-south Viking Orbiter (VO) 1 infrared scan on orbit 1243 (Lₙ = 29°) of the Viking Survey mission, covering longitudes 130°–145°W and near 1300 local time. The large T₁₁–T₂₀ negative values together with the nearly zero T₀−T₂₀ differences (Figure 3) indicate the presence of a water ice cloud on the flanks of Olympus Mons (at latitudes 16°–22°N), analogous to the Tharsis water ice cloud observed by the Mariner 9 IRIS [Curran et al., 1973]. Although this particular Viking scan did not obtain simultaneous visual coverage of this site, water ice clouds in this region and during this season (Lₙ = 29°) are well known from both earth-based [Smith and Smith, 1972; Martin and Baum, 1969] and spacecraft data [Leovy et al., 1973; Briggs et al., 1977; Kieffer et al., 1977; French et al., 1981].

Equally interesting, however, is the high-latitude transition from a negative to a positive T₁₁–T₂₀ value in the vicinity of the north polar cap edge. As we will show, this is also consistent with the presence of a water-ice cloud.

This interpretation of the T₁₁–T₂₀ difference would be relatively straightforward if the Martian surface were an ideal blackbody and if there were no other atmospheric absorbers at these wavelengths. However, Mars differs from an ideal blackbody in several aspects. Nonuniform surface temperatures, nonunit thermal emissivity, and the presence of atmospheric dust produce characteristic spectral differences which vary diurnally and between wavelength bands. These signatures have been modeled [e.g., Christensen, 1982] but here must be considered and separated from the signature of ice clouds.

The principal surface property which affects the daytime spectral differences is nonunit thermal emissivity. The emissivity of the nonpolar Martian surface at each wavelength is very closely related to albedo [Christensen, 1982] and decreases nearly linearly with decreasing surface albedo. Since the Planck function varies nonlinearly with wavelength, the spectral differences should then increase with decreasing albedo. Although the magnitude of this difference varies with surface temperature, daytime T₁₁–T₂₀ and T₀−T₂₀ spectral differences of 4 and 10 K, respectively, can be produced. Thus, between 25 and 60°N, the positive T₁₁–T₂₀ differences in Figure 2 are probably due to nonunit surface emissivities. The local maximum in T₁₁–T₂₀ observed at 58°N, for instance, corresponds to the lowest observed albedo (0.21). This maximum observed difference of 2.2 K could be produced by surface emissivities of 0.93 at both 11 and 20 μm, which is consistent with the values found globally for materials of similar albedo [Christensen, 1982].

Atmospheric absorbers will also affect the spectral differences. Mariner 9 IRIS observations indicate that airborne dust is most absorbing in the IRTRM 9-μm band region, somewhat less in the 20- and 11-μm bands, and least of all in the 7-μm band. In terms of the band opacities, τ₂₀ > τ₁₁ > τ₀ > τ₇ for airborne dust, while τ₁₁ > τ₂₀ ≥ τ₀ ≥ τ₇ ≈ 0 for a water ice haze. Thus, given a surface temperature greater than the temperature of some atmospheric haze with moderate or low opacity, T₁₁–T₂₀ will be positive for a dust haze and negative for a water ice cloud. If the ground is colder than the haze, both signatures will reverse. If the ground is very cold relative to the dust haze, T₁₁–T₂₀ can become positive again if, as suggested by the Mariner 9 IRIS [Toon et al., 1977] and Viking IRTRM [Martin et al., 1979; Hunt, 1979], the extinction due to airborne dust is only marginally larger in the 20-μm band than in the 11-μm IRTRM band. On this basis, there is little doubt that the nonpolar negative T₁₁–T₂₀ values observed during midday (i.e., over warm ground) are due to water ice clouds. The negative T₁₁–T₂₀ values at high latitudes are also likely to be due to water ice clouds. If the spectral signature were due to dust, the temperature of the dust haze would have to be considerably warmer than the bare ground at 64°N; this is unlikely. The transition from negative to posi-
tive values of \( T_{11} - T_{20} \) at high latitudes can be simply inter-
pred as due to a water ice cloud whose temperature is less than
the temperature of the bare ground but greater than that of
the increasingly icy ground farther northward. However, qua-
tative simulations of this transition must take into account
the surface contribution to the \( T_{11} - T_{20} \) difference when
both bare ground and surface ice are present.

**Polar Cap Boundary**

The inference that the fraction of the surface covered by
CO\(_2\) ice is increasing from zero at 63°N is based on the rapid
increase in the \( T_8 - T_{20} \) spectral difference shown in Figure 3b.
Note the lack of such a signature for the ice cloud at 14°N,
suggesting that the 9- and 20-μm bands are little, or nearly
uniformly, affected by water ice clouds. The rapid increase in
\( T_8 - T_{20} \) at 63°N marks the first occurrence of surface CO\(_2\)
frost. Bare ground temperatures in this region are warmer
than 210 K (see the \( T_{20} \) curve in Figure 3c), whereas CO\(_2\) frost
must be very close to the sublimation temperature of approx-
itimately 150 K. Again because of the nonlinearity of the Planck
function, viewing a scene whose area includes both relatively
warm bare ground and cold patches of ice will produce sub-
stantial spectral differences. Table 2 gives the expected \( T_8 - T_{20} \)
and \( T_{11} - T_{20} \) differences due to areal mixtures of materials at
kinetic temperatures of 210 and 150 K. Because the thermal
emissivity of CO\(_2\) ice is poorly known, differences have been
computed for CO\(_2\) emissivities of 0.8, 0.9, and 1.0. From Table
2 it is apparent that the predicted spectral differences are close
to those observed, reaching a maximum when 70-80% of the
surface is covered by ice. The spectral differences should de-
crease as more of the surface is covered, finally reaching the
value produced by nonunit emissivities alone when the entire
surface is at the CO\(_2\) ice temperature. The \( T_8 - T_{20} \) differences
poleward of 80°N have not been used to estimate this ice
emissivity, however, since there is considerable scatter in the
values (greater than the single-sample noise of both the 7-
and 9-μm bands which is itself greater than 1-2 K for tempera-
tures below 160 K [see Chase et al., 1978]). A warm dust haze
could produce increasingly large \( T_8 - T_{20} \) values as the surface
became colder and may be present on orbit 1243 over the
most northern latitudes, where the scatter in the \( T_8 - T_{20} \) values
could be due to variable opacity. Whether or not such a dust
haze can fully explain these differences remains to be seen.
Because we have not definitely characterized the orbit 1243
\( T_8 - T_{20} \) values poleward of 80°N, no estimates of the ice emis-
sivities or of the poleward edge of the water ice haze have been
made. Nonetheless, the observations for orbit 1243 (\( L_\alpha \sim 29° \))
are well matched by a CO\(_2\) ice cap first occurring at 63°N
with increasing ice cover poleward of this location. The
\( T_8 - T_{20} \) difference at 67°N is 5 K which, by reference to Table
2, indicates that the surface is covered by 20-30% CO\(_2\) ice at
150 K. Changing the ice temperature to 160 K changes the
spectral difference by less than 0.5 K.

**North Polar Water Ice Clouds**

The \( T_{11} - T_{20} \) data indicate that the polar ice cloud begins at
63°N and is relatively cold compared to the surface. At this
latitude the surface is mostly (90% or more) bare ground at


NORTH POLAR WATER ICE CLOUDS

Fig. 3. Idealized representation of the IRTM spectral differences and brightness temperatures, together with the averaged MAWD water vapor column abundances and 1.4-μm brightness, all observed in the north-south infrared scan taken by Viking Orbiter 1 on orbit 1243 (Ls = 29°). (a) $T_{11}$ vs. $T_{20}$ showing the rapid increase marking the edge of the seasonal CO$_2$ polar cap. (b) $T_{20}$ showing the large decrease observed in Olympus Mons, and the 1.4-μm brightness (dashed line) computed by dividing the average of the two MAWD continuum channel radiances by the cosine of the solar incidence angle and then normalizing to a standard sun-Mars distance.

$T_r \approx T_{20} \sim 230$ K, so the cloud temperature must be below 230 K. $T_{11}$ and $T_{20}$ should be nearly equal when the cloud and surface temperatures are the same. Assuming no surface spectral contrast, this point occurs at 66°N (Figure 3). If the $T_{11}$ vs. $T_{20}$ difference at 90°N is assumed to be due to CO$_2$ ice alone and is taken to be representative of the cap contribution between 90° and 71°N, while the variation between 63°N and 71°N is assumed to be similar in form to the $T_{20}$ vs. $T_{20}$ variation, the cloud contribution is zero at approximately 67°N. If either case the cloud temperature as estimated from the $T_{20}$ data is about 200–210 K in the region 67°–65°N. This temperature most likely decreases poleward due to thermal coupling with the cold surface. Viking radio occultation S (2.3 GHz) and X band (8.4 GHz) measurements taken during the early northern spring ($L_s = 21°–29°$) of the preceding Martian year over the northern polar ice cap and near its edge all show temperatures increasing sharply in the first 3 km above the surface, reaching maxima of 170–185 K in the 3–8 km range, and decreasing to values of 160 K or colder above 20 km [Lindal et al., 1979]. This is consistent with the VO 1 orbit 1243 data showing $T_{11}$ values between 157 and 162 K poleward of 65°N and agrees with our estimates of the cloud temperature if the cloud cools by ~20 K as it extends over the cap interior, as suggested by the circulation models of Haberle et al. [1979].

North of approximately 78°N the $T_{11}$ vs. $T_{20}$ difference drops to a value close to 5 K. This decrease suggests that either the cloud opacity has diminished or else the cloud temperature has cooled sufficiently to decrease the spectral contrast between the surface and the cloud. The $T_{20}$ brightness temperatures (Figure 3c) also show a sharp decrease north of 78°N. This observation is consistent with the poleward dissipation and/or cooling of a relatively warm cloud since the 20-μm band overlaps to some degree the water ice absorption band. Poleward of 80°N the 20-μm temperature is close to the expected temperatures of 148 K for CO$_2$ ice at 81 mbar or 155 K for a CO$_2$·6H$_2$O clathrate [Kieffer et al., 1976]. However, $T_{11}$ is still warmer than $T_{20}$ in this region, which requires that either the surface kinetic temperature be at the $T_{11}$ value (161 K) or else there is still a warm contribution to $T_{11}$ from the cloud.

Thus the spectral differences observed by the IRTM indicate the presence of a cold cloud over Olympus Mons at 20°N and a second cold cloud which first occurs at 63°N, perhaps even at 61°N if the decrease in $T_{11}$ vs. $T_{20}$ is not due to emissivity effects correlated with surface albedo. However, the $T_{20}$ vs. $T_{20}$ difference is also sensitive to surface emissivity effects; since it decreases at 61°N, the initial $T_{11}$ vs. $T_{20}$ decrease is unlikely to be due to the cloud. As determined from $T_{20}$ vs. $T_{20}$ observations, the most southerly occurrence of surface frost associated with the retreating cap edge also occurs at 63°N. The cloud overlies the transition to fully ice-covered ground poleward of 75°N and then either dissipates or cools to the surface temperature. Near 65°N the cloud temperature is estimated as between 200 and 210 K.

**Cloud Opacity**

The opacity of the polar ice cloud can be estimated from the $T_{11}$ vs. $T_{20}$ difference if the optical properties and the temperatures of the cloud and ground are known. The measurements made by the IRTM are not sufficient to retrieve all these properties, and there are, for the most part, no independent, nearly simultaneous data which can be used to constrain fur-

<table>
<thead>
<tr>
<th>Ice Cover, %</th>
<th>Ice Emissivity</th>
<th>$T_{20}$</th>
<th>$T_{11}$ vs. $T_{20}$</th>
<th>$T_{20}$ vs. $T_{20}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>1.0</td>
<td>210.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>20</td>
<td>0.9</td>
<td>210.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>40</td>
<td>0.8</td>
<td>210.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>60</td>
<td>0.8</td>
<td>210.0</td>
<td>0.0</td>
<td>0.0</td>
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<td>0.2</td>
<td>210.0</td>
<td>0.0</td>
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<tr>
<td>200</td>
<td>0.1</td>
<td>210.0</td>
<td>0.0</td>
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<td>220</td>
<td>0.0</td>
<td>210.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
</tbody>
</table>

**TABLE 2. Brightness Temperatures for Fields of View Covering Mixtures of CO$_2$ Ice at 150 K and Soil at 210 K**
ther the IRTM observations. Thus we proceed as follows. A δ-Eddington radiation code is used to compute the radiation emerging at a given angle from a plane-parallel, isothermal cloud above a surface of unit emissivity. The emergent radiation computed as a function of wavelength is convolved with the IRTM band shape functions [Kieffer et al., 1977] (Figure 1) in order to obtain the IRTM brightness temperatures and their spectral differences. The required optical parameters, namely, single scattering albedo, phase function asymmetry parameter, and extinction efficiency, have been computed as a function of wavelength using a Mie scattering program as developed by Hansen and Travis [1974]. The refractive indices for water ice (measured at 266 K) were taken from Schauf and Williams [1973]. For the most part, the radiative transfer model uses the cloud C.3 particle size distribution of Deirmendjian [1969] with a mode radius of 2.0 μm, since this gave the best fit for the Mariner 9 IRIS observations of the water ice clouds above Tharsis [Curran et al., 1973]. Figures 4a and 4b show the $T_{11}-T_{20}$ differences for various cloud and ground temperatures and as a function of the normal incident extinction optical depth at visible wavelengths, assuming a visible extinction efficiency $\sim 2.2$. The $\sim 10$ K difference for the Olympus Mons cloud indicates an optical depth of unity or greater (Figure 4a), while the $\sim 4$ K (negative) difference characteristic of the polar cloud indicates a visible opacity in the range 0.5–1.0 (Figure 4b). (These are opacities at visible wavelengths; the opacities in the 11-μm band are a factor of 2 smaller.) The sensitivity of the $T_{11}-T_{20}$ differences to our assumed particle size distribution can be judged by reference to Figure 4 where the dashed lines have been computed for the same functional form for the particle size distribution but with a mode radius of 4.0 μm.

The inference of cloud opacity and temperature poleward of 66°N is complicated by the surficial contribution to the $T_{11}-T_{20}$ difference due to the presence of partially frosted ground. In terms of the δ-Eddington model, this anisothermal ground produces a different effective surface temperature for each IRTM band. Figure 5 shows the different families of cloud and ground temperatures which reproduce the observed $T_{6}$, $T_{11}$, and $T_{20}$ values at 64°N. The different frames (Figures 5a–5c) were computed for three different opacities. If there were no surficial contributions to the spectral contrasts, then $T_{6}$, $T_{11}$, and $T_{20}$ should all yield the same ground as well as cloud temperature. Such a solution would be represented in Figure 5 by the simultaneous intersection of all three curves. The absence of such a solution indicates that there is some surface contribution to the spectral contrasts or that the cloud is not isothermal. If it is the former case and the surface contribution is simply due to a mixture of cold and warm patches of surface in the same IRTM field of view, the effective ground temperatures implied by the observed brightness temperatures may differ from one another, but they must do so in a specific way. At the very least, $T_{20}^{s} \leq T_{11}^{s} \leq T_{6}^{s}$, where the superscript indicates an effective surface temperature. The horizontal bar in each frame denotes a possible solution for the given opacity. The intersection of the bar with the three curves yields $T_{20}^{s}$, $T_{11}^{s}$, and $T_{6}^{s}$ values which are consistent with viewing a surface composed of two components, each of unit emissivity and one of which is assumed to be CO$_2$ frost at 150 K. When a similar diagram is constructed for 65°N, we find that a water ice haze having $T_{c} = 212$ K and $\tau_{c} = 1.0$ explains both sets of observed brightness temperatures. Table 3 indicates the cloud temperature, the temperature of the warm ground component, and the fraction of frost-covered surface computed for a cloud opacity of $\tau_{c} = 1.0$ and for latitudes 64°, 65°, 68° and 75°N. These can be regarded as refinements of values derived above, but one should remember the simplifying assumptions of the radiative model and the lack of independent information about the temperature and size of the ice particles forming the haze. The fractional frost cover-
TABLE 3. Model Haze and Surface Characteristics Which Reproduce the Observed Brightness Temperatures on Orbit 1243 Assuming \( \tau_e = 1.0 \)

<table>
<thead>
<tr>
<th>Latitude</th>
<th>( T_{45} ) K</th>
<th>( T_{45}^{11} ) K</th>
<th>( T_{45}^{20} ) K</th>
<th>( T_{45} ) K</th>
<th>( f_{sw}(\tau_e - 0) )</th>
<th>( T_{45}^{11} - T_{20} ) °K</th>
<th>( T_{45}^{11} - T_{20} ) °K</th>
</tr>
</thead>
<tbody>
<tr>
<td>64°N</td>
<td>232.5</td>
<td>224</td>
<td>197</td>
<td>174</td>
<td>0.085</td>
<td>0.5</td>
<td>0.8</td>
</tr>
<tr>
<td>65°N</td>
<td>227</td>
<td>220</td>
<td>194</td>
<td>173</td>
<td>0.05</td>
<td>0.09</td>
<td>0.45</td>
</tr>
<tr>
<td>68°N</td>
<td>231</td>
<td>222</td>
<td>188</td>
<td>163</td>
<td>-4</td>
<td>-2</td>
<td>6</td>
</tr>
<tr>
<td>75°N</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

\( T_{45}, T_{45}^{11}, T_{45}^{20} \) are observed. \( T_e \) is the temperature of the warm (unfrosted) surface component, and \( f_{sw}(\tau_e - 0) \) is the fraction of the surface area covered by CO\(_2\) frost at 150 K. \( T_{45} \) denotes the cloud temperature. \( T_{45}^{11} - T_{20} \) is the computed spectral difference due only to the anisothermal surface. An emission angle of 50° was used in the Eddington calculations.

age computed from the observed \( T_{45} - T_{20} \) values and assuming no haze is also shown for comparison. Table 3 compares the observed \( T_{45}^{11} - T_{20} \) difference with the contrast inferred for the surface alone. The negative \( T_{45}^{11} - T_{20} \) values are clearly due to the relatively cool water ice haze, but the transition to positive \( T_{45}^{11} - T_{20} \) values poleward of 67°N is largely due to the increasing fractional coverage of surface ice. There is still a significant haze contribution at 75°N, however. The opacity of the haze at 75°N is not well constrained by the IR 3.0 observations because of the large surface contribution; the opacity is also poorly determined near 68°N where the cloud temperature may not differ greatly from the effective surface temperatures for each band. However, if \( \tau_e < 0.4 \), the cloud temperature would be greater than 185 K at 75°N, whereas the temperature of a haze of opacity \( \tau_e \approx 1 \) would be less than 180 K; this latter value is more consistent with the radio occultation data from the previous spring [Lindal et al., 1979]. If the cloud opacity at 64°N is less than 0.4, the cloud temperature must be less than 180 K. As noted earlier, requiring the cloud opacity and temperature to be nearly the same at 64° and 65° suggests a water ice haze there having \( \tau_e = 1, T_e = 212 \) K. A dust haze present with the same thermal contrast and a visible opacity of 0.4 would produce a positive \( T_{45}^{11} - T_{20} \) spectral difference of 0.5–1 K for typical dust analogues, none of which could reasonably reproduce the large negative \( T_{45}^{11} - T_{20} \) values shown here.

For the assumed form of the particle size distribution, the column water amount \( M_e \) (in precipitable micrometers (pr \( \mu \)m)) is related to the visual opacity \( \tau_e \) by

\[
M_e = 0.65 \ r_m \ \tau_e
\]

where \( r_m \) is the mode radius of the distribution (in micrometers). Both the Olympus Mons cloud and the polar cloud appear to have the water equivalent of about 1 pr \( \mu \)m. This estimate is uncertain by at least a factor of 2.

The opacity of the Olympus Mons cloud is even less certain than that of the high-latitude cloud because the cloud temperature is not well constrained, and its contrast with the warm surface may be very large. Unlike the Mariner 9 IRIS, the Viking IR 3.0 cannot independently infer the temperature of low-lying clouds. Curran et al. [1973] estimated \( \tau_e \approx 0.4 \) for the Tharsis cloud viewed by the Mariner 9 IRIS but noted that only part of the IRIS footprint (i.e., the field of view of the instrument as projected onto the planet) was covered by clouds, as viewed by the Mariner 9 television cameras. They estimated that the opacity of the visually brightest regions could be as much as 4 times brighter. The footprint diameter of the Viking IR 3.0 channels for the \( T_{45}^{11} - T_{20} \) data shown here is 80–120 km. The fact that there is a range of \( T_{45}^{11} - T_{20} \) values for the Olympus Mons cloud indicates that some of the IR 3.0 spectral fields of view are partially or completely clear and that the clouds have elements spaced up to 80 km apart. Viking visual imaging of clouds above the Tharsis Plateau during the early afternoon and in early northeastern summer showed considerable variability and streakiness, even on spatial scales of 10 km [Briggs et al., 1977].

The absence of significant variations in the \( T_{45}^{11} - T_{20} \) spectral differences for the haze at high latitudes indicates that the characteristic spacing of the cloud elements in the region sampled by the IR 3.0 is much less than the 100-km FOV. High-resolution Mariner 9 television observations of the vicinity of the north polar cap during spring (1972) revealed faint clouds organized in characteristic wave cloud patterns [Leovy et al., 1973]. If the haze observed by Viking has a similar structure, the spacing between the wave clouds must be considerably less than 100 km.

Water Vapor

The latitudinal profile of the total amount of water vapor in a vertical atmospheric column as measured by MAWD is shown in Figure 3d for orbit 1243. The MAWD observations have been averaged together in 5° latitudinal bins in order to improve the signal-to-noise ratios for observing high latitudes where the water vapor abundances are low during early spring [Jakosky and Farmer, 1982]. This averaging is also consistent with the projected length of the MAWD instantaneous field of view (350–450 km) obtainable at these solar slant angles. Formal errors in the computed water vapor abundances are estimated to be less than 10%. The relative minimum in column water vapor amount at 15°–20°N coincides with the water ice cloud over Olympus Mons. Although the water equivalent of the cloud may be comparable to the observed deficit of \( \sim 5 \) pr \( \mu \)m, the observed minimum is more likely to be due to the reflection of sunlight at the cloud top masking the water vapor in and beneath the cloud from the MAWD instrument. Indeed, there is a slight brightening detected by the MAWD continuum channels which suggests this is the case. If water vapor tended to be uniformly mixed, there would be a relative minimum in the column vapor abundance above the high terrain of the shield volcanoes even if the cloud were absent. Jakosky and Farmer [1982] found that the column abundances above the Tharsis Plateau were low relative to the zonal average even when appropriately normalized by the column airmass. Masking of the water vapor by orographic water ice clouds could account for the observed deficit.

There may be a masking effect at higher latitudes as well because the observed decrease in polar water vapor begins when the IR 3.0 data suggest that the polar haze begins and is associated with a sharp increase in the MAWD continuum radiance, although this increased brightness could also be produced by surface CO\(_2\) or water ice. An isothermal atmospheric layer 10 km thick and at 190 K can hold the vapor equivalent of nearly 4 pr \( \mu \)m, while the same layer can hold less than 1 pr \( \mu \)m of water vapor at \( T_e = 180 \) K and less than 0.1 pr \( \mu \)m at 170 K, according to computed saturation water vapor pres-
Fig. 6. North polar cap regression observations. Polar cap edge determined from IRTM T_{16}-T_{20} observations are shown as open squares. IRTM observations were acquired during the final phase of the Viking orbiter mission (August 1979 to April 1980). Also shown are Viking orbiter imaging observations [James, 1982] and earth-based visual observations [Iwasaki et al., 1982] for the same season. Note the slowdown in cap regression between L_{s} = 20° and 50°.

The Survey/Completion Mission Data

The locations of water ice atmospheric hazes and of surface CO_{2} ice have been determined for the remaining Viking Survey/Completion mission latitudinal infrared scans by searching for the first occurrences of negative T_{16}−T_{20} values and of rapidly increasing positive T_{0}−T_{20} differences, as suggested by the analysis of the orbit 1243 data. The data from these scans are summarized in Table 1. This series of north-south scans spanned L_{s} ~ 348°−86°, covering the period of north polar cap retreat. Data taken at the very end of the completion mission (L_{s} > 86°) did not extend far enough north to intersect the polar cap edge.

Seasonal Variation of Cloud and Polar Cap Locations

The locations of the CO_{2} north polar cap edge determined from the T_{0}−T_{20} spectral differences observed during L_{s} ~ 348°−86° are presented in Figure 6. These are compared to the regression curves determined for the cap edge by Iwasaki et al. [1982] from earth-based observations and by James [1982] from Viking visual photographs for the same period. The cap edge determined from the IRTM data will generally be equatorward of the visual cap edge because the T_{0}−T_{20} data have been used to locate the first occurrence of CO_{2} ice on the surface as one goes poleward. Given this operational difference, the uncertainties in measuring the cap edge from visual observations, and the limited range of longitudes observed by the IRTM, the computed locations of the cap edge as a function of L_{s} agree quite well.

The slowdown in cap regression observed from earth between L_{s} = 20° and 58° [Iwasaki et al., 1982] is also observed in the IRTM data. Although this standstill is not seen in the Viking visual observations for this year, it was observed in the previous Martian year (1977−1978 [James, 1979]) and in terrestrial observations taken in 1962−1968 [Capen and Capen, 1970] and 1977−1978 [Iwasaki et al., 1979]. From visual observations there is some uncertainty as to whether this slowdown truly represented the cap behavior or was an artifact produced by atmospheric hazes [James, 1979]. The IRTM data, which can separate the surface and haze contributions, confirm that the cap did cease to retreat at some time between L_{s} = 1° and 25° and remained nearly constant until L_{s} = 49°. These data, together with the apparent slowdown seen in previous years, indicate that such behavior is typical for the north polar cap regression.

The general agreement between the data sets indicates that the polar haze did not significantly affect the estimates of the cap location, probably because the polar cloud often did not extend equatorward of the cap edge. However, there are important differences in the timing of changes which occur in regression rates. These differences may be due to clouds obscuring the visual view of the cap edge and masking the retreat of the surface ice. For example, the IRTM data indicate that cap regression resumes at L_{s} = 48°, while terrestrial observations place this point at L_{s} = 58°. During this period, clouds extend as much as 2° in latitude south of the cap edge (see Figure 7), a point very close to the visually determined edge. It therefore appears that the IRTM data can provide a better estimate of the detailed polar cap regression because of the higher spatial resolution when compared to terrestrial observations and the ability to discriminate between surface ice and atmospheric clouds. These data should better constrain dynamic models of polar cap retreat.

In Figure 7 the IRTM-determined cap and cloud locations are given for the observations studied. The cloud location marks the southernmost extent: no attempt was made to estimate possible northern limits to the clouds. As can be seen, the southern edges of the cap and the cloud differ at most by a few degrees of latitude. For two observations (L_{s} = 27° and 56°) clouds were not observed, possibly because of the very
limited longitudinal coverage. The characteristic signature of
the polar clouds was not observed by the IRTM after $L_e = 65^\circ$. The terrestrial observations suggest that the cap con-
tinued to recede up to $L_e = 82^\circ$ at which time it had reached the
permanent cap location at 80°N ([wasaki et al., 1982]). IRTM
data from orbit 1379 suggest that the cap has again stopped
receding by $L_e = 70^\circ$, at a latitude of 73°N. This cessation in
retreat coincides with the lack of clouds and suggests that the
two processes may be coupled. One should note that the $L_e = 86^\circ$
data were taken at the northernmost extent of the obser-
vation scan and may not be as reliable as the earlier data.
Data beyond $L_e = 86^\circ$ are not available for this year.

Source for the Polar Clouds

As noted above, the Viking radio occultation data reveal
the presence of a temperature inversion over the CO$_2$ ice cap;
numerical simulations of the atmospheric climate in the vicin-
ity of the receding CO$_2$ seasonal ice cap indicate that a warm
atmospheric layer below 10 km but above the planetary
boundary layer (i.e., above ~2 km) may extend to and per-
haps equatorward of the CO$_2$ seasonal cap edge [Haberie et
al., 1979]. The cloud temperatures derived here indicate that
the water ice haze is likely to be in this warm atmospheric
layer. If so, the most likely source of the water ice haze is
advection of water vapor from the warmer latitudes equator-
ward of the CO$_2$ frost-covered ground. The 4–5 pr $\mu$m
of water vapor observed there by MAWD are sufficient to supply
the 1–2 pr $\mu$m forming the cloud. The source of this atmo-
spheric vapor is not known. The necessary vapor could be
supplied locally by water ice subliming or vapor desorbing
from the largely CO$_2$ frost-free ground. As the surface warms,
the CO$_2$ frost should disappear first, followed by the subli-
mation of water frost. As the ground continues to warm, some
of the water adsorbed just beneath the surface will also be
driven into the atmosphere. The amount of water ice required
to form the observed cloud is only 1–2 pr $\mu$m. Since the sur-
ficial roughness is probably considerably greater than that
deptch, such a thin layer of water frost would not be easily seen
from orbit. Indeed, very thin water frosts could extend visually
undetected to the south of the CO$_2$ cap edge; a patchwork of
surface (water) frost and bare ground is also unlikely to pro-
duce detectable IRTM spectral differences because of the rela-
tively small contrast between the temperatures of the water
frost and the bare ground. (The strong signature of CO$_2$ frost
patches next to bare ground is produced by the large thermal
contrast due to the very low CO$_2$ sublimation temperature.)
Also, the surface emissivities for a water ice deposit are likely
to be near unity [Warren, 1982].

As one moves into the cap interior, it is unlikely that the
surface can provide even a few tenths of a precipitable mi-
crometer of water vapor, and even that vapor may be confined
to the lowest layers of a very stably stratified polar atmos-
phere. Thus the polar cloud should either thin out over the
CO$_2$ cap interior or the condensate must be advected in from
lower latitudes.

Comparison With Earlier Viking Data

The approach used here for the analysis of the Viking Sur-
vey/Completion mission infrared data is currently being ap-
plied to the more abundant data taken during earlier phases
of the Viking mission. The IRTM data covering the previous
Martian northern winter and spring are of particular interest
since those data have much better temporal resolution and
longitudinal coverage than the Viking Survey/completion mis-
sion data used here.

Kieffer [1979] has previously described the IRTM spectral
contrasts observed by Viking for high southern latitudes
during the retreat of the south polar cap. These contrasts are
larger and more complicated than those presented here, for
the north polar cap retreat, primarily because of effects due to
the atmospheric dust raised by local and planetary-scale dust
storms. Prior to the onset of the great dust storms (at $L_e \approx
205^\circ$ and 279°), the observed correlation of high albedo and
warm $T_{20}$ values suggested the presence of either water frost
or water-ice clouds over the southern seasonal CO$_2$ caps
[Kieffer, 1979], and there was an apparent release of a few
precipitable micrometers of water vapor as the ice cap re-
treated [Davies and Wainio, 1981]. However, far less water
vapor appeared in the southern hemisphere than in the north
during its corresponding season [Davies and Wainio, 1981;
Jakosky and Farmer, 1982]. Kieffer [1979] found large positive values of $T_{11}-T_{20}$ in the regions where $T_{20}$ values suggested that CO$_2$ ice was present on the surface. These $T_{11}-T_{20}$ differences were closely correlated with the apparent position of the residual CO$_2$ south polar cap and thus were attributed largely to nonunit emissivity of CO$_2$ frost at 20 $\mu$m rather than to the presence of a water ice haze in the relatively warm atmosphere. The $T_{11}-T_{20}$ positive values found over the seasonal CO$_2$ ice cap in the data analyzed here (Figure 3) also suggest a surface emissivity less than unity at 20 $\mu$m for CO$_2$ ice, but, as noted earlier, the possible presence of a water ice cloud and/or of airborne dust precludes a reliable estimate of the emissivity itself.

**Summary**

Viking infrared observations of the Martian north polar regions taken during the 1979–1980 final phase of the Viking orbiter mission have been used to locate, to determine the composition, and to estimate the temperature and opacity of hazes above the edge of the retreating seasonal cap for the period from late winter to midsummer ($L_s = 240^\circ–86^\circ$). These hazes, composed of water ice, are remarkably uniform and are present for most of the period of observations. They extend from at most a few degrees in latitude south of the cap edge, overlie the transition from frost-free to fully frost-covered ground, and extend an unknown distance into the cap interior. The cloud temperatures appear to vary from ∽210 K near the cap edge to values below 180 K over the cap interior. Comparison of these estimated temperatures with Viking radio occultation profiles from the previous spring and with simultaneous IRTM $T_{13}$ values suggests that the polar haze is 3–8 km above the surface. The visible opacities of the clouds are approximately unity, equivalent to 1–2 pr $\mu$m of water vapor. The polar hazes may provide a temporary reservoir for water which appears as vapor during midsummer just equatorward of the edge of the seasonal CO$_2$ ice cap, although this is by no means certain based on the available data.

The edge of the polar cap, defined as the most equatorward occurrence of CO$_2$ ice on the surface, can be distinguished in the IRTM data from the edge of the haze above it. IRTM data taken during the 1979–1980 Martian spring yield a regression curve for the cap edge which generally agrees with results produced by earth-based and spacecraft visual imaging. However, there are differences in the details of cap regression. For instance, the cap edge as determined from the IRTM data does extend south of 60$^\circ$N during late winter, and the cap retreat resumes in midsummer at $L_s = 50^\circ$.

Because the negative $T_{11}-T_{20}$ spectral difference is such a clear-cut indicator of water ice clouds over a warm surface, the IRTM data could be used to provide an atlas of cloud locations for comparison with MAWD observations, with visual imaging data or with data from previous Martian years. Such a survey is now underway using the more abundant data from the earlier Viking orbiter missions. Unfortunately, the opacities of such water ice clouds will be difficult to determine because of the lack of independent temperature information for the lower atmosphere. Constraints on the cloud temperature, and thus opacity, are best obtained for the water ice hazes near the seasonal cap edge with their characteristic signature of negative $T_{11}-T_{20}$ values giving way to positive values.

The IRTM data should also provide more accurate observations of the polar cap retreat for the previous Martian springs observed by Viking by utilizing the higher spatial resolution of the IRTM and by using different spectral contrasts to distinguish between surface ice and atmospheric clouds.

**Acknowledgments.** We wish to thank D. Paige for his assistance and comments. P. R. Christensen's work was supported by the Planetary Geophysics and Geochemistry Program and represents one phase of research carried out at Arizona State University, supported by National Aeronautics and Space Administration grant NHEW-263. R. Zurek's work was supported by the Planetary Atmospheres Program Office, NASA, and represents one phase of research carried out at the Jet Propulsion Laboratory, California Institute of Technology, under contract NAS7-100, sponsored by the National Aeronautics and Space Administration.

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