The Hemispherical Asymmetry of the
Residual Polar Caps on Mars

Contract No. NASW-4444

Semi-Annual Progress Report

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A model of the polar caps of Mars has been created which allows for light penetration into the cap, allows ice albedo to vary with age, latitude, hemisphere, dust content and solar zenith angle, includes the radiative effects of clouds and dust, allows for diurnal variability, and includes heat transport as represented by a thermal wind. The model reproduces polar cap regression data very well, including the survival of CO$_2$ frost at the south pole and not the north pole, and reproduces the general trend in the Viking Lander pressure data, although further improvement is needed. Papers were presented at six conferences and in the Journal of Geophysical Research, and have been submitted to Icarus. The research plan for the next reporting period involves further publishing and presenting of our results.
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I. Abstract for the project

Computer simulation of the condensation and sublimation of CO$_2$ frost in the martian polar caps has been fairly successful in reproducing the annual cycle in atmospheric pressure observed by the Viking Landers. However, these studies have not been able to uniquely explain why CO$_2$ frost survives southern summer. We propose to study several mechanisms which might explain this phenomenon: an improved treatment of cloud and atmospheric-dust effects; the decrease in the importance of heat conduction due to existing CO$_2$ ice; an improved treatment of sublimation microphysics; the variance in ice albedo with solar zenith angle, ice age, and dust/ice ratio; the bidirectional reflectance of ice; snowfall accumulation in addition to surface frost formation; insulating residues on the ice; wind shifting of ice and brightening of ice under exposure to sunlight. To perform this study, we will integrate two existing models of the martian polar caps and combine them with an accurate model of martian atmospheric radiation and parameterizations of the processes described above. The model will be used to identify combinations of parameters which will allow a reproduction of seasonal observations of atmospheric pressure and the latitudinal and seasonal extent of the polar cap, and the hemispherical asymmetry in the residual polar caps. A better understanding of the martian polar caps will have direct implications for our understanding of the climate and atmospheric dynamics of Mars and the Earth, as well as provide insight for the observations planned for the Mars Observer mission.
II. Summary of Research to date

II.1. Jakosky Model Acquired

The Principal Investigator, Dr. Lindner, flew to Colorado in September 1989 to meet with the Co-Investigator, Dr. Jakosky. The polar cap model of Dr. Jakosky was obtained. Dr. Lindner became familiar with the code in discussions with Dr. Jakosky, and through exercising the model on Dr. Jakosky's computer system. Dr. Lindner and Dr. Jakosky also held in-depth discussions of the overall approach to be taken on the project to meet the objectives.

II.2. Lindner and Jakosky Models Integrated

The Jakosky model was loaded onto the AER computer system and combined with the Lindner model, as discussed in the proposal. The hybrid model was debugged of errors.

II.3. Comparison to Kieffer's Surface Model

Kieffer (1990) described a model which is similar to ours, which he used for ice metamorphism studies. I set up the code to use the same inputs as Kieffer did in his Figure 3, and plotted the results on the same figure (deleting all but the case I compared to). The agreement is good, as can be seen in Figure 1. Since I don't know exactly what all of his input parameters were (such as emissivity, seasonal ice albedo, etc.), I consider the agreement a validation of our code. The only different behavior is why his figure doesn't show a seasonal temperature "pulse" which makes it's way down in depth (i.e.
Fig. 1. Temperatures within the north polar ice cap (shown are the annual minimum and maximum). Lindner and Jakosky values are for the current study ($L_s = 260$ and 280 degrees), and Kieffer values are taken from Kieffer (1990). The albedo of the residual ice is 0.41, and the thermal inertia is 0.04 cal cm$^{-2}$ sec$^{-0.5}$ K$^{-1}$. Surface values are shown as 1 cm depth.
when the surface is at a maximum, the temperature at 10 m should be at a relative minimum, as our results show). Perhaps Kieffer is showing the maximum temperature for all year at each depth, rather than a snapshot in time when the surface reaches a maximum (as I've plotted).

II.4. Ice Microphysics Parameterization

Section 2.3 of the proposal discussed the inclusion of improved ice microphysics in the model. Discussions were held with Dr. Gary Clow as to the nature of his ice microphysics models and the results he has obtained recently. In the proposal, we mentioned that the depth to which radiation penetrates into the cap has never been included in polar cap sublimation models, although it could be important. We have done calculations that show that this is not important for thick CO$_2$ ice, or thin CO$_2$ ice overlying a dust surface. For the thick ice case, the integral of sublimation is independent of where it occurs within the ice. The distribution of absorption within the polar cap is important for studies of the internal physics of the cap, but not for overall sublimation rate. For the thin CO$_2$ ice over dust case, the surface thermally reradiates the absorbed solar radiation primarily back to the polar cap where most is reabsorbed. The top few grains of dirt absorb the incoming solar radiation which penetrates the polar cap, and it is much easier for these top few grains to conduct or radiate heat back to the overlying polar cap than deep into the surface.
A thin CO₂ ice polar cap overlying a residual water ice deposit is different. Solar radiation which passes through the CO₂ ice cap may penetrate quite deeply into the H₂O ice. This will effect CO₂ ice lifetimes over the residual polar cap, but won’t affect overall polar-cap recession or the annual variation in surface pressure, both of which current models reproduce. I incorporated this effect in the model by apportioning the absorbed solar flux between the sublimation of the ice, and the heating of the subsurface. Clow (1987) has already calculated the absorption of solar flux as a function of depth in CO₂ ice on Mars, and I have reproduced a figure from his paper in Figure 2, which we used in the model. Based on Clow’s figure, I obtained a % of solar flux absorption as a function of depth for 41 depth steps. The code already calculates the thickness of CO₂ ice, and I use this depth with Clow’s work to determine how much sunlight penetrates the seasonal ice. This fraction that penetrates the seasonal ice is further apportioned between the individual subsurface layers, again using Clow’s figure.

Both Kieffer (1990) and Moore (1988) have deduced that ice grain radii in the residual cap are over 100 μm and that dust concentrations are less than 1/1000, so any one of Clow’s profiles could be legitimate. I’ve tried them all, and found the largest deviation was for clean, fine ice. This is understandable, since solar radiation penetrates the deepest in clean, fine ice. In our work, we have assumed that the seasonal ice was fine, and the residual ice was coarse.
FIG. 2. The integral of net shortwave flux with respect to depth for clean and dusty snow, using the mean-annual incident solar flux at latitude $-38^\circ$. Mean-annual temperature profiles in snow are proportional to the flux integral $I_s(z)$, which is generally greater for coarse-grained than for fine-grained snows. Curves (A) correspond to a snow with ice grain radii of 50 $\mu$m and bulk density 50 kg m$^{-3}$ while curves (B) are for a snow with 1000-$\mu$m ice grains and a density of 400 kg m$^{-3}$. [from Clow (1987)]
I have plotted in Figure 3 the difference in the depth profile of temperature between the code with no penetration of light and the modified code using penetration appropriate for clean ice for $L_S = 260^\circ$ (just before the last of the seasonal ice sublimes). The radiation which penetrates the cap heats the subsurface by up to 3 K. Most of the radiation gets absorbed near the surface, but that easily conducts away to the surface. The greatest heating occurs at 1 m depth. The scale on the right side of the figure is for the "difference" curve. The surface temperature is plotted as 1 cm depth on this plot.

Figure 4 shows the profiles for $L_S=280^\circ$, when the maximum surface temperature occurs, as well as the maximum deviation at depth between the codes (9 K). The surface is actually a couple of degrees cooler for the penetrating radiation case, as we expected. Again, the maximum subsurface heating occurs at 1 m depth.

Figure 5 shows how the surface temperature varies with $L_S$ for both codes. Surface temperatures are cooler from $L_S = 270^\circ$ to $L_S = 310^\circ$ for the penetration code, since solar radiation is absorbed at depth, and must conduct to the surface. From $L_S = 310^\circ$ to 360°, the surface is actually warmer than for the original code, due to the increased conduction by the warmer subsurface.

Figure 6 shows depth profiles of temperature for every 10 sols through the year for clean ice penetration, and Figure 7 shows the same for the no-penetration code. Again, the heating of the subsurface can be seen for summer (just before the
Fig. 3. Temperature versus depth in the south polar cap for $L_s$ of 260 degrees simulated including the effect of light penetration for "clean" ice, and simulated without penetration. The difference between with and without is also plotted, using the scale on the right side. Surface values are shown at 1 cm depth.
Fig. 4. Temperature versus depth in the south polar cap for Ls of 280 degrees simulated including the effect of light penetration for "clean" ice, and simulated without penetration. The difference between with and without is also plotted, using the scale on the right side. Surface values are shown at 1 cm depth.
Fig. 5. Surface temperature of the south pole versus Solar Longitude simulated including the effect of light penetration for "clean" ice, and simulated without penetration. The difference between with and without is also plotted, using the scale on the right side.
Fig. 6. Temperature versus depth in the south polar cap for every 10 sols through one year simulated including the effect of light penetration for "clean" ice. Surface values are shown at 1 cm depth.
Fig. 7. Temperature versus depth in the south polar cap for every 10 sols through one year simulated without the effect of light penetration. Surface values are shown at 1 cm depth.
surface ice sublimes away) and in the warm subsurface at the maximum temperatures. We also see that very little annual variation in temperature occurs at 4 m depth.

Figures 8 and 9 show the annual variation in temperature, with each curve representing a different depth (at \( z=0, 2.8, 9.5, 19.0, 32.2, 50.6, 76.5, 112.7, 163.4, 234.4, 333.7, 472.8, 667.5, 940.1, 1321.8, 1856.2, 2604.2, \) and \( 3651.5 \) cm). The deepest depths show the least variation (almost a straight line at 36 m. The ice-covered heating of the subsurface can be seen between \( L_S = 180^\circ \) and \( 270^\circ \).

Figure 10 shows the effect of light penetration on surface frost between the two models in the first year of iteration (starting from the same initial conditions of an isothermal surface of 154K, and no frost). Since the sun doesn't shine from \( L_S=0^\circ \) to \( 180^\circ \), no difference is seen then. As the cap thins in the spring, the effect of sunlight penetrating the cap becomes more important (as expected from Clow's figure). Thus the ice in the model which includes penetration lasts longer in Year 1.

Figure 11 shows the same for Year 6, by which time the models have converged. Now the model which includes penetration has less ice all year. This is because the subsurface heats up more in the summer due to light penetration, and this heat decreases the amount of condensing frost in the fall and winter, resulting in an increasing difference between the models from \( L_S=0 \) to \( 180^\circ \). After \( L_S=180^\circ \), this difference decreases because of light penetration as we saw for Year 1, but not enough to
Fig. 8. Surface temperature of the south pole versus Solar Longitude, simulated including the effect of light penetration for "clean" ice. Each curve represents another layer in the model, ranging from the surface to 3.5m depth.
Fig. 9. Surface temperature of the south pole versus Solar Longitude, simulated without the effect of light penetration for "clean" ice. Each curve represents another layer in the model, ranging from the surface to 3.5m depth.
Fig. 10. Frost amount at the south pole versus solar longitude, simulated including the effect of light penetration for "clean" ice, and simulated without penetration. The difference between with and without is also plotted, using the scale on the right side. Results are for the first year of model iteration, starting from an isothermal 154K surface and 10g cm⁻³ of frost.
Fig. 11. Frost amount at the south pole versus solar longitude, simulated including the effect of light penetration for "clean" ice, and simulated without penetration. The difference between with and without is also plotted, using the scale on the right side. Results are for the sixth year of model iteration.
compensate completely, and the ice sublimes away earlier for the model which includes penetration.

Figures 12 and 13 show the depth profiles of temperature for the dirty ice case versus the no-penetration model. The differences in temperature exhibit the same behavior as with clean ice, but of a much lower magnitude (as we expected). Figure 14 shows surface temperature as a function of $L_S$ for dirty ice and the original code, again exhibiting the same behavior as for clean ice, but with a smaller magnitude. Figure 15 shows the effect of light penetration on the frost budget for dirty ice, again showing the same behavior as for clean ice, but with a smaller magnitude.

I have checked to make sure that these deviations in temperature are not caused by the modifications to the code. To check that these changes weren't altering the code in its original form, I ran one case where all solar radiation is absorbed in the top mm of ice, rather than distributed over the top meter or so. This case gave the same results as Jakosky's original code, and Kieffer's model.

Obviously, the code had to be run with very thin layers at the surface (a 5 cm surface layer thickness for this case). Thicker layering caused an error, which was determined by comparing to the original Jakosky code. Even a 26 cm top layer caused a 3 sol increase in the number of ice-free days, and a 100 cm top layer caused a 15 sol error. This is obvious, since the layering needs to be on the same scale as the penetration depth.
Fig. 12. Temperature versus depth in the south polar cap for Ls of 260 degrees simulated including the effect of light penetration, and simulated without the effect, for "dirty" ice. The difference between with and without is also plotted, using the scale on the right side. Surface values are shown at 1 cm depth.
Fig. 13. Temperature versus depth in the south polar cap for Ls of 280 degrees simulated including the effect of light penetration for "dirty" ice, and simulated without penetration. The difference between with and without is also plotted, using the scale on the right side. Surface values are shown at 1 cm depth.
Fig. 14. Surface temperature of the south pole versus solar longitude, simulated including the effect of light penetration for "dirty" ice, and simulated without penetration. The difference between with and without is also plotted, using the scale on the right side.
Fig. 15. Frost amount at the south pole versus solar longitude, simulated including the effect of light penetration for "dirty" ice, and simulated without the effect. The difference between with and without is also plotted, using the scale on the right side. Results are for the sixth year of model iteration.
The results do depend on the input parameters (albedo of ice, inertia, albedo of surface, emissivity, etc.). For example, a low thermal inertia (0.01) yields a much hotter subsurface and dramatic reduction in ice-covered sols due to light penetration. Also, the albedo needed to keep ice year-round is ~0.04 higher for frost emissivity of 0.95 versus 1.0 (obviously, since a reduction in emissivity reduces amount of heat lost, and a higher albedo reduces the absorbed solar radiation to maintain balance). The results do not depend strongly on which hemisphere is studied. However, the results do depend on the type of ice (coarse/fine and clean/dirty). A dirty, coarse ice has much less light penetration than a clean, fine ice. But all cases result in fairly small changes, and can be neglected given the uncertainties in other parameters.

I did run one case with soil as the underlying surface. As we had stated earlier, there is no difference between the no-penetration code and a code which allows light to penetrate the seasonal cap and get absorbed in the top mm of soil. Almost all the penetrating radiation returns to sublime the seasonal ice from this depth.

However, this does imply that the seasonal CO₂ frost sublimes away more quickly over the residual cap in the spring than over the neighboring soil, and condenses more slowly over the residual cap in the fall than over the surrounding soil. This might tend to move the residual cap around a bit on a climatic timescale, but would probably be a small effect.
II.5. Solar zenith angle dependence of albedo.

The albedo of ice depends on solar zenith angle (SZA). This SZA dependence has been theoretically predicted for CO₂ frost on Mars by Warren et al. (1990). I assumed that the co-albedo (1-albedo) varied with the SZA dependence given by Warren, no matter what the albedo was. This will give reasonable albedos when near the 0.7 base albedo used in Warren’s calculations, but is speculative for albedos far from this value. Fortunately, the values considered here are very close to this value.

As mentioned by Warren et al. (1990), his calculations are for a flat surface. Because of surface roughness on a natural snow surface, the effective zenith angle is usually much less, and rarely over 80° (Warren et al., 1990). I have simply adjusted the SZA to be 10° less than given by orbital mechanics as a simple approximation for surface roughness.

I find that the influence of the sun on ice stability is significantly reduced in spring and fall, when SZA’s are large. This significantly lowers the albedo required to maintain ice on the south pole year-round, by about 5%.

II.6. All parameterizations considered together, with the best fit to the data.

The results discussed up to now dealt strictly with ice at 90°N and 90°S latitude. The problem with polar cap models to date has been that they cannot maintain CO₂ ice on the south pole without also maintaining it on the north pole, unless a
hemispherical asymmetry in ice albedo is assumed. Having noted that the effects of clouds and dust can correct this problem (Lindner, 1990), and having seen that the SZA effect on albedo can allow for reasonable albedos to accomplish this, I set out to run the model at all latitudes and seasons to see if the observed data on polar cap regression and atmospheric pressure could be reproduced.

The Lindner (1985) polar cap model was modified to allow for light penetration into the cap, to allow ice albedo to vary with age, latitude, hemisphere, dust content and solar zenith angle, to include the radiative effects of clouds and dust, to allow for diurnal variability, and to include heat transport as represented by a thermal wind, as discussed in the proposal. Ice albedo was chosen to have either the albedo for old ice or new ice, depending upon whether frost was condensing or subliming, with a gradual transition from the albedo of new ice to that of old ice over a period of 10 days. The albedo of old ice was usually assumed to be less than that of new ice, although several cases of brightening of ice with age of the type postulated by Paige (1985) were investigated. Ice albedo also was allowed to vary linearly with distance from the edge of the polar cap, being up to 10% higher if the edge of the polar cap were 50° in latitude away. The transport of dust into the winter polar region is ineffective and there are no sources of dust when the soil is ice-covered. Changes in albedo due to dust content within the ice and hemispherical asymmetries in albedo were allowed to vary at random to obtain the best fit to the data.
The radiative effects of clouds and background dust (i.e. other than during dust storms) as computed by Lindner (1990) were incorporated. The cloud opacity was assumed to be hemispherically asymmetrical, with an opacity of 0.5 in the north and 0.2 in the south, and to vary linearly with the annual cycle in atmospheric surface pressure observed by the Viking Lander. Clouds were assumed to exist only where CO$_2$ frost existed. The annual cycle in background dust opacity was taken from Pollack et al. (1977). Dust opacity was assumed to vary linearly with distance from the edge of the polar cap, so that it was only half that at the equator if the edge of the polar cap was 50° in latitude away. Both background dust and cloud opacity were also assumed to vary linearly with the changes in airmass due to changes in elevation. The radiative effects of CO$_2$ vapor were also included (NIR absorption of solar radiation and emission of IR radiation). The scientific basis for all of these approximations were discussed in Lindner (1990) and in the proposal to this contract and will not be repeated here.

Heat transport was simply incorporated by the use of the thermal wind approximation. Future versions of this model will use a full GCM calculation. The thermal wind is based on the temperature difference across the edge of the polar cap, and various efficiencies for heat transport were chosen for model comparisons.

The model was also run with time steps of 1/50 of a day, small enough to account for diurnal variations in solar zenith angle, surface temperature, and frost condensation/sublima-
tion. The model was run for every 2° in latitude to account for latitudinal variation in model parameters. An albedo of 0.45, a thermal inertia of 0.03 cal cm\(^{-2}\) s\(^{-1/2}\) K\(^{-1}\) and a surface density of 0.93 g cm\(^{-3}\) were chosen for the residual polar caps (Kieffer, 1990; Jakosky and Haberle, 1990), while values for bare soil of 0.25 for albedo, 0.006 cal cm\(^{-2}\) s\(^{-1/2}\) K\(^{-1}\) for thermal inertia, and 1.5 g cm\(^{-3}\) for surface density were chosen for other latitudes. Frost temperature was allowed to vary with atmospheric pressure.

I ran hundreds of combinations and permutations of model parameters. Rather than present hundreds of graphs, charts, etc., allow me to summarize the results and show the best fit to the data. The penetration of light into the polar cap was not deemed to be important, as discussed in section II.4. The division of albedo between old and new ice was also not very important. The fits to the data were not markedly different whether old ice was chosen to be higher or lower than new ice. The dependence of ice albedo on SZA and the radiative effects of clouds and background dust were only important right at the pole itself, and not for the overall polar cap regression or atmospheric pressure, as postulated by Lindner (1990). The radiative effects of global dust storms were not investigated, and could be important. Running the model including diurnal variability versus in a diurnally-averaged mode also made little difference to the fit to the data. The inclusion of heat transport was significant, and therefore warrants the inclusion of more accurate GCM simulations.
Figures 16 through 19 show the regression of the polar caps as predicted by the model versus observations. Agreement is good, particularly in the northern hemisphere. An exact comparison is difficult, considering that the edge of the polar cap is usually patchy and ill-defined, in large part due to terrain. Further, observations of the polar cap are conducted during the daylight. Near the cap edge, ice frequently forms at night and sublimes away completely during the day. Thus, the edge of the polar cap is also diurnally-variable.

Figures 20 and 21 show the comparison of the atmospheric pressure predicted by the model for the elevations of the Viking Landers to the actual observations. The agreement is good, although the annual cycle in pressure predicted by the model is slightly out of phase with the data. This is a problem noted with all runs of the model, no matter what sets of input parameters were used. The model is still ignoring or poorly treating some process which forms ice sooner in the fall, and sublimes it away more quickly in the spring. Some suggestions for what these may be are presented in the section on plans for future work.

Figures 22 and 23 show model predictions for the amount of frost at particular latitudes throughout the year. These values can be converted to a depth of frost by using a density of 0.93 g cm$^{-3}$. Frost forms as equatorward as 50°S and 54°N latitude with typical maximum depths of a meter. Note that frost forms later and sublimes earlier at lower latitudes. Note also that once ice begins to sublime, it does so rapidly.
Fig. 16. The regression of the south polar cap, as observed for various years (taken from Iwasaki et al., 1990) and as simulated by our model (thin line), as a function of the aerocentric longitude of the sun ($L_s$). The cap radius is that which would be measured on a polar stereographic projection of the south polar region; the units of the radius are fractions of the planetary radius of Mars.
Fig. 17. The regression of the south polar cap, as observed in 1986 (solid circles), 1971 (crosses), and 1977 (plus signs) [taken from James et al., 1990] and as simulated by our model (thin line), as a function of the aerocentric longitude of the sun (Lₚ). Two-sigma error bars are indicated for the 1986 data; the errors are smaller for the denser 1971 data and for the 1977 Viking data.
Fig. 18. The regression of the north polar cap, as observed for various years [taken from Iwasaki et al., 1982; symbols refer to Dollfus (1973), Baum (1974), Briggs (1974), James (1979), Capen and Parker (1981), Iwasaki et al. (1982)] and as simulated by our model (thin line), as a function of the aerocentric longitude of the sun ($L_s$).
Fig. 19. The regression of the north polar cap, as simulated by our model (thin line) and as observed, as a function of the aerocentric longitude of the sun ($L_s$). Data are taken from James et al. (1987); also showing data of Iwasaki et al. (1979; 1982). The symbols stand for 1977-1978 and 1980 Viking data (x's and +'s respectively) and ground-based 1975-1980 data (circles).
Fig. 20. Daily mean surface pressure at Viking Lander 2, showing observations (line with high-frequency oscillations) and simulations (smooth line) over the course of one martian year starting from the vernal equinox. Observations are taken from Hess et al. (1980), and gaps in the data are due to irretrievably lost data.
Fig. 21. Daily mean surface pressure at Viking Lander 1, showing observations (line with high-frequency oscillations) and simulations (smooth line) over the course of one martian year starting from the vernal equinox. Observations are taken from Hess et al. (1980), and gaps in the data are due to irretrievably lost data.
Fig. 22. Mass of CO₂ frost per cm² of surface at every 2 degrees of latitude in the northern hemisphere over one martian year starting at vernal equinox. The solid line is 90°N latitude.
Fig. 23. Mass of CO$_2$ frost per cm$^2$ of surface at every 2 degrees of latitude in the southern hemisphere over one martian year starting at vernal equinox. The solid line is 90°S latitude.
II.7. Presentations made at conferences

Several conferences were attended which were supported in whole or in part by this contract. Reprints of papers and abstracts published at these meetings are included in the Appendix.

1989 DPS Conference

Dr. Lindner attended the AAS/DPS conference in Providence in October, 1989, and presented a paper entitled "The Martian Polar Cap: Radiative effects of ozone, clouds, and airborne dust", which discussed the theoretical aspects of this current work. The work was well received, and Dr. Lindner benefitted from discussions with Dr. Jakosky and others at the conference, and from listening to other talks. The abstract was published in the Bulletin of the American Astronomical Society, Volume 21, p. 979 (1989). NASA contract NASW-4444 paid for expenses.

LPS Conference

Dr. Lindner attended the Lunar and Planetary Science Conference in Houston in March, 1990 and presented a paper entitled "Solar and IR radiation near the martian surface: A parameterization for CO₂ transmittance" with T. Ackerman, J. Pollack, O.B. Toon and G.E. Thomas as secondary authors. This paper was published in Lunar and Planetary Science XXI, pp. 696-697. NASA contract NASW-4444 paid for some expenses (mostly just my labor), with the Lunar and Planetary Institute paying the rest of the expenses (airfare, rental car, and some food and
hotel) with a grant from the Mars Surface and Atmosphere Through Time (MSATT) program at NASA.

EGS Conference

Dr. Lindner attended the European Geophysical Society Assembly in Copenhagen, Denmark in April, 1990 and presented a paper entitled "CO$_2$ transmittance in the Mars atmosphere: An exponential-sum fit for use in multiple-scattering models" with J. Pollack and T. Ackerman as secondary authors. This abstract was published in Annales Geophysicae. NASA contract NASW-4444 paid all expenses.

Atmospheric Radiation Conference

Dr. Lindner attended the seventh conference on atmospheric radiation in San Francisco in July and presented a paper entitled "The effects of polar clouds and dust on the radiative budget of the martian polar cap". This paper was published in the conference proceedings by the American Meteorological Society. A. D.O.D. contract paid all expenses except for the publication and registration costs related to the Mars work as Dr. Lindner was there to present other research as well (Multispectral Cloud Property Retrieval by B.L. Lindner and R.G. Isaacs).

Cloud Physics Conference

Dr. Lindner attended the cloud physics conference in San Francisco in July (concurrent with the atmospheric radiation
conference) and presented a paper entitled "Clouds and Chemistry on Mars". This paper was published in the conference proceedings by the American Meteorological Society. A. D.O.D. contract paid all expenses except for the publication and registration costs related to the Mars work.

1990 DPS Conference


II.7 Journal Publications

The text of the proposal for this contract was edited and published in the special Mars Polar Processes edition of the Journal of Geophysical Research, Solid Earth and Planets, Volume 95, pages 1367-1379, February 1990. A reprint is included in the Appendix.

A paper by Dr. Lindner entitled "Ozone heating in the martian atmosphere" has been submitted to Icarus. A preprint will be included in the final report.
III. Program of Research for the Next 6 Months

III.1. Research Tasks

Seeing that the primary contract objectives have been satisfied, and that all contract funds have been spent, the next and last 6 months of this contract will be spent preparing the results of this contract for publications and presentations, as outlined below.

III.2. Conferences

Dr. Lindner will attend the International Union of Geodesy and Geophysics Assembly in Vienna, Austria, in August 1991 to present a paper entitled "Mars seasonal CO₂-ice lifetimes and the angular dependence of albedo" in a special Mars climate session. The abstract will appear in the conference proceedings.

Dr. Lindner will also attend the International Symposium on the chemistry and physics of ice, held in Sapporo Japan in Sept. 1991, to present a paper entitled "Why is the north polar cap on Mars different than the south polar cap?" in the extraterrestrial ice session. The abstract will appear in the conference proceedings.

III.3. Publications in Progress

We are in the process of submitting a short paper to Geophysical Research Letters which will present the research presented in sections II.3 and II.4 of this report.
We are also preparing a manuscript to submit to either Nature or Science describing the results of sections II.5 and II.6.

III.4 Proposal to continue this work.

We intend to submit a proposal to NASA to continue this work. We would integrate more accurate simulations of heat transport into our polar-cap model. We would also investigate some processes touched on in the initial proposal to this work which haven't been investigated, such as terrain, snowfall, Warren et al. (1990) predictions of ice emissivity, and the wavelength dependence and bidirectional nature of ice albedo. (As predicted by several peer reviewers who read the proposal for this contract, I did not request sufficient funds for this research).
Publications under this contract
(reprints are in the appendix)


Lindner, B. L., Ozone heating in the martian atmosphere, submitted to Icarus, 1991b.

Lindner, B. L., Why is the north polar cap on Mars different than the south polar cap?, Proceedings, International Symp. on the Phys. and Chem. of Ice, in Press, 1991c.


Bibliography


Warren, S. G., Wiscombe, W. J., and J. F. Firestone, Spectral albedo and emissivity of carbon dioxide in martian polar
Appendix

Reprints of publications made under this contract

(in the order listed earlier)
The Martian Polar Cap: Radiative Effects of Ozone, Clouds, and Airborne Dust

B.L. Lindner (AER)

The solar and thermal flux striking the polar cap of Mars is computed for various ozone, dust, and cloud abundances and for three solar zenith angles. Ozone does not significantly affect the total energy budget of the polar cap. Hence, the observed hemispherical asymmetry in ozone abundance causes only an insignificant hemispherical asymmetry in the polar caps. Vertical optical depths of dust and cloud ranging from zero to one cause little change in the total flux absorbed by the polar cap near its edge, but increase the absorbed flux significantly as one travels poleward. Hemispherical asymmetries in dust abundance, cloud cover, and surface pressure combine to cause a significant hemispherical asymmetry in the total flux absorbed by the residual polar caps, which helps to explain the dichotomy in the residual polar caps on Mars. Other processes which affect the energy budget of the polar cap are proposed and reviewed, particularly with respect to their interaction with the radiative effects of clouds and dust.
CLOUDS AND CHEMISTRY ON MARS

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1. INTRODUCTION

Martian O₃ shows strong seasonal and latitudinal variation, with a mixing ratio ranging from 2x10⁻⁷ppm at equatorial latitudes to a maximum of 0.6ppm over the winter polar atmosphere (Barth et al., 1973). The wealth of data provided by Mariner 9 spawned many theoretical models to understand martian aeronomy. However, the polar hood (a layer of clouds at winter polar latitudes) has been neglected by both observational and theoretical studies of martian aeronomy, even though there have been no studies that have shown whether these effects are indeed negligible. Indeed, the abundance of O₃ within the polar hood is virtually unknown because of the inability of reflectance spectroscopy observations of O₃ to pierce the large optical depths (r) of aerosols. This study examines the variation in the O₃ abundance on Mars with changes in cloud opacities from a theoretical basis; it also questions the efficacy of reflectance spectroscopy for observing O₃ abundances on Mars.

2. MODEL

The O₃ abundance is obtained by solving the coupled one-dimensional continuity equations for O₂, HO₂, O₃, O(3P), O(1D), H, OH, HO₂, and H₂O₂ (discussed in detail in Lindner and Jakosky, 1985; Lindner, 1988). H₂O₂ is allowed to condense on a cold surface. The surface is assumed covered with old ice with a albedo of 0.5. The vertical transport term in the one-dimensional continuity equation uses the eddy diffusion coefficient, K, to describe the processes which mix the atmosphere vertically. Heterogeneous catalytic decomposition of O₃ at the surface is accounted for using the parameterization of Kong and McElroy (1977), modified to account for the lower temperatures and snow cover at the winter poles. The radiative transfer used to calculate the intensity is described in detail elsewhere (Lindner et al., 1990). The discrete ordinate method of Stamnes et al. (1988) is used to treat the scattering and absorption of radiation. The flux is calculated for 57 solar wavelength intervals, including Rayleigh scattering and the absorption and scattering by O₃, CO₂, O₂, H₂O, H₂O₂, HO₂, cloud ice, and dust. Three types of clouds are frequently observed in the polar hood and are selected to test the interaction between O₃ and the polar hood: a low-lying cloud or fog, a cloud layer, and a hazy cloud. The vertical optical depth at the surface (rₒ) used here is unity, which is typical of the polar hood. The single-scattering albedo of ice is taken to be unity and the Heneyy-Greenstein phase function is used.

3. OZONE AND THE POLAR HOOD

The effect of a hazy cloud on the number densities of O₃ family members is shown in Figure 1. H₂O and H₂O₂ are controlled by temperature in the winter polar atmosphere and are not shown. O₃ number densities decrease at all altitudes, primarily because the dominant O₃ loss occurs at an altitude of 20km through HO₂ and HO₃, which increases at this altitude. Unlike the other species, O₃ does not experience strong altitude-dependent changes because O₃ is nearly in diffusive equilibrium (0x is nearly in diffusive equilibrium, and O₃ is the predominant Oₓ species below 30km). Decreases in O₃ number densities near the surface are nearly balanced by increases in O₃ number densities at high altitudes. H, OH, HO₂, O(1D), and O(3P) number densities all exhibit increases above and within the upper part of the cloud, and decreases in the lower part of the cloud. As H, OH, O(3P), and O(1D) all have very short lifetimes, they are more dependent on the local photodissociation rates (J) than are O₃ and HO₂, and exhibit the same altitude dependence as the J’s. J’s increase as much as 10x above the cloud and within the upper 10km of the cloud because of the increase in the effective solar flux resulting from the scattering by the cloud. J’s decrease over 20x in the lower part of the cloud because of a significant reduction in the solar radiation resulting from the scattering to space of the solar flux by the upper portion of the cloud.

Changing the optical depth of the cloud gives results similar to those shown, except that the magnitudes varied slightly. A cloud with rₒ = 3 yields virtually the same results shown, indicating that large clouds (rₒ = 1-3) affect martian aeronomy similarly and irrespective of rₒ. Indeed, O₃ number densities are found to vary by at most a few percent for a cloud of any optical depth, which agrees with Mariner 9 observations (Barth and Dick, 1974).

The most surprising result is that J’s are decreased by only 20% near the surface, when one would expect that a cloud with rₒ = 1 should cause far greater reductions in the available solar flux. The explanation lies in the large solar zenith angle. Without a cloud, the solar flux must traverse an effective optical depth of rₒ/μₒ, where μₒ is the vertical optical depth measured from the surface to infinity and μₒ.

Figure 1: Ratio of the number densities for the hazy cloud case to those of the cloudless case.
is the cosine of the solar zenith angle. In the winter polar regions, $\mu_0$ is quite large. If a cloud is in the path of the solar rays, then most rays are scattered in all directions, although with strong forward scattering. Some solar rays are scattered to space and lost. However, a significant percentage of the solar photons are scattered downward at some angle $\theta$ which is usually less than the solar zenith angle. As a result, $r_v/\cos\theta$ is actually less than the optical depth traversed had the cloud not been there.

When solar zenith angles are as large as they are in the winter polar regions, this is an important effect, explaining why the reduction in J's is only 20% when a factor of 2 or so could be expected. This effect is even more dramatic farther poleward where J's are actually larger at the surface when a cloud is present than they would be if no cloud were present.

The overall pattern in number densities with a flat-lying cloud or fog is similar to the case of the hazy cloud, only shifted toward the surface. Again, $O_3$ is reduced 3-4% at all heights. When a cloud layer is present, the J's and number densities develop a discontinuity at the altitude of the cloud layer. All short-lived species also show a small increase near the surface, caused by a similar increase in the J's near the surface. Below the cloud, the effect of the cloud is to increase J's as the surface is approaches because the solar flux passes through less atmosphere than it would if it traveled along the solar zenith angle, as explained before. The effect is stronger for the optically thick wavelengths that are important for J$(CO_2)$ and J$(H_2O)$.

Other latitudes and seasons are also examined. In general, $O_3$ varies only a few percent with the occurrence of clouds, sometimes increasing slightly, sometimes decreasing slightly.

4. DISCUSSION

Figure 2 shows the column abundance of $O_3$ observed by Mariner 9 in late winter (Barth, 1985). While a general trend in J's is apparent in Figure 2, significant scatter of data points can be seen, particularly at polar hood latitude (poleward of 40° latitude). Furthermore, significant variability occurs in the $O_3$ measurements made over the polar hood, but not in the observations made over the polar cap when no clouds are present (Barth et al., 1973). As demonstrated by the modeling results, the presence of clouds does little to change the $O_3$ abundance, or even to change the altitude dependence of $O_3$. Variations in temperature (cold fronts) have been claimed to account for much of the scatter in the data points at a particular latitude, because the water vapor abundance would vary as well (see Barth et al., 1973; Barth and Dick, 1974; Barth, 1985). However, water vapor is a small source of odd hydrogen in the winter polar atmosphere, and may not account for most of the variability in Figure 2.

Masking by clouds may also account for some of the observed $O_3$ variability, because the nature and opacity of the clouds in the polar hood change dramatically in latitude and even on a day-to-day basis. As the maximum $O_3$ abundance resides near the surface, spacecraft must be able to observe through the entire cloud in order to actually see 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 $r_v$ of the cloud and dust must be traversed. Indeed, part of the observed latitudinal variation in $O_3$ may be due to the inability of the spacecraft to observe through the increasing effective optical depths ($r_v/\mu$) as one goes poleward.

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5. REFERENCES


Fig. 2. Mariner 9 measurements of the $O_3$ column abundance during the northern winter, $L = 330-360^\circ$, in the northern hemisphere (taken with permission from Barth, 1985).
THE EFFECT OF POLAR CLOUDS AND DUST ON THE RADIATIVE BUDGET OF THE MARTIAN POLAR CAP

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1. INTRODUCTION

One of the most puzzling mysteries about the planet Mars is the hemispherical asymmetry in the polar caps. Every spring the seasonal polar cap of CO₂ recedes until the end of summer, when only a small part, the residual polar cap, remains. During the year that Viking observed Mars, the residual polar cap was composed of water ice in the northern hemisphere [Kieffer et al., 1976] but was primarily carbon dioxide ice in the southern hemisphere [Kieffer, 1979]. Scientists have sought to explain this asymmetry by modeling Viking lander observations of atmospheric pressure (since the seasonal polar caps are primarily frozen atmosphere, they are directly related to changes in atmospheric mass). These models are hereafter called "polar cap/atmospheric pressure" models because they examine the energy balance at the surface at all latitudes and seasons, thus simulating the condensation and evaporation of CO₂ ice over the winter polar region. They can therefore be compared to observations of atmospheric pressure and polar cap recession, which are reproduced fairly well, except for the asymmetry in the residual polar caps (Leighton and Murray, 1966; Cross, 1971; Briggs, 1974; Davies et al., 1977; James and North, 1982; Lindner, 1985, 1986). This paper will focus on how ozone, clouds, and airborne dust affect CO₂ ice formation and sublimation to see if they help explain the asymmetry in the residual polar caps.

2. MODELING PROCEDURE

The radiative flux striking the surface is calculated for 57 solar and 10 infrared wavelength intervals from 0 to 100 μm including the absorption, scattering, and emission by O₃, CO₂, clouds, and dust (see Lindner, 1990). Ultraviolet absorption by O₂, H₂O, HO₂, and H₂O₂ is also included, as is Rayleigh scattering. The discrete ordinate method of Stamnes et al. (1988) treats the scattering, emission, and absorption of monochromatic radiation. The exponential sum method allows CO₂ transmittance to be incorporated in a scattering model [see Lindner et al., 1990]. The albedo for the polar cap is 0.5 at solar wavelengths [Kieffer, 1979; James and Lumme, 1982]. The IR albedo of the polar cap is assumed to be zero [Kieffer, 1970; Smythe, 1975; Wiscombe and Warren, 1980]. Three latitude cases were studied: 57°N, 70°N, and 90°N, and late winter conditions were assumed (Ls = 343°). These three cases effectively simulate the edge of the polar cap, the edge of polar night, and polar night, respectively, for all solar longitudes and both polar caps. The temperature profile at 57°N latitude rises linearly with altitude from 150 K at the surface to 130 K at 40 km. The value used for albedo and temperature had little effect on our conclusions.

3. THE EFFECT OF CLOUDS AND DUST

Sublimation depends on the total flux that is absorbed, which is a function of the cap albedo. Weighting the fluxes we calculate with the co-added results in the absorbed flux, as shown in Table 1. Almost no change in the absorbed flux occurs over the range that dust and cloud optical depth experiences in the atmosphere, a result also found by Davies [1979] and Paige [1985]. The increase in the infrared flux due to thermal emission by dust balances the decrease in the solar flux due to dust absorption. It is at the polar night latitudes that the radiative effects of clouds and background dust are most significant, as shown in Table 2. IR fluxes absorbed by the polar cap increase up to an order of magnitude as the cloud opacity and dust opacity increase. Even simply a cloud optical depth of 0.5 and dust optical depth of 0.2 would result in the sublimation of the order of 40 cm of CO₂ ice over 1 martian year. When inserted in a model, this would accelerate the predicted loss of all the CO₂ ice on the northern polar cap.

TABLE 1. Flux Absorbed by the Polar Cap Near Its Edge for Various Cloud and Dust Opacities (57°N Latitude, Ls = 343°)

<table>
<thead>
<tr>
<th>Cloud Opacity</th>
<th>0.0</th>
<th>0.2</th>
<th>0.5</th>
<th>1.0</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>348.2</td>
<td>340.9</td>
<td>333.0</td>
<td>326.7</td>
</tr>
<tr>
<td>0.2</td>
<td>339.0</td>
<td>336.5</td>
<td>333.6</td>
<td>330.1</td>
</tr>
<tr>
<td>0.5</td>
<td>333.1</td>
<td>334.7</td>
<td>336.1</td>
<td>332.6</td>
</tr>
<tr>
<td>1.0</td>
<td>329.8</td>
<td>333.5</td>
<td>338.2</td>
<td>339.2</td>
</tr>
</tbody>
</table>

The vertical optical depth is given. Flux values are given in units of J cm⁻² day⁻¹.

TABLE 2. Flux Absorbed by the Polar Cap in Polar Night For Various Cloud and Dust Opacities (90°N Latitude, Ls = 343°)

<table>
<thead>
<tr>
<th>Cloud Opacity</th>
<th>0.0</th>
<th>0.2</th>
<th>0.5</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>14.7</td>
<td>43.7</td>
<td>74.8</td>
</tr>
<tr>
<td>0.2</td>
<td>36.5</td>
<td>60.9</td>
<td>88.7</td>
</tr>
<tr>
<td>0.5</td>
<td>60.9</td>
<td>81.7</td>
<td>105.9</td>
</tr>
<tr>
<td>1.0</td>
<td>90.3</td>
<td>107.9</td>
<td>128.4</td>
</tr>
<tr>
<td>3.0</td>
<td>152.4</td>
<td>164.7</td>
<td>178.1</td>
</tr>
</tbody>
</table>

The vertical optical depth is given. Flux values are given in units of J cm⁻² day⁻¹.
polar cap by approximately 40 sols (based on ice depth vs. season from Lephoton and Murray [1966], Cross [1971], Briggs [1974], Davies et al. [1977], Lindner [1985], and Jakosky and Haberle [1990]).

While polar cap/ambient pressure models have been unable to explain the dichotomy of the residual polar caps, these models have been fairly successful at reproducing observations of atmospheric pressure and polar cap recession up to 75° or 80° latitude. The radiative effects of clouds and background dust have the added benefit of not significantly affecting model predictions of the annual variation in atmospheric pressure. Clouds and background dust have a neutral effect along the cap edge, becoming more important as one approaches the pole, particularly within 80° latitude. The integral of seasonal CO₂ ice over the 80°S to 90°S latitude band (where the residual polar cap exists) is an order of magnitude less than the integral of seasonal CO₂ ice over the planet. Hence, a process which affects primarily the residual polar cap area will not significantly affect predictions of atmospheric pressure. Also, the rate of recession of the seasonal polar cap would remain essentially the same as that predicted by polar cap/ambient pressure models equatorward of 75° latitude, where good agreement with observations was previously obtained. The behavior near the pole would be significantly modified, making for better agreement with observations than previously obtained. The rate of recession of the polar cap is an important test, especially since it repeats reliably over several years of observation [James and Lumme, 1982; James, 1982].

4. IMPLICATIONS FOR THE POLAR CAP ASYMMETRY

The radiative effects of clouds and background dust would be most noticeable near the residual polar cap, which spans all of the winter in polar night. Observational evidence suggests a greater abundance of clouds over the northern polar cap than over the southern polar cap. Moreover, it is also expected that the formation and sublimation phases of the seasonal southern polar cap will occur with higher background dust opacities than the formation and sublimation phases of the seasonal northern polar cap. Both of these asymmetries would preferentially increase the fluxes absorbed by the residual polar cap in the north.

Furthermore, the surface pressure is lower over the southern residual polar cap than the northern residual polar cap. If mixing ratios of cloud and dust are assumed to be the same for both poles, then the pressure differential between the poles would result in a dust and cloud opacity over the residual polar cap in the south that is 30% less than over the north during the crucial winter months. Mixing ratios of dust may actually be lower over the southern polar cap during winter months because the mechanisms for raising and maintaining atmospheric dust are pressure dependent [Pollack and Toon, 1982]. Lower mixing ratios over the southern pole would accentuate the asymmetry in cloud and dust opacity. These hemispherical asymmetries would allow relatively less frost to accumulate in the north during the fall and winter and cause the CO₂ surface ice to sublimate more rapidly in the north during the following spring and summer.

However, the effects of clouds and background dust would not allow CO₂ ice to survive year-round in the south, as is observed. Other processes which might allow that include penetration of radiation into and through the ice, bidirectional surface reflectance, ice albedo dependence on solar zenith angle, season, and latitude [Wiscombe and Warren, 1980; Squyres and Veverka, 1982; Paige 1985], lower ice emissivity (S. Warren, personal communication, 1988), decreased heat conduction from the surface to existing ice [Jakosky and Haberle, 1990], snowfall [Pollack and Haberle, 1988], surface residues [Saunders et al., 1986], surface roughness, and wind shifting of ice [Briggs, 1974; James et al., 1979; Kieffer, 1979]. The importance of each of these processes is currently being assessed by B.L. Lindner and B.M. Jakosky. Most of these processes would lengthen CO₂ ice survivability either by increasing condensation or decreasing sublimation of CO₂ ice and would counter the increased sublimation caused by clouds and dust. However, most of these processes work equally effectively at both poles and do not generate an asymmetry. Hemispherical asymmetries in cloud and dust opacity and in surface elevation would result in a hemispherical asymmetry in the net radiative flux from clouds and dust which is absorbed by the polar cap, helping explain the asymmetry in the polar caps.

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5. REFERENCES

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Abstract. The solar and thermal flux striking the polar cap of Mars is computed for various ozone, dust, and cloud abundances and for three solar zenith angles. Ozone does not significantly affect the total energy budget of the polar cap. Hence the observed hemispherical asymmetry in ozone abundance causes only an insignificant hemispherical asymmetry in the polar caps. Vertical optical depths of dust and cloud ranging from zero to 1 cause little change in the total flux absorbed by the polar cap near its edge but increase the absorbed flux significantly as one travels poleward. Hemispherical asymmetries in dust abundance, cloud cover, and surface pressure combine to cause a significant hemispherical asymmetry in the total flux absorbed by the residual polar caps, which helps to explain the dichotomy in the residual polar caps on Mars. Other processes which affect the energy balance at the polar cap are proposed and reviewed, particularly with respect to their interaction with the radiative effects of clouds and dust.

Introduction

One of the most puzzling mysteries about the planet Mars is the hemispherical asymmetry in the polar caps. Every spring the seasonal polar cap of CO₂ recedes until the end of summer, when only a small part, the residual polar cap, remains. During the year that Viking observed Mars, the residual polar cap was composed of water ice in the northern hemisphere [Kieffer et al., 1976] but was primarily carbon dioxide ice in the southern hemisphere [Kieffer, 1979]. Scientists have sought to explain this asymmetry by modeling Viking lander observations of atmospheric pressure (since the seasonal polar caps are primarily frozen atmosphere, they are directly related to changes in atmospheric mass). These models are hereafter called "polar cap/atmospheric pressure" models because they examine the energy balance at the surface at all latitudes and seasons, thus simulating the condensation and evaporation of CO₂ ice over the winter polar region, and can therefore be compared to observations of atmospheric pressure and polar cap recession [Leighton and Murray, 1966; Cross, 1971; Briggs, 1974; Davies et al., 1977; James and North, 1982; Lindner, 1985, 1986]. Polar cap/atmospheric pressure models reproduce most aspects of the observed annual variation in atmospheric pressure fairly accurately. Furthermore, the predicted recession of the northern polar cap in the spring agrees well with observations, including the fact that the CO₂ ice is predicted to completely sublime away. However, these models all predict that the carbon dioxide ice will also sublime away during the summer in the southern hemisphere, as shown in Figure 1. This paper will focus on how ozone, clouds, and airborne dust affect CO₂ ice formation and sublimation to see if they help explain the hemispherical asymmetry in the residual polar caps.

Observations of Ozone, Clouds, and Airborne Dust

Ozone has been observed on Mars by Mariner 6 [Barth and Hord, 1971; Lane et al., 1973], by Mariner 9 [Barth et al., 1973; Lane et al., 1973; Barth and Dick, 1974], by MARS 5 [Krasnopolskaia et al., 1977], and by Earth-based experiments [Broadfoot and Wallace, 1970; Noxon et al., 1976; Traub et al., 1979]. Global-average ozone abundances integrated from the surface to space are of the order of 1 µm atm⁻¹ (1 micron atmosphere = 1 µm atm⁻¹ = 2.69 x 10⁻⁵ cm⁻²). Ozone experiences strong latitudinal variation, increasing significantly over the winter polar regions. In the northern hemisphere, the maximum abundance of ozone observed was 60 µm atm⁻¹ near 60°N latitude in late winter, while the maximum abundance observed in the southern hemisphere was 30 µm atm⁻¹. This hemispherical asymmetry in ozone is reproduced by theoretical modeling [Kong and McElroy, 1977; Shimazaki and Shiotzu, 1979; Lindner, 1985]. There is also strong day-to-day variability in ozone, which arises from the variability in temperature and in the polar hood [Barth and Dick, 1974; Lindner, 1988].

The polar hood, a blanket of clouds which covers the winter polar region, has been observed on a latitudinal and seasonal basis by Mariner 9, MARS 3, and Viking [Leovy et al., 1972; Masursky et al., 1972; Briggs and Leovy, 1974; Martin, 1975; Horoz, 1976; Briggs et al., 1977; Anderson and Leovy, 1978; Tillman et al., 1979; Kondrat'ev and Hunt, 1982; Kahn, 1984; Christensen and Zurek, 1984]. The following synopsis is extracted from these papers. The polar hood is first observed at high latitudes in the northern hemisphere at about the time of autumnal equinox, gradually spreading to obscure the surface to latitudes as low as 60°, and then retreats to the pole in the spring. The disappearance of the hood occurs at about the vernal equinox at lower latitudes, the polar hood has strong daily variation and highly variable and often complex vertical structure and is diffuse on scales of less than 1 km [Briggs and Leovy, 1974]. The main bulk of the polar hood is below 20 km altitude. The thickness at one point is estimated at 1-2 km, centered at 15 km height, with an average mixing ratio of ice of the order of 10⁻⁵ g cm⁻³. The hood is often a wave cloud pattern at lower latitudes, and clouds 10 km high with half an optical depth can appear at the hood's edge. Equatorward of 60° latitude in winter the composition of the clouds may be primarily water ice. Poleward of that, the hood is more diffuse,
there is no evidence to suggest any other behavior.

The optical depth of dust in the atmosphere also experiences significant seasonal variations. Global dust storms tend to occur during southern spring, when Mars is near aphelion [Martin, 1984; Tillman and Leovy, 1984], and tend to have vertical optical depths of dust of 1.0 or greater at visible wavelengths [Pollack et al., 1979; Thorpe, 1981]. During the remainder of the year, atmospheric dust opacities are substantially lower, although still appreciable, and are hereafter called background dust. The vertical optical depth of background dust was observed to vary from 0.3 to 1.0 in the visible as determined by imaging the Sun from Viking landers 1 and 2 [Pollack et al., 1979]. Similar background dust optical depths were inferred from contrast observations by Viking orbiter imaging [Thorpe, 1981], inferred from Viking lander pressure oscillations [Zurek, 1981], deduced by studying polar cap reflection from Earth-based observations [Lumme and James, 1984], and inferred from modeling the annual variation in pressure [James and North, 1982].

The frequency and intensity of dust storms have apparently varied greatly from year to year over the past 60 years [Martin, 1984]. The second and third Mars years that Viking observed were relatively calm, while the fourth year had the most intense global dust storm observed by Viking, during which Viking lander 1 failed [Tillman and Leovy, 1984]. Earlier Mariner 9 imaging of dust storms noted significantly less dust than observed by Viking, with optical depths of dust ranging from 0.1 to 2 [Leovy et al., 1972; Masursky et al., 1972]. The same variation in dust optical depth was inferred from Mariner 9 infrared interferometer spectrometer (IRIS) spectra at 3000 Å [Toon et al., 1977] and was determined by the reduction in the solar flux reflected by the polar cap, as observed by the Mariner 9 UV spectrometer [Pang and Hord, 1973].

Observations are inadequate to accurately describe any latitudinal variation in background dust opacity; orbiter data fail to note any significant latitudinal variation, other than a consistently lower dust opacity over both winter poles. Dust opacities in the winter polar atmosphere have been observed to be a factor of 2 or so lower than elsewhere over the planet when the global dust storms exist [Masursky et al., 1972; Leovy et al., 1972; Pang and Hord, 1973; Kieffer, 1979; James et al., 1979; Lumme and James, 1984; Jakosky and Martin, 1987], and the same may be true during background dust conditions. Surface dust is covered by ice at winter polar latitudes, hindering that source of dust; the winter polar atmosphere is stable, facilitating dust settling; and CO₂ snow forms using dust particles as nuclei, further accelerating dust removal [Pollack et al., 1979]. All of these processes may be active during background dust conditions, resulting in dust opacities that may be becoming gradually lower as one approaches the winter pole.

However, the dust seasonal cycle results in a background dust opacity over the Viking lander which is higher from Lₚ = 180° to 360° (southern winter and spring) than from Lₚ = 0° to 180° (northern winter and spring) [Pollack et al., 1979]. Assuming that little hemispherical asym-
metric occurs for the same time frame in background dust opacity [Martin, 1986], then the formation and sublimation phases of the southern polar cap will occur with higher background dust opacities than those of the northern polar cap. As a result, background dust could affect the formation and sublimation of the southern cap more than the northern cap at polar night latitudes.

Previous Work in This Area

Kuhn et al. [1979] suggested that ozone could play an important role in the energy budget of the polar cap. Kuhn et al. showed that ozone heating of the winter polar atmosphere was non-negligible compared to CO$_2$ heating, particularly near 10 km altitude, and could therefore affect any CO$_2$ condensation in the atmosphere. However, atmospheric heating by aerosols in the winter polar atmosphere has since been shown to be greater than either O$_3$ or CO$_2$ heating, particularly near 10 km altitude [Lindner and Thomas, 1983; Lindner, 1985]. Thus, ozone does not make a large enough contribution to the atmospheric temperature to significantly affect downwelling thermal radiation which strikes the polar cap or any condensation of CO$_2$ in the atmosphere. Ozone also affects the surface energy budget by absorbing incoming solar radiation which would otherwise strike the surface. This issue is addressed in the results section of this paper.

Clouds will also affect the energy budget by scattering solar radiation and emitting thermal radiation. The radiative effects of clouds were shown to be important by Briggs [1974] and James and North [1982]. However, the magnitude of this importance was not accurately determined, since they ignored the effects of clouds at solar wavelengths. Furthermore, they included infrared effects with an atmospheric emissivity of 0.3 when clouds existed, an emissivity of 0.15 for cloudless conditions, and a single atmospheric temperature, assumptions which ignore the variations in cloud opacity and temperature which occur.

The attenuation of solar radiation due to absorption and scattering by dust and the increase in thermal radiation due to emission by dust will also influence the energy budget of polar cap, as previously suggested by Briggs [1974], Davies [1979], James et al. [1979], Kieffer [1979], Martin and Kieffer [1979], James and North [1982], and Paige [1985]. These studies all examined the radiative effects of the global dust storms, which occur primarily in southern spring, and showed that global dust storms do not appear to significantly affect polar cap recession. However, appreciable amounts of background dust exist year-round. Davies [1979], James and North [1982], and Paige [1985] examined the radiative effects of background dust and noted a neutral contribution on the whole (a slight decrease in some cases and a slight increase in others).

However, these background dust studies all ignored polar night latitudes (those latitudes and seasons where the Sun remains below the horizon all day). In addition, Paige [1985] studied only the 90° latitude in his study. Furthermore, all studies ignored the overlap of the radiative effects of CO$_2$ and dust in the 15-µm wavelength region, where both CO$_2$ and dust emit the most thermal radiation [Lindner, 1985], as well as the overlap of CO$_2$ and dust at near-infrared (NIR) wavelengths. Improper treatment of this overlap will result in errors in the downwelling thermal radiation of the order of 30% [Lindner, 1985]. Moreover, Davies [1979] and James and North [1982] accurately modeled the radiative effects at solar wavelengths, but included thermal heating of the surface by simply assuming that half of all the solar radiation absorbed by dust is radiated thermally to be absorbed by the surface. This hypothesis assumes that dust is the primary source of radiative heating and cooling; however, CO$_2$ heating and cooling is also appreciable during background dust conditions (e.g., Lindner, 1985). Furthermore, atmospheric absorption of planetary thermal emission is also appreciable. A better approach to including IR heating of the polar cap is to compute the thermal emission based on the observed atmospheric temperatures, CO$_2$ opacities, and dust opacities, as is done here.

Modeling Procedure

The radiative flux striking the surface is calculated for 57 solar and 10 infrared wavelength intervals from 0 to 100 µm including the absorption, scattering, and emission by O$_3$, CO$_2$, clouds, and dust. Ultraviolet absorption of solar flux by O$_3$, H$_2$O, H$_2$O$_2$, and H$_2$O is also included, as is Rayleigh scattering. Atmospheric composition is taken as 95% CO$_2$ and 0.13% O$_2$ [Owen et al., 1977]. Season-dependent CO$_2$ abundances are taken from Hess et al. [1980], after correcting for elevation [Lindal et al., 1979; Jakosky and Farmer, 1982] and accounting for possible circulation-induced pressure gradients [Haberle et al., 1979]. Ozone abundances were taken from Barth et al. [1973], assuming that ozone mixing ratios in the polar night were the same as at the edge of polar night and adopting altitude profiles from Lindner [1988]. H$_2$O abundances are based on the work of Jakosky and Farmer [1982] and Lindner [1988], and H$_2$O$_2$ and H$_2$O$_2$ abundances are taken from Lindner [1988].

Transmission functions for the 2.7-, 4.3-, and 15-µm bands of carbon dioxide are taken from the line-by-line model results of Gal'tsev and Osipov [1979]. The transmission function, $T_{Re}(T,P,U)$, as a function of temperature $T$, pressure $P$, and CO$_2$ column abundance $U$, was extrapolated from the Gal'tsev and Osipov results (subscript C) to temperatures below 200 K by

$$T_{Re}(T,P,U) = 1 - (1 - T_{Re}(200 K,P,U)) (T/200 K)^Q$$

The exponential Q was found to be 0.45, 0.3, and 0.8 for the 2.7-, 4.3-, and 15-µm bands, respectively, when the temperature dependencies for the Pollack et al. [1981] transmission functions were recast in this form. Using a modified version of the FASCOD transmission model (Clough et al., 1986), the accuracy of these transmission functions was confirmed, and transmission functions were obtained for the 1.3-, 1.4-, 1.6-, 2.0-, 4.8-, and 5.2-µm bands of CO$_2$ (B. L. Lindner et al., manuscript in preparation, 1989).

The properties of clouds and dust are extrapolated from observations at mid-latitudes, given
the lack of observations at winter polar latitudes. Background dust opacities over winter polar latitudes may be less than the Viking lander latitudes, where they vary from 0.2 to 1.0 [Pollack et al., 1979; Lumme and James, 1984]. However, the optical depths of background dust vary from 0 to 1.0 in this work to account for all possible scenarios. The wavelength dependence of the dust opacity is taken from Toon et al. [1977]. A Gaussian profile describes the vertical distribution of dust, with dust opacities confined mostly below 20 km altitude for normal conditions and 50 km altitude for dust storm conditions [see Anderson and Leovy, 1978; Zurek, 1982]. The single-scattering albedo of airborne dust as a function of wavelength is taken from Zurek [1978, 1982] and Toon et al. [1977] at solar and infrared wavelengths, respectively, using a solar average of 0.86 [Pollack et al., 1979]. The Haze-L phase function is used to describe the scattering of radiation by Martian dust [Toon et al., 1977]. The emissivity of airborne dust is high and has been calculated as a function of wavelength from theory and observations [Tarbell et al., 1977; Simpson et al., 1981].

Clouds were assumed to be diffuse with a distribution in altitude of the mixing ratio of cloud ice like that assumed for airborne dust [Briggs and Leovy, 1974]. Cloud opacities are observed to vary significantly, and a range of 0 to 3 is used in this work. The single-scattering albedos of both CO2 and H2O ice clouds are high and are assumed to be unity at all wavelengths from UV to IR. This agrees with single-scattering albedos for Martian water ice clouds derived from IRTH data [Hunt, 1979] and derived for CO2 ice clouds from Mie theory [Hunt et al., 1980] for all wavelengths, to the extent that ice particle radii are known. Single-scattering albedos for the clouds are also given by Moroz [1976] and Freeman and Liu [1979]. These observations agree well with theoretical calculations [Wiscombe and Warren, 1980]. The analytic phase function which has been used most extensively in the literature was introduced by Henyey and Greenstein [1941]. The Henyey-Greenstein phase function has been shown to be valid for ice [Hansen, 1969], and asymmetry factors as a function of wavelength from the UV to IR have been derived for Martian clouds, with a great deal of uncertainty [Hunt, 1979; Hunt et al., 1980].

The discrete ordinate method of Stamnes and Conklin [1984] treats the scattering, emission, and absorption of monochromatic radiation through the Martian atmosphere. The exponential sum method allows the banded wavelength structure of CO2 in the infrared (IR) and near-infrared (NIR) to be treated as monochromatic, allowing for its incorporation in a scattering model [see Freeman and Liu, 1979]. Eight terms in the exponential sum were found to give a good approximation to the transmission in the 15-μm band of CO2, while four terms were used for the NIR bands of CO2 (R. L. Lindner et al., manuscript in preparation, 1989). The use of the exponential sum technique allows the overlap of CO2 and dust opacities in the 15-μm and NIR wavelength regions to be properly treated.

Atmospheric properties are zonally averaged and assumed axially independent. The region from the surface to 40 km altitude is broken into 20 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 [e.g., 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. The model is diurnally averaged [e.g., Cogley and Borucki, 1976]. An albedo for the polar cap at solar wavelengths (0.3-5.4 μm) of 0.5 is used, which is appropriate for late northern winter [Kieffer, 1979; Jane and Lumme, 1982]. The IR (5.4-100 μm) albedo of the polar cap is assumed to be zero [Kieffer, 1970; Smythe, 1975; Wiscombe and Warren, 1980]. Three latitude cases were studied: 57°N, 70°N, and 90°N, and late winter conditions were assumed (LD = 243°). These three cases effectively simulate the edge of the polar cap, the edge of polar night, and polar night, respectively, for all solar longitudes and both polar caps.

The temperature profile at 57°N latitude that is used in this study rises linearly with altitude from 150 K at the surface to 170 K at 6 km altitude and then falls with increasing altitude to 130 K at 40 km. This temperature profile is extracted from radio occultation observations made near winter polar latitudes during background dust conditions [Lindal et al., 1979]. These temperatures also agree with those obtained by the Viking IRTH experiment [Kieffer, 1979; Martin, 1984]. The diurnally averaged heating and cooling rates calculated using these temperatures were very nearly in balance. This indicates that the winter polar atmosphere is nearly in radiative equilibrium, which is the assumption that is used for calculating atmospheric temperatures farther poleward. The temperature profile at 70°N latitude was the same as that used at 57°N latitude except that it had a peak temperature of 155 K at 6 km. Polar night temperatures were assumed to have no inversion and fell linearly from 150 K at the surface to 130 K at 40 km. Temperatures are markedly higher at the time of the global dust storms [e.g., Pollack, 1978], but global dust storms are not considered here.

Radiative Fluxes at the Polar Cap Surface

Table 1 shows the radiative fluxes from 0 to 100 μm integrated over UV, visible, NIR, and three IR wavelength bands. O3 absorbs almost two thirds of the available solar flux from 0 to 0.3 μm at 57°N latitude. The flux in the 0.3-1.0-μm bin is 6.1 cm2 s−1 for 0 dust and no O3 scenarios for this latitude and season 1 sol = 1 Martian day.) However, ozone absorption has only a minor effect on the overall flux at the surface, due to the small contribution that solar radiation in the UV makes to the overall energy budget. Very little O3 or CO2 absorption occurs in the 0.3- to 1.0-μm wavelength bin. Absorption in the CO2 NIR bands events approximately 5% of the available solar flux from 1 to 5 μm from striking the surface at 57°N latitude and LD = 343°. The 15-μm band of CO2 is the dominant gaseous source of thermal radiation which strikes the polar cap [see Lindner, 1985]. The IR flux from the Sun can be seen in the 3.4- to 10-μm and 21- to 100-μm wavelength bins for the dust-free case.
Dust affects the energy budget of the polar cap by absorbing and scattering to space solar radiation that would otherwise strike the cap and by emitting thermal radiation, some of which strikes the cap. Dust absorbs and scatters strongly in the UV and moderately in the visible and NIR as seen in Table 1. Dust also emits thermally at all wavelengths. Note that the flux which strikes the surface from thermal emission by even small abundances of airborne dust is of the same order as thermal emission in the 15-µm band of CO$_2$. Clearly, both CO$_2$ and dust thermal emission are important for the energy budget of the polar cap. It is also important to note that in the 15-µm wavelength band, both dust and CO$_2$ emission are important. Hence, it is important to treat dust and CO$_2$ simultaneously as is done here, and not to calculate cooling rates individually and sum them.

The total wavelength-integrated flux decreases with increasing dust optical depth, as the increase in thermal radiation does not equal the decrease in solar flux. However, sublimation depends on the total flux that is absorbed, which is a function of the cap albedo. Whereas the typical albedo of old ice is 0.5 at solar wavelengths [Kieffer, 1979; James and Lumme, 1982], the albedo at IR wavelengths is close to zero [Kieffer, 1970; Smythe, 1975; Wiscombe and Warren, 1980]. Weighting the flux with the coalbedo (1-albedo) results in the absorbed flux, given at the bottom of Table 1. Almost no change in the absorbed flux occurs over the range that dust optical depth experiences in the Mars atmosphere (omitting dust storms), a result also found by Davies [1979] and Paige [1985]. The increase in the infrared flux due to thermal emission by dust balances the decrease in the solar flux due to dust absorption, although not completely, resulting in a slight decrease in absorbed flux. Global dust storms are another matter, as the atmospheric temperature and the meridional transport of heat are much higher. However, the effect of dust storms on polar cap recession was studied by Briggs [1974], James and North [1982], and Paige [1985], and dust storms do not appear to offer an explanation for the persistence of CO$_2$ ice in the southern hemisphere.

As shown in Table 2, all solar fluxes at 70°N latitude are lower than at 57°N latitude due to the increased solar zenith angle and decreased daylight hours. The larger slant path allows O$_3$, CO$_2$, and dust to be more effective at preventing solar flux from reaching the surface; effective optical depths are nearly an order of magnitude larger than vertical optical depths. Thermal fluxes at 70°N latitude are also lower than thermal fluxes at 57°N latitude (Table 1) because atmospheric temperatures are lower [Lindal et al., 1979]. However, the net effect on the flux absorbed by the polar cap is the same as at 57°N latitude, in that changes in dust opacities from 0 to 0.5 cause virtually no change in absorbed flux.

However, while dust has little influence on the total flux absorbed by the polar cap at sunlight latitudes, a noticeable result occurs in the polar night region (that part of the planet where the Sun does not illuminate the surface at all during an entire day). Table 3 shows the thermal...

### TABLE 1. Flux Striking the Surface Near the Edge of the Polar Cap (57°N Latitude, $L_a$ = 343°)

<table>
<thead>
<tr>
<th>Wavelength Interval, µm</th>
<th>0.0</th>
<th>0.2</th>
<th>0.5</th>
<th>1.0</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0-0.3</td>
<td>2.132</td>
<td>1.429</td>
<td>0.823</td>
<td>0.364</td>
</tr>
<tr>
<td>0.3-1.0</td>
<td>455.608</td>
<td>391.262</td>
<td>321.688</td>
<td>247.947</td>
</tr>
<tr>
<td>1.0-5.4</td>
<td>194.296</td>
<td>167.001</td>
<td>138.951</td>
<td>109.501</td>
</tr>
<tr>
<td>5.4-10.0</td>
<td>2.221</td>
<td>3.255</td>
<td>4.361</td>
<td>5.315</td>
</tr>
<tr>
<td>10.0-21.0</td>
<td>19.894</td>
<td>41.189</td>
<td>61.311</td>
<td>79.603</td>
</tr>
<tr>
<td>21.0-100.0</td>
<td>0.044</td>
<td>16.603</td>
<td>36.640</td>
<td>62.910</td>
</tr>
<tr>
<td>Total flux</td>
<td>674.195</td>
<td>620.739</td>
<td>563.774</td>
<td>505.640</td>
</tr>
<tr>
<td>Absorbed flux</td>
<td>348.177</td>
<td>340.893</td>
<td>333.043</td>
<td>326.734</td>
</tr>
</tbody>
</table>

The vertical optical depth of dust is given. Flux values are given in units of J cm$^{-2}$ sol$^{-1}$.

### TABLE 2. Flux Striking the Surface of the Polar Cap Near the Edge of Polar Night (70°N Latitude, $L_a$ = 343°)

<table>
<thead>
<tr>
<th>Wavelength Interval, µm</th>
<th>0.0</th>
<th>0.2</th>
<th>0.5</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0-0.3</td>
<td>1.105</td>
<td>0.552</td>
<td>0.264</td>
</tr>
<tr>
<td>0.3-1.0</td>
<td>220.890</td>
<td>164.344</td>
<td>121.753</td>
</tr>
<tr>
<td>1.0-5.4</td>
<td>93.550</td>
<td>73.743</td>
<td>58.566</td>
</tr>
<tr>
<td>5.4-10.0</td>
<td>1.078</td>
<td>1.371</td>
<td>2.432</td>
</tr>
<tr>
<td>10.0-21.0</td>
<td>16.348</td>
<td>32.923</td>
<td>48.696</td>
</tr>
<tr>
<td>21.0-100.0</td>
<td>0.021</td>
<td>14.497</td>
<td>32.051</td>
</tr>
<tr>
<td>Total flux</td>
<td>332.992</td>
<td>287.990</td>
<td>263.762</td>
</tr>
<tr>
<td>Absorbed flux</td>
<td>175.220</td>
<td>168.571</td>
<td>173.471</td>
</tr>
</tbody>
</table>

The vertical optical depth of dust is given. Flux values are given in units of J cm$^{-2}$ sol$^{-1}$.
flux absorbed by the polar cap for three dust opacities (solar fluxes are obviously zero). Again, note that thermal fluxes are lower at polar night latitudes than at sunlit latitudes (i.e., Tables 1 and 2) due to the extremely cold temperatures. Even so, there is a substantial increase in absorbed flux with dust opacity, even for very small opacities. For comparison, the thermal emission by the surface to space is only 260 J cm⁻² sol⁻¹, assuming unit emissivity. In terms of a sublimation rate, 10 J cm⁻² sol⁻¹ would sublimate approximately 0.02 cm of CO₂ ice per sol (this assumes an ice density of 1 g cm⁻³). Hence, a period of 1 Martian year with a vertical optical depth of 0.2 would result in a substantial loss of the order of 20 cm of CO₂ ice.

The Effect of Clouds

Clouds also affect the energy budget of the polar cap. At 57°N latitude, the radiative effects on the absorbed flux of a combination of clouds and dust are fairly neutral, with a slight downward bias (Table 4). Clouds slightly decrease the flux absorbed by the polar cap for small dust opacities (rD = 0 or 0.2), with the extent of the decrease proportional to the cloud opacity. However, clouds actually slightly increase the flux absorbed by the polar cap at 57°N latitude for moderate or large dust opacities (rD = 0.5 or 1). The explanation lies in the fact that cloud ice is strongly scattering and in the fact that the solar zenith angles are very large in the winter polar atmosphere. With a cloud, the solar flux must traverse an effective optical depth of rµ₀, where r is the vertical optical depth measured from the surface to infinity, and µ₀ is the cosine of the solar zenith angle. In the winter polar regions, µ₀ is quite small (diurnally averaged µ₀ is approximately 0.32 for this case). If a cloud is in the path of the solar rays, then most rays are scattered in all directions, although with strong forward scattering. Some solar rays are scattered to space and lost. However, a significant percentage of the solar photons are scattered downward at some angle θ which is usually less than the solar zenith angle. As a result, r/cos θ is actually less than the optical depth traversed had the cloud not been there. When solar zenith angles are as large as they are in the winter polar regions, this is an important effect. This effect is even more dramatic farther poleward. The same effect is easily noticed on Earth, especially just after sunset. If a cloud is present, then the surface illumination can be significantly greater than if no cloud were present because the cloud scatters the solar flux down to the surface.

For low dust opacities, the effect of lowering the effective solar zenith angle is outweighed by the scattering of solar flux to space. For moderate or high dust opacities, lowering the effective solar zenith angle decreases the effective optical depth of dust that must be traversed and allows more solar radiation to reach the surface. This is even more important in NIR and UV wavelengths, where CO₂ and O₃ have absorption bands. This effect was noted to be particularly important for ozone photochemistry on Mars [Lindner, 1988]. Thermal emission by clouds increases the IR flux absorbed by the polar cap, most noticeably in the 21- to 100-µm wavelength range. Thermal emission by clouds in the 15-µm and can get absorbed by CO₂ before striking the surface. At 70°N latitude, the radiative effect is also neutral for small cloud opacities and all dust opacities, as shown in Table 5. However, moderate and large clouds cause up to a 15% increase in the flux absorbed by the polar cap. At 70°N

---

**TABLE 4. Flux Absorbed by the Polar Cap Near Its Edge for Various Cloud and Dust Opacities (57°N Latitude, Lₙ = 343°)**

<table>
<thead>
<tr>
<th>Cloud Opacity</th>
<th>0.0</th>
<th>0.2</th>
<th>0.5</th>
<th>1.0</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>348.177</td>
<td>340.893</td>
<td>333.043</td>
<td>326.734</td>
</tr>
<tr>
<td>0.2</td>
<td>339.769</td>
<td>336.545</td>
<td>333.569</td>
<td>330.136</td>
</tr>
<tr>
<td>0.5</td>
<td>333.106</td>
<td>334.744</td>
<td>334.163</td>
<td>332.564</td>
</tr>
<tr>
<td>1.0</td>
<td>329.832</td>
<td>333.467</td>
<td>338.226</td>
<td>339.172</td>
</tr>
</tbody>
</table>

The vertical optical depth is given. Flux values are given in units of J cm⁻² sol⁻¹.
TABLE 5. Flux Absorbed by the Polar Cap Near the Edge of Polar Night for Various Cloud and Dust Opacities (70°N Latitude, \( L_\phi = 343° \))

<table>
<thead>
<tr>
<th>Cloud Opacity</th>
<th>Dust Opacity</th>
<th>0.0</th>
<th>0.2</th>
<th>0.5</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>175.220</td>
<td>168.571</td>
<td>173.471</td>
<td></td>
</tr>
<tr>
<td>0.2</td>
<td>167.139</td>
<td>170.316</td>
<td>181.022</td>
<td></td>
</tr>
<tr>
<td>0.5</td>
<td>169.478</td>
<td>178.580</td>
<td>192.287</td>
<td></td>
</tr>
<tr>
<td>1.0</td>
<td>184.456</td>
<td>208.631</td>
<td>195.178</td>
<td></td>
</tr>
</tbody>
</table>

The vertical optical depth is given. Flux values are given in units of J cm\(^{-2}\) sol\(^{-1}\).

While cloud optical depths as large as 3 could occur, optical depths of 0.5 are probably more common. However, even a cloud of optical depth of 0.3 and dust optical depth of 0.2 would result in the sublimation of the order of 40 cm of CO\(_2\) ice over 1 Martian year. When inserted in a polar cap/atmospheric pressure model, this would accelerate the predicted loss of all the CO\(_2\) ice on the northern polar cap by approximately 40 sols (based on ice depth versus season as shown by Leighton and Murray [1966], Cross [1971], Briggs [1974], Davies et al. [1977], Lindner [1985], and Jakosky and Haberle [this issue]).

As discussed in the introduction, while polar cap/atmospheric pressure models have been unable to explain the dichotomy of the residual polar caps, these models have been fairly successful at reproducing observations of atmospheric pressure and polar cap recession up to 75° or 80° latitude. The radiative effects of clouds and background dust have the added benefit of not significantly affecting polar cap/atmospheric pressure model predictions of the annual variation in atmospheric pressure. Clouds and background dust have a neutral effect along the cap edge, becoming more important as one approaches the pole, particularly within 80° latitude. The integral of seasonal CO\(_2\) ice over the 80°S to 90°S latitude band (where the residual polar cap exists) is an order of magnitude less than the integral of seasonal CO\(_2\) ice over the planet [Leighton and Murray, 1966; Briggs, 1974; Davies et al., 1977; Lindner, 1985]. Hence, a process which affects primarily the residual polar cap area will not significantly affect model predictions of atmospheric pressure. Also, the rate of recession of the seasonal polar cap would remain essentially the same as that predicted by polar cap/atmospheric pressure models equatorward of 75° latitude, where good agreement with observations was previously obtained. The behavior near the pole would be significantly modified, making for better agreement with observations than previously obtained. The rate of recession of the polar cap is an important test, especially since it repeats reliably over several years of observation [Weaver and Goguen, 1973; James et al., 1979; James and Lumme, 1982; James, 1982].

The Effect of Albedo and Temperature on Our Results

A polar cap albedo of 0.5 at visible wavelengths was selected for the calculations presented here, which is appropriate during the sublimation phase of the polar cap (late winter, spring) [Kieffer, 1979; James and Lumme, 1982]. One would expect higher albedos during the formation phase of the polar cap (fall, early winter) because of the freshly formed ice; unfortunately, observations of the polar cap albedo during the formation phase of the polar cap are not available, due to the extensive cloud cover at this time. Higher polar cap albedos would lessen the amount of absorbed solar flux, making the thermal effects of clouds and dust relatively more important to the energy budget of the polar cap. A polar cap albedo of 1 would change the polar cap energy budget by nullifying the solar contributions in Tables 1 and 2. For dust-free, cloud-free cases, the total flux absorbed by the polar...
cap at 57°N latitude in late winter would be only 5% of the value obtained for an albedo of 0.5. However, a scenario with a cloud of vertical optical depth 1 and dust of vertical optical depth 0.5 at 57°N latitude in late winter would still result in a flux absorbed by the polar cap for an albedo of 1 that is 50% of the flux absorbed for an albedo of 0.5. Hence, the presence of clouds and dust lessens the ability for a high albedo to preserve CO₂ ice. Furthermore, polar cap albedo does not have a linear effect on the energy budget, as clouds and dust will reflect some of the flux reflected from the polar cap back onto the polar cap. Also, a high albedo at solar wavelengths is of no consequence at polar night latitudes where the Sun never shines.

Similarly, a low albedo at solar wavelengths is not as damaging for the preservation of CO₂ ice as might be expected, because a substantial portion of the energy budget of the polar cap is received from IR emission by clouds and background dust. Hence, any increase in absorbed flux at solar wavelengths from a decrease in albedo is not as important to the energy budget as it would be in the absence of clouds and dust. A polar cap albedo of zero would approximately double the absorbed solar flux given in the tables, but the percentage increase to the total energy budget is smaller when the effects of clouds and dust are included. Furthermore, a low albedo at solar wavelengths has no importance at polar night latitudes.

Atmospheric temperatures used in this model for the edge of the polar cap were some of the coldest observed at polar cap edge latitudes by the Viking orbiter [Lindal et al., 1979]. Higher atmospheric temperatures than those used here would actually increase the thermal emission by dust and cloud even more, making their radiative effects yet more important to the overall energy budget. However, the temperatures used are not as important as one might expect, because most atmospheric dust and cloud lie near the surface (about 60% is within 8 km of the surface; one atmospheric scale height at 160 K), and atmospheric temperatures near the surface remain very close to the CO₂ frost point temperature during winter except during dust storms [Lindal et al., 1979].

Implications for the Polar Cap Asymmetry

As ozone is more prevalent at northern latitudes [Barth et al., 1973], ozone was proposed to be partly responsible for the hemispherical asymmetry in residual polar caps [Kuhn et al. 1979]. The maximum ozone abundance observed by Mariner 9 is used to test this hypothesis. Higher ozone abundances probably exist under cloudy and dusty conditions through which the reflectance spectroscopy technique employed by Mariner 9 researchers would have had difficulty detecting ozone [Lindner, 1988]. However, both ozone heating of the atmosphere and the reduction in light striking the surface would be less important under very cloudy, dusty conditions [Lindner, 1988].

Ozone does absorb most of the ultraviolet insolation, which increases the lifetime of CO₂ ice. However, ultraviolet insolation is only 1% of the total insolation. Thus, even if O₃ absorbs all ultraviolet insolation, CO₂ ice lifetimes would only be marginally changed. Ozone heating of the atmosphere is also minor compared to CO₂ and dust heating [Lindner and Thomas, 1983; Lindner, 1985]. Hence, ozone does not cause any appreciable hemispherical asymmetry in atmospheric temperature, and therefore in the atmospheric thermal emission which strikes the polar cap, or in any CO₂ condensation which may take place in the atmosphere.

However, the cloud and dust results have more important implications regarding the hemispherical asymmetry of the residual polar caps. Although the results are dependent on the polar cap albedo and atmospheric temperatures, it does appear that the radiative effects of both clouds and background dust will appreciably increase sublimation of the polar cap at polar night latitudes. Hence, the radiative effects of clouds and background dust would be most noticeable near the residual polar cap, which spends all of the winter in polar night. As discussed in the introduction, observational evidence suggests a greater abundance of clouds over the northern polar cap than over the southern polar cap. Moreover, as also discussed in the introduction, it is also expected that the formation and sublimation phases of the seasonal southern polar cap will occur with higher background dust opacities than the formation and sublimation phases of the seasonal northern polar cap. Both of these hemispherical asymmetries would preferentially increase the fluxes absorbed by the polar cap at polar night latitudes in the north, particularly by the residual polar cap area in the north.

Furthermore, the surface pressure during winter months is lower over the southern residual polar cap than the northern residual polar cap, due to differences in elevation [Voiceshyn, 1974; Lindal et al., 1979] and due to the annual cycle in surface pressure [Hess et al., 1980]. If mixing ratios of cloud and dust are assumed to be the same for both poles, then the pressure differential between the poles would result in a dust and cloud opacity over the residual polar cap in the south that is 30% less than over the residual polar cap in the north during the crucial winter months. Mixing ratios of dust may actually be lower over the southern polar cap than over the northern polar cap during winter months because the mechanisms for raising and maintaining atmospheric dust are pressure dependent [Pollack et al., 1976, 1979; Pollack and Toon, 1982]. Lower mixing ratios over the southern pole would accentuate the asymmetry in cloud and dust opacity due to the asymmetry in surface pressure.

The net result of all these hemispherical asymmetries is that the solar and infrared flux absorbed by the residual polar cap area in the north during the winter is greater than that absorbed by the residual polar cap area in the south. This would allow relatively less frost to accumulate over the residual polar cap in the north during the fall and winter than in the south and cause the CO₂ surface ice to sublime more rapidly in the north during the following spring and summer. These suspected hemispherical asymmetries in background dust and atmospheric pressure have not been included in prior polar cap/atmospheric pressure studies.
Other Processes Neglected by Prior Polar Cap/Atmospheric Pressure Models

The radiative effects of cloud and dust are not the only processes responsible for the hemispherical asymmetry in the residual polar caps. Prior polar cap/atmospheric pressure studies determined that CO₂ ice would completely sublime away from both polar caps. This study has shown that clouds and background dust serve to increase sublimation over the poles, albeit to a greater extent over the northern pole. Some other phenomenon that has not been included in prior polar cap/atmospheric pressure models is causing the ice to sublime more slowly, to allow CO₂ ice to survive southern summer.

Previous polar cap/atmospheric pressure models have assumed that all absorption of sunlight occurs at the top surface of the polar cap, when in reality it can be distributed over as much as several meters depth [Clow, 1987]. This simplifies the treatment of ice microphysics in the model, but it means that the effects of the finite grain size of ice, background dust, and the energy balance vary significantly with solar zenith angle. This is an important effect that is not considered in the winter polar regions where solar zenith angles vary significantly. The inclusion of the dependence of ice albedo on the solar zenith angle is important for thin ice, as much of the incoming radiation would go through the ice and get absorbed by the surface under the cap. Considering that the seasonal cap is at most 1 m thick [Leighton and Murray, 1966; Briggs, 1974; Davies et al., 1977; Lindner, 1985], this effect alone should extend the predicted stability of surface ice in the summer. Third, previous models of the sublimation of Martian ice used an empirical relationship for sublimation developed for turbulent-driven evaporation from terrestrial oceans [Clow, 1987]. This parameterization fails to account for environmental differences between the Earth and Mars and differences in the microphysics of ice, because models without such a term yield incorrect sublimation rates for thin ice.

First, the rate of sublimation/condensation of a column of ice depends on the integral of the energy balance at each level within the ice, which is not the same as assuming that all sources and sinks occur at the top surface. Second, accurate treatment of ice microphysics is particularly important for thin ice, as much of the incoming radiation would go through the ice and get absorbed by the surface under the cap. Considering that the seasonal cap is at most 1 m thick [Leighton and Murray, 1966; Briggs, 1974; Davies et al., 1977; Lindner, 1985], this effect alone should extend the predicted stability of surface ice in the summer. Third, previous models of the sublimation of Martian ice used an empirical relationship for sublimation developed for turbulent-driven evaporation from terrestrial oceans [Clow, 1987]. This parameterization fails to account for environmental differences between the Earth and Mars and differences in the microphysics of ice, because models without such a term yield incorrect sublimation rates for thin ice.

Furthermore, all prior models ignored CO₂ absorption by the atmosphere in the NIR, which is more important at winter polar latitudes where the slant paths are so large [Lindner, 1985]. As much as 10% of the solar flux can be absorbed, which would allow CO₂ ice to linger longer into the summer months. Moreover, heating within the polar cap in near-infrared wavelengths is actually larger than heating in visible wavelengths in the top few millimeters of ice, because near-infrared radiation does not penetrate the ice as deeply as visible radiation [Clow, 1987]. This accentuates the effect that atmospheric absorption of near-infrared sunlight has on polar cap sublimation.

Previous models assumed a Lambertian (isotropic) surface albedo. However, ice does not scatter isotropically, particularly for solar zenith angles above 66° where ice is strongly forward scattering [Taylor and Stowe, 1984]. Since the solar zenith angle in the winter polar region of Mars is usually much greater than 66°, the assumption of an isotropic albedo for the polar cap is in error. This error becomes important when the atmospheric feedbacks of clouds and dust become important, because clouds and dust do scatter a significant amount of solar radiation which has been reflected by the polar cap rather than on the polar cap. This is particularly important for high-albedo objects such as the polar caps and the polar hood where solar radiation can reflect back and forth many times before escaping to space or getting absorbed.

Furthermore, the total albedo of ice has a strong dependence on solar zenith angle [Lins and Cess, 1977; Wiscombe and Warren, 1980; Squyres and Werren, 1982; Paige 1985], which is neglected by prior polar cap/atmospheric pressure models. Ice brightens significantly at high solar zenith angles. This is an important effect to consider in the winter polar regions where solar zenith angles vary significantly. The inclusion of the dependence of ice albedo on the solar zenith angle in the model will result in a smaller error for thin ice, as much of the incoming radiation would go through the ice and get absorbed by the surface under the cap. Considering that the seasonal cap is at most 1 m thick [Leighton and Murray, 1966; Briggs, 1974; Davies et al., 1977; Lindner, 1985], this effect alone should extend the predicted stability of surface ice in the summer. Third, previous models of the sublimation of Martian ice used an empirical relationship for sublimation developed for turbulent-driven evaporation from terrestrial oceans [Clow, 1987]. This parameterization fails to account for environmental differences between the Earth and Mars and differences in the microphysics of ice, because models without such a term yield incorrect sublimation rates for thin ice.

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mation of the polar cap occurs in the near-infrared [Clow, 1987]. The albedo of ice is sensitive to grain size in the near-infrared and sensitive to the dust to ice mixing ratio in the visible [Warren and Wiscombe, 1980; Wiscombe and Warren, 1980].

Seasonal and latitudinal variations in ice albedo occur on the Earth due to changes in dust mixing ratios and ice age [Wiscombe and Warren, 1980; Warren and Wiscombe, 1980] and should also exist on Mars. Latitudinal variations in ice albedo have been inferred from Mariner 7 spectra [Martin, 1988], and a consistently higher albedo for surface ice is inferred to exist in the southern hemisphere based on Viking orbiter observations [Paige, 1985]. This would lower the energy absorbed by the southern polar cap and allow the CO$_2$ ice to survive the summer. Seasonal and latitudinal variations in polar cap albedo have been noted to make better agreement between polar cap/anthmospheric pressure models and observations of atmospheric pressure [Lindner, 1986].

Previous polar cap/anthmospheric pressure models have only examined the condensation of frost and have ignored snowfall. Snowfall could be an appreciable component of the polar cap [Pollack and Haberle, 1988], as it is on Earth, and therefore could account for the lingering of carbon dioxide ice. Carbon dioxide clouds probably exist over part of the polar cap [Briggs and Leovy, 1974; Hunt et al., 1980; Paige, 1985], and carbon dioxide snow could also be expected to form. Furthermore, snowfall should occur mostly at polar night latitudes. Sunlit polar winter latitudes have a strong inversion which keeps atmospheric temperatures well above the CO$_2$ condensation temperature even while frost is condensing on the surface [Lindal et al., 1979]. However, in polar night the heat sources which support the temperature inversion at lower latitudes are not as effective, allowing for lower atmospheric temperatures which favor the formation of CO$_2$ clouds and snow. Hence, snowfall might not be an important contributor to the polar energy budget at latitudes equatorward of 70°, which is where most of the CO$_2$ frost occurs [Leighton and Murray, 1966; Briggs, 1974; Davies et al., 1977; Lindern, 1985]. Thus, including snowfall in polar cap/anthmospheric pressure models will not appreciably change the predictions of the annual cycle in pressure or the predictions of polar cap recession equatorward of 70° latitude, both of which agree fairly well with observations. However, snowfall might be appreciable poleward of 70° latitude, particularly at 90° latitude, where previous models have been incorrect. Snowfall would also help counterbalance the radiative effects of clouds and dust, which were also most important in polar night. Furthermore, southern winter is colder and longer than northern winter due to the large eccentricity of the Martian orbit. Thus, the southern pole might accumulate more snowfall than the northern pole, which also might help explain why CO$_2$ frost remains year-round at 90°S latitude and not at 90°N latitude.

Jakosky and Haberle [this issue] noted that prior polar cap/anthmospheric pressure models always began with the assumption that the south pole was bare of a CO$_2$ ice residual polar cap. This allowed the surface to warm during summer and hence delayed the onset of frost in the fall. Jakosky and Haberle showed that the assumption that a CO$_2$ ice residual polar cap existed in southern summer resulted in CO$_2$ frost accumulating earlier the following fall and hence allowed it to survive longer the next summer. This idea has the added benefit of being most important at polar night latitudes and would therefore help to counter the radiative effects of clouds and dust, which are also most important at polar night latitudes.

Recent radar observations of the south polar cap suggest the presence of significant roughness [D. O. Muhleman, personal communication through B. M. Jakosky, 1988]. Theoretically, surface roughness in the form of penitentes (spikes of ice) may occur in the polar caps [Svitek and Murray, 1988]. Surface roughness will change the effective solar zenith angle. Surface roughness will also allow radiation to be reflected from one part of the polar cap directly to another part of the polar cap without atmospheric scattering as an intermediary.

Recent calculations of the emissivity of CO$_2$ ice as a function of wavelength from the UV to the IR under Marslike conditions show strong variability in emissivity with dust content and ice age (S. Warren, personal communication, 1988). Unidirectional surface effects, the emissivity can be quite low, which would decrease the ability of the ice to radiate in the IR.

Other possibilities include an insulating residue on the polar ice [Saunders et al., 1986], which would reduce the rate of sublimation. Wind shifting of the polar cap ice may also change the recession rate of the polar cap by moving ice toward the pole [Briggs, 1974; James et al., 1979; Kieffer, 1979; Saunders et al., 1985].

Summary

Polar cap/anthmospheric pressure models have been unable to explain the observed hemispherical asymmetry in the composition of the residual polar caps on Mars. However, the radiative effects of ozone, clouds, and background dust were not fully modeled. Thus, including snowfall in the energy budget of the polar cap. Clouds and dust have a net radiative effect that is quite small near the edge of the polar cap, but it becomes more substantial in increasing the sublimation rate as one goes poleward. In fact, the net radiative effect of clouds and dust at the pole itself increases the loss of CO$_2$ ice over 1 Martian year of the order of 40 cm or more, which would result in the loss of CO$_2$ ice of the order of 40 sols earlier than previously predicted.

However, several other processes have also not been fully included in polar cap/anthmospheric pressure models. These processes include penetration of radiation into and through the ice, bidirectional surface reflectance, ice albedo dependence on solar zenith angle, season, and latitude, and other ice emissivity, decreased heat conduction from the surface due to existing ice, snowfall, surface residues, surface roughness, and wind shifting of ice. The importance of each of these processes to the energy budget is currently being assessed by B. L. Lindern and B. M. Jakosky. Most of these processes would lengthen...
CO₂ ice survivability either by increasing condensation or decreasing sublimation of CO₂ ice and would counter the increased sublimation caused by clouds and dust. However, most of these processes work equally effectively at both poles and do not generate an asymmetry. Hemispherical asymmetries in cloud and dust opacity and in surface elevation would result in a hemispherical asymmetry in the net radiative flux from clouds and dust which is absorbed by the polar cap. The key point is that the net radiative effect of clouds and background dust is to shorten ice lifetimes on the north pole relative to the south pole, helping to explain the asymmetry in the residual polar caps on Mars. Furthermore, the radiative effects of clouds and dust do not appear to significantly affect model predictions of polar cap recession equatorward of 75° latitude or model predictions of the annual cycle in atmospheric pressure, both of which agreed fairly well with observations. The presence of clouds and dust also lessens the ability for a higher albedo to preserve CO₂ ice and for a lower albedo to increase sublimation, decreasing the importance of albedo in determining ice sublimation rates.

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Penetration of Light Into the Martian Polar Cap: Implications for the Energy Budget

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We have modified the Jakosky and Haberle (J. Geophys. Res., 95, 1359, 1990) model of the seasonal polar cap on Mars to include penetration of solar radiation into the cap itself, based on the theoretical work of Clow (Icarus, 72, 95, 1987). We find that the inclusion of light penetration slightly decreases the albedo needed in the model to keep CO$_2$-ice year-round at the south pole by on the order of 1%. The required albedo is decreased because some solar radiation is used to heat the subsurface, and not all of this heat is transported back to the surface. Furthermore, the number of sols for which the north pole is free of CO$_2$-ice is increased slightly by approx. 5 sols. By allowing solar radiation to be absorbed at depth, the surface does not become as warm. This decreases the infrared radiation emitted by the surface and increases the heat conduction from the subsurface, which in turn delays the onset of frost formation in the fall.

Overall, we conclude that the penetration of light into the polar cap has only a small effect on the energy budget of the polar cap. Given the uncertainties currently present in albedo and other parameters, the effect of light penetration is second order, and can be neglected in models of the polar cap energy budget.
AN EFFICIENT AND ACCURATE TECHNIQUE TO COMPUTE THE ABSORPTION, EMISSION, AND TRANSMISSION OF RADIATION BY THE MARTIAN ATMOSPHERE. Bernhard Lee Lindner, AER Inc., 840 Memorial Drive, Cambridge, MA 02139; Thomas P. Ackerman, Dept. of Meteor., Penn. State Univ., University Park, PA 16802; and James B. Pollack, NASA/ARC, Moffett Field, CA 94035.

INTRODUCTION. CO$_2$ comprises 95% of the composition of the martian atmosphere [1]. However, the martian atmosphere also has a high aerosol content. Dust opacities vary from less than 0.2 to greater than 3.0, primarily on a seasonal basis with the occurrence of global dust storms during southern spring [2]. Ice-cloud opacities vary from 0 to greater than 1, with large amounts occurring at winter polar latitudes [3]. CO$_2$ is an active absorber and emitter in near-IR and IR wavelengths; the near-IR absorption bands of CO$_2$ provide significant heating of the atmosphere, and the 15 $\mu$m band provides rapid cooling [4-7]. However, dust and ice-cloud aerosols have high scattering albedos in solar wavelengths, and are highly absorbing at infrared wavelengths, and are as important as CO$_2$ in the atmospheric energy budget [5].

Including both CO$_2$ and aerosol radiative transfer simultaneously in a model is difficult. Aerosol radiative transfer requires a multiple-scattering code, while CO$_2$ radiative transfer must deal with complex wavelength structure, as shown in Fig. 1. The problem can be solved exactly by inserting the CO$_2$ absorbance for each spectral line into a multiple-scattering code, but the 15$\mu$m band alone has on the order of 10,000 lines, making such a computation tedious and expensive. It is this difficulty of simultaneously treating aerosol multiple scattering and the banded absorption structure of CO$_2$ that prompts most radiative-transfer studies of the martian atmosphere to consider either a pure-CO$_2$ or pure-dust atmosphere. This approximation simplifies treatment, but is inaccurate.

One alternative technique that has recently been developed for atmospheric applications is the exponential-sum or k-distribution approximation [8-21]. The transmission of a homogeneous atmosphere is actually independent of the ordering of the absorption coefficient, $k$, in frequency space within a spectral interval, depending only upon the percentage of the spectral interval that has a particular value of $k$. The percentage of the spectral interval which has values between $k$ and $k + \Delta k$ can be formulated in a probability density function $f(k)$ shown schematically in Fig. 2. The chief advantage of the exponential-sum approach is that the integration over $k$ space of $f(k)$ can be computed more quickly than the integration of $k$ over frequency. The exponential-sum approach is superior to the photon-path-distribution and emissivity techniques for dusty conditions [22, 19, 23]. Our work is the first application of the exponential-sum approach to martian conditions.

THEORETICAL APPROACH. The transmittance of the 15 $\mu$m band and the near-IR bands of CO$_2$ was computed using the FASCOD line-by-line transmittance model [24], modified for martian conditions. Computations with the modified FASCOD model were made at 3 temperatures (125K, 200K, 300K) and 5 pressures (100 mb, 10 mb, 1 mb, 0.1 mb, 0.01 mb); these cover the range of temperature and pressure currently observed in the atmosphere at all latitudes, seasons, and altitudes up to 40 km, and also can be used for early Mars, dense atmosphere studies. The near-IR and 15$\mu$m bands were broken into spectral sub-intervals (see Fig. 1), and the

![Figure 1. CO$_2$ 15 $\mu$m band transmission at 20 km altitude looking upward, at temperatures of 200 and 300K [25]. Also shown are the sub-intervals used for the 8 term and 16 term fits.](image1)

![Figure 2. A schematic illustration portraying the essence of the exponential-sum approach. (a) shows a schematic of absorption line spectra at two different pressures. In (b) the two probability density functions $f(k)$ associated with (a) are illustrated. The shaded area depicts the strongest absorption (i.e., largest $k$) for the same spectral interval (i.e., $f(k)$ for different pressures are correlated). Integration of $f(k)$ over $k$ replaces the integration of $k$ over $\nu$ (modified from [28], [29]).](image2)
transmittance for each layer, $T_r$, as a function of CO$_2$ column abundance within that layer, $u$, was fit with a series of weighting coefficients, $a_i$, and exponential coefficients, $b_i$:

$$T_r(u) = \sum_{i=1}^{n} a_i \exp(-b_i u)$$  \hspace{1cm} (1)

We tried fitting procedures based on Wiscombe and Evans [14] and an improved version of Ackerman et al. [13], and found both yielded similar results. Both of these procedures avoid the ill-conditioning of earlier exponential-sum routines, and produce more accurate and unique solutions [13,14]. The exponential-sum fit reproduced the FASCOD transmittances to better than 10$^{-4}$ for all CO$_2$ abundances considered. The frequency sub-intervals are picked to try to minimize the variation in line strengths within the sub-interval. As shown in Fig. 1, 2 sub-intervals covered the band center, 2 covered the far line wings, 2 covered the near line wings, and 2 covered the transition from wings to band center. The vertical inhomogeneity of the atmosphere is treated by using homogeneous layers, with the absorption coefficients $k$ for all layers correlated in frequency space, i.e. the sub-intervals of the spectral band which have the maximum absorption also have the largest $k$ values (Ackerman et al., 1976). An interpolation is used for temperatures and pressures which fail in between the values at which the exponential-sum coefficients were computed. We compared a logarithmic interpolation to a linear interpolation, and found the logarithmic interpolation to be more accurate.

**INCORPORATION IN MULTIPLE-SCATTERING MODELS.** Vertical optical depths of CO$_2$ absorption for each term number, frequency sub-interval, and atmospheric layer are $b_i u$. CO$_2$ is combined with dust and cloud in that the total optical depth $T_i$, single-scattering albedo $\omega_i$, and phase function $P$ for each term, frequency sub-interval, and layer are given by [16]:

$$T_i = b_i u + r_s + r_p + r_s + r_p$$  \hspace{1cm} (2)

$$\omega_i = \frac{(r_i + r_s)}{T_i}$$  \hspace{1cm} (3)

$$P = \frac{D_i R + C_i C + R_i R}{r_s + r_p}$$  \hspace{1cm} (4)

where:  
- $r_i$ = Rayleigh scattering optical depth for that layer
- $D_i, C_i, a_i$ = Dust (D) and Cloud (C) scattering (s) and absorption (a) optical depth for that layer
- $P = D_i, C_i, R_i$ = dust (D), cloud (C), and Rayleigh scattering (R) phase function

The multiple-scattering code is run once for each term in the sum using the $T_i$, $\omega_i$, and $P$ appropriate to that term, and the resultant fluxes, $F_i$, (or intensities) are then summed and weighted by $a_i$ to give the total flux, $F$, (or intensity) over the frequency sub-interval:

$$F = \sum_{i=1}^{n} a_i F_i$$  \hspace{1cm} (5)

Tables of the exponential-sum coefficients, $a_i$ and $b_i$, can be obtained from the authors.

**NUMERICAL STUDIES.** The number of terms in the series of exponentials (n in equation 1) and the number of sub-intervals into which the spectral band is broken can be varied to increase the desired accuracy. Figure 3 shows a comparison between the 15 $\mu$m cooling rates computed using various numbers of frequency sub-intervals and terms within each sub-interval. The temperature of the atmospheric model rose linearly from 150K at the surface to 160K at 10km altitude, and then fell linearly to 130K at 40km altitude. These temperatures are typical of the winter polar atmosphere on Mars. Using 4 terms and 4 sub-intervals results in an error of less than 10% in the lowest 10 km, but results in substantial errors at higher altitudes. The explanation is that the 4 term fit could not capture the effects of both the band center and the wings at all altitudes, and in this case we emphasized the wings, which are more important at lower altitudes where the band center is saturated (see Fig. 1). Adding additional terms from 8 to 16 is far less noticeable.

**Figure 3.** 15 $\mu$m band cooling rates calculated with an exponential-sum approach using 4 frequency sub-intervals and 4 terms in the sum, 10 sub-intervals and 8 terms, and 10 sub-intervals and 16 terms. A winter polar temperature profile is used (see text), but only 200K exponential-sum coefficients are used for comparison purposes.
We also separately derived exponential-sum coefficients based on the Gal'tsev and Osipov [25] line-by-line calculations of the 15 μm band. The number of terms in the fit (n in equation 1) were varied, to confirm our FASCOD work on the number of terms required to give an accurate fit. Figure 4 shows cooling rates computed with several fits to the Gal'tsev and Osipov parameterizations, as well as to the FASCOD transmittances. Cooling rates compare favorably in the lower atmosphere (below 10 km altitude). Gal'tsev and Osipov only considered temperatures as cold as 200K. We extrapolated the temperature dependence to colder temperatures [26]. Discrepancies between the Gal'tsev and Osipov and FASCOD cooling rates are due to the inaccuracy in the temperature extrapolation, particularly at the higher, colder altitudes. Again, note that the 4 and 5 term fits become inaccurate above 10 km altitude, as in Fig. 3. Also, the 15 term fit yields no marked improvement over the 8 term fit, as also shown with the FASCOD fits in Fig. 3.

Cooling rates computed using the FASCOD exponential-sum transmittances also compared well below 10 km altitude with cooling rates computed using the Pollack et al. [27,4] parameterizations of CO₂ transmittance. At higher altitudes, discrepancies exist due to the use of the strong-line approximation by Pollack et al, which emphasizes the effects of the line wings. However, comparisons of our exponential-sum transmittances with techniques commonly used for the terrestrial atmosphere have indicated that our exponential-sum transmittances could be in error above 20 km altitude. J. Pollack is seeking to modify our exponential-sum approach to use two sets of sums, one applied to the line centers and one applied to the line wings. Initial tests seem to show better agreement in the upper atmosphere of Mars.

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SOLAR AND IR RADIATION NEAR THE MARTIAN SURFACE: A PARAMETERIZATION FOR CO₂ TRANSMITTANCE, Barnard Lee Lindner, AER Inc., 860 Memorial Drive, Cambridge, MA 02139; Thomas P. Ackerman, Dept. of Meteor., Penn. State Univ., University Park, PA 16802; James B. Pollack and O. Brian Toon, NASA/ARC, Moffett Field, CA 94035; Gary E. Thomas, APAS Dept., Univ. of Colorado, Boulder, CO 80309

INTRODUCTION. CO₂ comprises 95% of the atmospheric composition of the martian atmosphere [1]. CO₂ is an active absorber and emitter in near-IR and IR wavelengths; the near-IR absorption bands of CO₂ provide significant heating of the atmosphere, and the 15 µm band provides rapid cooling [2-9]. However, the Mars atmosphere also has a high aerosol content. Dust opacities vary from less than 0.2 to greater than 3.0, primarily on a seasonal basis with the occurrence of global dust storms during southern spring [6]. Ice-cloud opacities vary from 0 to greater than 1, with large amounts occurring at winter polar latitudes [7]. Dust and ice-cloud aerosols have high scattering albedos in solar wavelengths, and are highly absorbing at infrared wavelengths, and are as important as CO₂ in the atmospheric energy budget [5].

Including both CO₂ and aerosol radiative transfer simultaneously in a model is difficult. Aerosol radiative transfer requires a multiple-scattering code, while CO₂ radiative transfer must deal with complex wavelength structure. The problem can be solved exactly by inserting the CO₂ absorbance for each spectral line into a multiple-scattering code, but the 15µm band alone has on the order of 10,000 lines, making such a computation tedious and expensive. It is this difficulty of simultaneously treating aerosol multiple scattering and the banded absorption structure of CO₂ that prompts most radiative-transfer studies of the martian atmosphere to consider either a pure-CO₂ or pure-dust atmosphere. This approximation simplifies treatment, but is inaccurate if the atmosphere has been developed for atmospheric applications or the exponential-sum or k-distribution approximation [8-21]. The transmission of a homogeneous atmosphere is actually independent of the ordering of the absorption coefficient, k, in frequency space within a spectral interval, depending only upon the percentage of the spectral interval that has a particular value of k. The percentage of the spectral interval which has values between k and k + Δk can be formulated in a probability density function f(k) shown schematically in Figure 1. The chief advantage of the exponential-sum approach is that the integration over k space of f(k) can be computed more quickly than the integration of k over frequency. The exponential-sum approach is superior to the photon-path distribution and emissivity schemes for dusty conditions [22, 19, 23]. Our work is the first application of the exponential-sum approach to martian conditions.

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RESULTS. The number of terms in the series of exponentials and the number of sub-intervals into which the spectral band is broken can vary to increase the desired accuracy. Figure 2 shows a comparison between the FASCOD cooling rates computed using various numbers of frequency sub-intervals and terms within each sub-interval. The temperature of the atmospheric model rose linearly from 1500K at the surface to 1600K at 10 km altitude, and then fell linearly to 1300K at 40 km altitude. These temperatures are typical of the winter polar atmosphere. Using 4 terms and sub-intervals results in an error of less than 1% in the lowest 10 km, but results in substantial errors at higher altitudes. The explanation is that the 4 term fit could not capture the effects of both the band center and the wings at all altitudes, and in this case we emphasized the wings, which are more important at lower altitudes where the band center is saturated. Adding additional terms from 8 to 16 is far less noticeable. However, very recent work by J. Pollack comparing these exponential-sum transmittances with techniques commonly used for the terrestrial atmosphere have indicated that our exponential-sum transmittances could be in error above 40 km altitude. J. Pollack is seeking to apply our exponential-sum approach to use two sets of sums, one applied to the line centers and one applied to the line wings. Initial tests seem to show better agreement in the upper atmosphere of Mars.

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CO₂ TRANSMITTANCE IN THE MARTIAN ATMOSPHERE: AN EXPONENTIAL-SUM FIT FOR USE IN MULTIPLE-SCATTERING MODELS

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CO₂ and airborne dust are both important in heating and cooling the martian atmosphere. However, theoretical modeling of both CO₂ and dust radiative transfer simultaneously is difficult because dust radiative transfer requires a multiple scattering code, while CO₂ radiative transfer must deal with complex wavelength structure.

We have approximated the CO₂ transmittance at near-IR and IR wavelengths in the lower martian atmosphere with an exponential-sum fit. Exponential-sum fitting allows for the incorporation of CO₂ absorption and emission in a multiple scattering computer model in a straightforward, efficient and accurate manner. This makes it possible for CO₂ and dust heating and cooling of the martian atmosphere to be easily treated simultaneously.

Comparison of our CO₂ cooling and heating rates to those derived from several other techniques shows good agreement in the lower atmosphere below 20 km altitude, particularly near the surface. An improved version of the exponential-sum approach is being developed to treat higher altitudes as well.

Tables of exponential-sum coefficients are available from the authors.
MARS SEASONAL CO$_2$-ICE LIFETIMES AND THE ANGULAR DEPENDENCE OF ALBEDO

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The albedo of the polar caps on Mars brightens appreciably at high solar zenith angle (Warren et al., J. Geophys. Res., 95, 14717, 1990), an effect not included in prior polar-cap energy-balance models. This decreases absorption of sunlight by the polar cap, hence decreasing sublimation of CO$_2$ ice. Lindner (J. Geophys. Res., 95, 1367, 1990) has shown that the radiative effects of clouds and airborne dust will increase sublimation of CO$_2$ ice over that predicted by prior polar-cap energy-balance models. Furthermore, observations hint that more clouds may exist in the northern hemisphere, which Lindner (1990) has shown would sublime CO$_2$ ice more quickly in the north than in the south. I show here that the effects of the solar zenith angle dependence of albedo and the radiative effects of clouds and dust offset each other, but act to extend the lifetime of CO$_2$ ice on the south pole more than on the north pole, possibly explaining the observed hemispherical asymmetry in the residual polar caps without the need of a hemispherical asymmetry in polar-cap albedo required by prior models. Another positive aspect of this solution is that neither the inclusion of the solar zenith angle dependence of albedo nor the radiative effects of clouds and dust should appreciably change prior model agreement with observations of the annual cycle of surface pressure and the recession of the polar caps equatorward of 75° latitude.
WHY IS THE NORTH POLAR CAP ON MARS DIFFERENT THAN THE SOUTH POLAR CAP?

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Introduction. One of the most puzzling mysteries about the planet Mars is the hemispherical asymmetry in the polar caps. Every spring the seasonal polar cap of CO$_2$ recedes until the end of summer, when only a small part, the residual polar cap, remains. During the year that Viking observed Mars, the residual polar cap was composed of water ice in the northern hemisphere [Kieffer et al., Science, 194, 1341, 1976] but was primarily carbon dioxide ice in the southern hemisphere [Kieffer, J. Geophys. Res., 84, 8263, 1979]. Scientists have sought to explain this asymmetry by modeling observations of the latitudinal recession of the polar cap and seasonal variations in atmospheric pressure (since the seasonal polar caps are primarily frozen atmosphere, they are directly related to changes in atmospheric mass). These models reproduce most aspects of the observed annual variation in atmospheric pressure fairly accurately. Furthermore, the predicted latitudinal recession of the northern polar cap in the spring agrees well with observations, including the fact that the CO$_2$ ice is predicted to completely sublime away. However, these models all predict that the carbon dioxide ice will also sublime away during the summer in the southern hemisphere, unlike what is observed. This paper will show how the radiative effects of ozone, clouds, and airborne dust, light penetration into and through the polar cap, and the dependence of albedo on solar zenith angle affect CO$_2$ ice formation and sublimation, and how they help explain the hemispherical asymmetry in the residual polar caps. These effects have not been studied with prior polar cap models.

Ozone, Clouds, and Airborne Dust. Since O$_3$ is more prevalent in the northern hemisphere than in the southern hemisphere, O$_3$ was suggested as a cause for the hemispherical asymmetry in the residual polar caps by Kuhn et al. (J. Geophys. Res., 84, 8341, 1979). However, Lindner (submitted to Icarus, 1991) has shown that O$_3$ has a minor effect on the atmospheric temperature, and hence on the infrared radiation which strikes the polar cap, and Lindner (J. Geophys. Res., 95, 1367, 1990) has shown that O$_3$ absorbs less than 1% of the total solar radiation absorbed by the polar cap. Thus, O$_3$ is not an important consideration in the polar cap energy budget.

Lindner (1990) has computed the solar and thermal flux striking the polar cap of Mars for various ozone, dust, and cloud abundances and for three solar zenith angles. These calculations have been inserted in the polar-cap models...
of Lindner (Eos Trans. AGU, 67, 1078, 1986) and Jakosky and Haberle (J. Geophys. Res., 95, 1359, 1990). Vertical optical depths of dust and cloud ranging from zero to 1 cause little change in the total flux absorbed by the polar cap near its edge but increase the absorbed flux significantly as one travels poleward. Observed hemispherical asymmetries in dust abundance, cloud cover, and surface pressure combine to cause a significant hemispherical asymmetry in the total flux absorbed by the residual polar caps, which helps to explain the dichotomy in the residual polar caps on Mars.

**Light Penetration.** Penetration of solar radiation into the cap itself is included in my polar cap model, based on the theoretical work of Clow (Icarus, 72, 95, 1987). I find that the inclusion of light penetration slightly decreases the albedo needed in the model to keep CO$_2$-ice year-round at the south pole by on the order of 1%. The required albedo is decreased because some solar radiation is used to heat the subsurface, and not all of this heat is transported back to the surface. Overall, I conclude that penetration of light into the polar cap has only a small effect on the polar cap energy budget.

**Albedo and the Solar Zenith Angle.** Warren et al. (J. Geophys. Res., 95, 14717, 1990) has computed the dependence of the albedo of the martian polar caps on solar zenith angle, and these calculations have been included in my polar cap model. Since the albedo of ice increases and becomes more forward scattering at higher solar zenith angles, and since the solar zenith angle becomes higher as one approaches the pole, the albedo is greatest at the pole. This decreases absorption of sunlight, hence increasing survivability of CO$_2$ ice. In fact, this increases the survivability of ice enough to offset the decrease in survivability of ice due to the radiative effects of clouds and dust.

**Discussion.** The combination of the effects of solar zenith angle on albedo and the radiative effects of clouds and dust act to extend the lifetime of CO$_2$ ice on the south pole relatively more than on the north pole, explaining the hemispherical asymmetry in the residual polar caps without the need of a hemispherical asymmetry in polar cap albedo. Another positive aspect this solution is that neither the inclusion of solar zenith angle effects on ice albedo nor the radiative effects of clouds and dust should appreciably change model predictions of the annual cycle of pressure or polar cap recession equatorward of 75° latitude, since approximately 90% of the seasonal CO$_2$ frost is equatorward of 80° latitude. Hence, the good model agreement noted by prior researchers to the seasonal cycle in atmospheric pressure and to the recession of the polar cap equatorward of 80° latitude is retained.