

Polar stratospheric cloud impacts on Antarctic stratospheric heating rates

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SUMMARY

The impact of polar stratospheric clouds (PSCs) on the stratospheric radiative heating rate is analysed by including a nominal PSC in heating-rate calculations that incorporate realistic atmospheric variables including tropospheric clouds. The use of realistic atmospheric conditions constrains the possible radiative effects of PSCs, which previous studies have shown to be very sensitive to such variables as temperature, tropospheric clouds, and solar zenith angle. Over the pole, winter heating rates within the stratospheric polar vortex are decreased substantially by the presence of a PSC, while a PSC increases the heating rates equatorward of 75–85°S. Although the PSC always increases the short-wave heating, the effect in the long-wave region depends on the ground temperature, the stratospheric temperature, and presence of a tropospheric cloud. For the thickest PSCs (Type II), the effect in August 1994 varies from cooling by 0.25 K d^{-1} (potential-temperature difference $\Delta\dot{\theta} = 0.5 \text{ K d}^{-1}$) at the pole to heating by 0.3 K d^{-1} ($\Delta\dot{\theta} = 0.6 \text{ K d}^{-1}$) at 65°S to slight cooling equatorward of 57°S. September 1994 results are similar. Calculated heating rates over the pole including PSCs are near -0.5 K d^{-1} ($\dot{\theta} = -1 \text{ K d}^{-1}$) for both months, and positive heating rates of up to 0.25 K d^{-1} ($\dot{\theta} = 0.5 \text{ K d}^{-1}$) occur near the vortex edge. Thinner PSCs (Type I) have less of an effect; for example, heating rates of 0.375 K d^{-1} ($\dot{\theta} = 0.75 \text{ K d}^{-1}$) occur over the pole in August when a Type I PSC is included. These results should be viewed as an upper bound to the effect of PSCs since the calculations specify 100% PSC cover; satellite results show that this assumption is not unreasonable within the vortex during winter and early spring, however. The increased latitudinal gradient in descent rates in the presence of a PSC is consistent with the behaviour of long-lived trace-gas observations, and strengthens the vortex relative to a PSC-free case.

KEYWORDS: Heating rates Polar stratospheric clouds

1. INTRODUCTION

Descent rates are an important component of the dynamics of the winter polar stratospheric vortex. Together with horizontal mixing, the descent rate governs the behaviour of trace gases, and so knowledge of the descent rate is essential for understanding the distribution of chemical species in the stratosphere. Descent rates have been computed in several ways, with a wide range of results. Large descent rates support the notion that ozone-depleted air can move to the subvortex region where mixing to lower latitudes occurs at a relatively rapid rate. Small descent rates indicate that this effect is not substantial. Analysis of the behaviour of trace gases has been used by a variety of studies. Some studies computed descent rates assuming no horizontal mixing, and produced very low values (e.g. Schoeberl *et al.* 1992). Other studies of trace gases using different techniques, including correlations between different species, produced much larger descent rates (expressed as rate of change of temperature $\dot{T} = 0.5\text{--}1.5 \text{ K d}^{-1}$; all heating rates quoted are in temperature per day; $\dot{\theta}$ describes the vertical velocity in potential-temperature coordinates, and $\dot{\theta} \simeq 2\dot{T}$ at lower-stratospheric altitudes; Tuck 1989; Proffitt *et al.* 1989). White and Bryson (1966) published cooling rates greater than 1 K d^{-1} based on radiometersonde observations in Antarctica. Radiative transfer calculations are yet another method of computing descent rates. Previous studies of the radiative heating rates reported lower values of the descent rates (e.g. Rosenfield *et al.* 1994; Hicke and Tuck 1999) on the order of 0.3 K d^{-1} .

The latitudinal gradient of descent rates within the vortex in the lower stratosphere is also a matter of current research. Schoeberl *et al.* (1992) proposed that the maximum

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descent occurs at the vortex edge. In contrast, Danielsen and Houben (1988), Tuck and Proffitt (1997), and Hicke and Tuck (1999) discuss the maximum descent occurring at the pole. The shape of the latitudinal descent rate gradient has implications for interpreting the behaviour of trace gases in the lower stratosphere.

In the austral winter, temperatures within the Antarctic stratospheric vortex decrease to values low enough to form polar stratospheric clouds (PSCs). PSCs are well known for their role in ozone depletion (Solomon 1999 and references therein). In addition to playing a role in ozone depletion, PSCs also affect the radiation and dynamics of the stratosphere, and influence the amount of descent that occurs in the vortex. Little has been quantified regarding the effect of PSCs on the descent rates, however. Several studies have shown that PSCs could have a substantial effect on the radiative heating rate (Blanchet 1985; Pollack and McKay 1985; Kinne and Toon 1990). These studies showed that changes between -0.5 and $+1$ K d $^{-1}$ could occur when PSCs were included. The magnitude and sign of the changes vary substantially within and among studies, depending on many factors relating to both the PSC (e.g. optical thickness) as well as the environment (e.g. the amount of solar radiation, or the presence or absence of tropospheric clouds). In particular, Kinne and Toon (1990) show large changes in the PSC impacts when a tropospheric cloud was included in the calculations.

To investigate the impact of PSCs on the radiative heating rate over month-long time periods, Rosenfield (1992) modelled PSC composition and used temperatures to estimate the location and thickness of PSCs. She found that PSCs make only a small difference in month-long time periods. Rosenfield (1992) explored the effect of tropospheric clouds only as a sensitivity study, however, with highly parametrized tropospheric clouds, and the sign of the PSC impact depended on the specification of tropospheric clouds.

Hemispheric observations of PSC composition and particle size throughout the winter and spring are needed to completely assess the impacts of PSCs. However, satellites currently lack such capability. PSCs are optically very thin, making detection by nadir-viewing satellites difficult. Limb-sounding satellites, on the other hand, integrate the observed extinction over a large spatial area (200 km 2), and thus they are unable to distinguish between a thick PSC with limited horizontal extent and a thin PSC spread over a wide area. Thus, improvements to the PSC modelling by Rosenfield (1992) as regards the radiative heating rate await the advent of new developments in PSC detection.

Recently, Hicke and Tuck (1999) published radiative heating-rate calculations that included the realistic representation of tropospheric clouds through the outgoing long-wave radiation (OLR)-matching technique. The present study uses the OLR-matching technique to investigate the sensitivity of PSCs on the radiative heating rate. By accurately including the effects of tropospheric clouds, this study improves on previous studies to better understand the role of PSCs given realistic atmospheric conditions. However, this study only includes a nominal PSC with fixed characteristics; the actual effect of PSCs depends on the (currently unavailable) PSC distribution and composition. We note, however, that airborne lidar studies from the DC-8 aircraft over Antarctica in August and September 1987 recorded dense PSCs all the way from the edge to the centre of the vortex in late August to mid September (Murphy *et al.* 1989; Tuck 1989; Watterson and Tuck 1989; McKenna *et al.* 1989).

2. HEATING-RATE CALCULATIONS

Data used in this study were the monthly, zonal mean quantities of: temperature from the UKMO/UARS assimilations (Swinbank and O'Neill 1994); and water vapour

and ozone from the Halogen Occultation Experiment (HALOE) and Microwave Limb Sounder (MLS) instruments as generated by Rosenlof (1999). The primary trace gas source used was observations from HALOE; MLS observations, adjusted to match HALOE values, were used to fill in the parts of the atmosphere that HALOE did not observe. The dataset includes an ozone hole. Latitudes not covered by either instrument, usually within 5° of the pole, were filled in using the most poleward data. The climatology of Logan (1999) provided tropospheric ozone values, while tropospheric water vapour from the European Centre for Medium-Range Weather Forecasts (Trenberth 1988) was used. The vertical extent of the data is from the surface to 0.3 hPa, with a vertical resolution of approximately 2.5 km. For this study monthly mean inputs were used to compute the monthly mean heating rates; a number of studies have shown the close agreement between these types of calculations and mean heating rates computed by averaging daily heating rates over a month (Gille and Lyjak 1986; Rosenfield *et al.* 1987; Hicke and Tuck 1999).

Zonal mean temperatures for August and September 1994 are shown in Fig. 1. Temperatures below the PSC thresholds for Type I (~ 196 K) and Type II* (~ 192 K) can be seen poleward of about 65° S in August and 68° S in September. The zonal mean temperatures equatorward of this are too warm to support PSCs, but the minimum zonal temperatures are cold enough to support PSCs for an additional 5° equatorward (that is, poleward of about 60° S). Thus, heating rates including PSCs equatorward of about 60° S should be viewed as hypothetical. Stratospheric temperatures during August and September 1994 were typical of the middle 1990s: values were within 1 K of the 1992–1998 mean temperatures.

Short-wave heating rates were computed using the Column Radiation Model of the Community Climate Model 3 (CCM3/CRM) developed by the National Center for Atmospheric Research (NCAR; Hack *et al.* 1993). The plane-parallel assumption in CCM3/CRM underestimates the solar heating in the lower stratosphere by 0.01 to 0.2 K d $^{-1}$ (2 to 20%; Lary and Balluch 1993). The Rapid Radiative Transfer Model (RRTM; Mlawer *et al.* 1996) was used to compute the long-wave heating rate owing to its more detailed treatment of clouds. Though RRTM ignores scattering, calculations with another radiative transfer model that does include scattering, Streamer (Key 1996; Pinto *et al.* 1997), indicated that scattering in the long-wave region plays only a minor role in computing the effect of PSCs on the heating rate. Comparisons between these models and MODTRAN (Bernstein *et al.* 1996), a narrow-band model with cm $^{-1}$ spectral resolution, showed good agreement (within 0.05 K d $^{-1}$) for both long- and short-wave heating rates (Hicke *et al.* 1999).

The radiative impact of PSCs depends heavily on the upwelling long-wave radiation from the troposphere, consequently accurate treatments of ground temperature and tropospheric clouds were incorporated. Ground temperatures from the National Centers for Environmental Prediction (NCEP) Reanalyses (Kalnay *et al.* 1996) were used; 1994 was within 2 K of 40-year mean temperatures. The OLR-matching technique developed by Hicke and Tuck (1999) was used to account for tropospheric clouds. The OLR-matching technique uses collocated satellite observations of OLR to adjust a nominal tropospheric cloud in the profile until the OLR computed by the radiative transfer model matches the collocated satellite OLR. While the algorithm does not necessarily retrieve accurate tropospheric cloud information, it does accurately predict the upwelling long-wave radiation from the troposphere into the stratosphere, and that is the important effect

* Type I PSCs consist of nitric acid trihydrate while Type II PSCs consist of water ice. Type I PSCs are generally optically thinner than Type II PSCs.

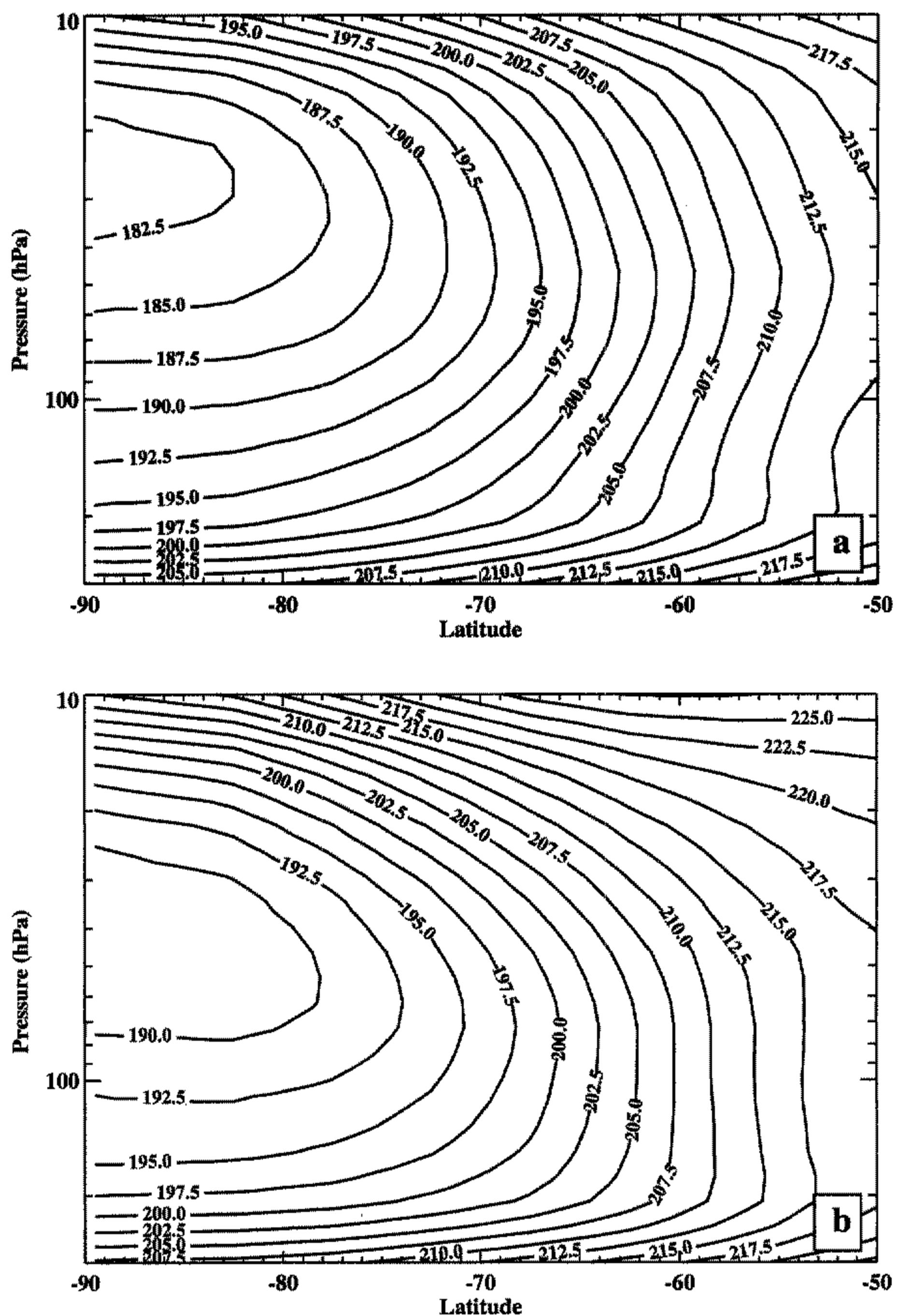


Figure 1. Zonal mean temperatures (K) for (a) August 1994, and (b) September 1994.

of tropospheric clouds on the stratospheric heating rate; see Hicke and Tuck (1999) for more details. The OLR-matching technique was updated to include zonally averaged cloud amounts from the International Satellite Cloud Climatology Project (ISCCP; Rossow and Schiffer 1991); this results in a more accurate computation of the short-wave heating.

Following Kinne and Toon (1990), a nominal Type I or Type II PSC was inserted into the profile at each latitude between 90° and 50°S for the months of August and September 1994. The PSC was placed between 80 and 100 hPa and was approximately 2 km thick, with a cloud fraction of 1.0. Unpublished Polar Ozone and Aerosol Measurement (POAM) data from M. Fromm (personal communication) reveal PSC observation frequencies within the Antarctic vortex of 85–98% in August and 70–99% in September. These results are from 1994 through 2000 using the algorithm of Fromm *et al.* (1999). Poole and Pitts (1994) discussed somewhat lower results (60%) for Stratospheric Aerosol Measurement II (SAM II) satellite observations. Thus, the assumption of a cloud fraction/frequency of unity throughout the month is probably too high, particularly for the latitudes nearest to the equator where the temperatures are higher than at higher latitudes. On the other hand, airborne lidar data show a fraction of one in the inner vortex in late September at altitudes up to 19 km, particularly in the inner vortex; solar occultation limb scanners are necessarily restricted in their view of this region before equinox. However, this study was designed to demonstrate the possible effect of PSCs, not the actual effect, and so the results should be considered as an upper limit at the altitude of the PSC, although the actual values will be close to this upper limit.

Though PSCs were only placed between 80 and 100 hPa (approximately 16–18 km) for this study, PSCs observed from satellite (Poole and Pitts 1994; Fromm *et al.* 1997) occurred at altitudes from 13 to 28 km. Furthermore, dehydration has been observed up to 25 km (approximately 30 hPa; Vömel *et al.* 1995), implying that PSCs form at those altitudes. PSCs thus occur at altitudes throughout the lower stratosphere.

Mie calculations provided the single scattering parameters using a log normal distribution of particle sizes (mode radius = $8\text{ }\mu\text{m}$, geometric standard deviation = $1.25\text{ }\mu\text{m}$). The number concentration was adjusted so that the PSC optical depth, τ , at $0.55\text{ }\mu\text{m}$ equalled the Type I (0.01) and Type II (0.04) optical depths specified by Kinne and Toon (1990). An optical depth of 0.04 at $0.55\text{ }\mu\text{m}$ corresponds to an optical depth of 0.015 at $10\text{ }\mu\text{m}$. The volume extinction coefficient $\beta_{\text{ext}} = \tau/\Delta z = 0.02\text{ km}^{-1}$ for Type II PSCs is probably at the upper limit of PSCs. Blanchet (1985) and Pollack and McKay (1985) used SAM II values of 0.01 km^{-1} between roughly 11 and 13 km (from McCormick *et al.* 1982) in their sensitivity studies. Rosenfield (1992) reported modelled and SAM II extinction coefficients of 0.01–0.02 over 2–5 km in the lower stratosphere. Thus, the Type II results reported in this study should be considered an upper bound.

The RRTM allows investigation into the spectral long-wave effect of the PSCs. Figure 2 shows the heating-rate difference (*with* minus *without* a Type II PSC) at PSC altitude for latitudes 65.5° and 89.5°S in August. At wave numbers less than 500 cm^{-1} large decreases in the heating rate (increases in the cooling rate) occur when a PSC is included for both latitudes. Water vapour rotational lines dominate this spectral region, and the effect of adding a PSC is similar to increasing the water vapour concentration.

For the profile at 65.5°S , the window region between 800 and 1400 cm^{-1} is strongly affected by the presence of a PSC. In this spectral region, in the presence of a PSC the heating rate is influenced by the effective emitting temperature in the troposphere (that is, considering both the ground temperature and the tropospheric clouds) compared to

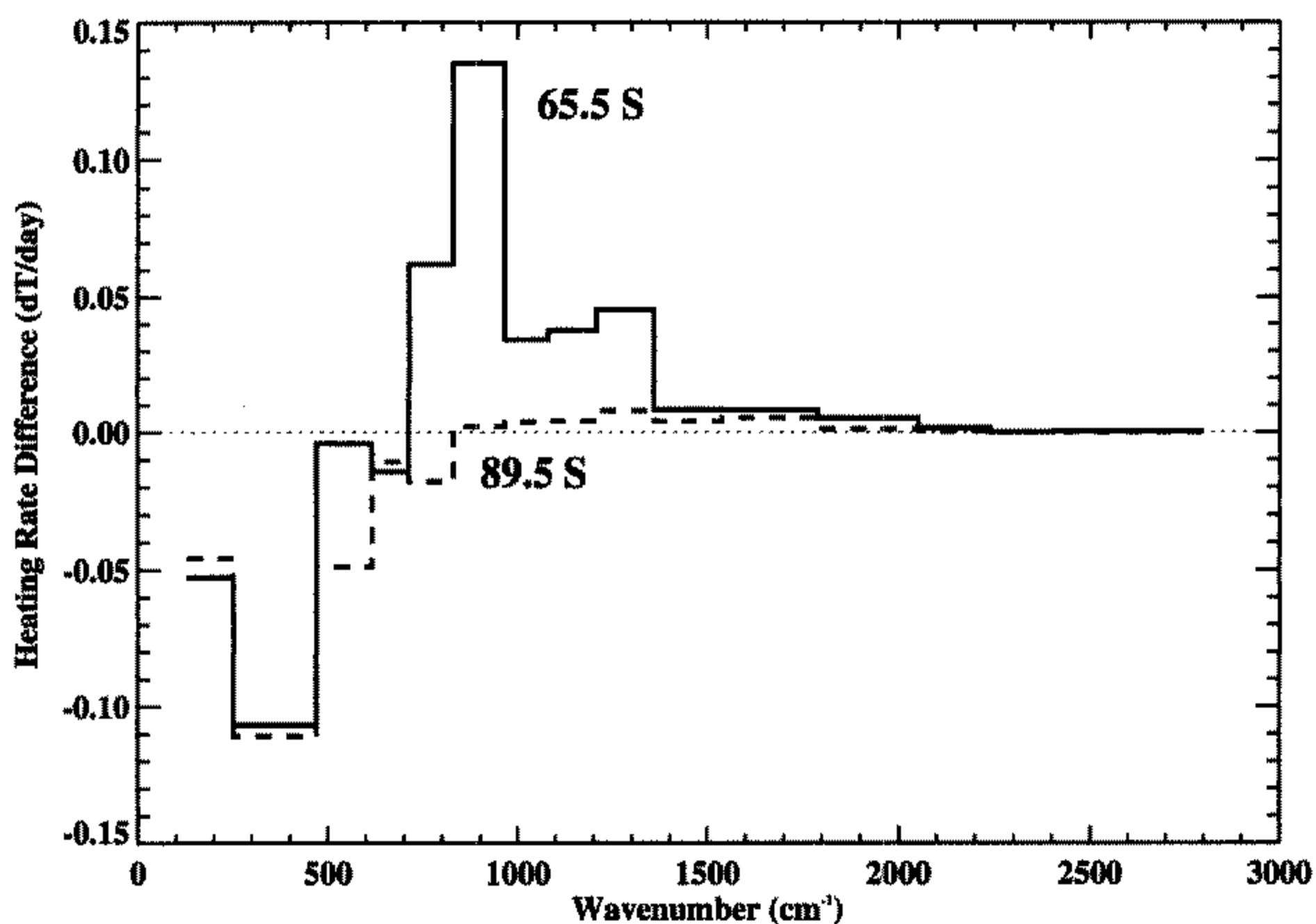


Figure 2. Heating-rate difference (K d^{-1}) for each Rapid Radiative Transfer Model (RRTM) long-wave band after inserting a Type II (see text) polar stratospheric cloud (PSC) into the profile (PSC minus no PSC) for August 1994. Differences at 65.5°S (solid curve) and 89.5°S (dashed curve) are shown; the altitude is the PSC level (approximately 80 hPa).

the PSC temperature. At 89.5°S , the heating between 800 and 1400 cm^{-1} as a result of a PSC is almost negligible owing to the extremely low ground temperatures over the interior of the Antarctic continent. In contrast, substantial differences exist for the profile at 65.5°S ; tropospheric clouds do occur (cloud fraction is 0.7), but both the underlying ground and the atmosphere are warmer than at the pole. Thus, the broad-band long-wave difference at 65.5°S is of opposite sign compared to that at 89.5°S as a result of this heating in the window region.

3. RESULTS

Figure 3 shows the heating rates at PSC altitude (approximately 80 hPa) for results without a PSC as well as with Type I or Type II PSCs for August 1994. Net heating rates (Fig. 3(a)) without a PSC are -0.25 K d^{-1} and are approximately constant from 50° to 90°S . The addition of a Type II PSC decreases the heating rate at the pole by 0.25 K d^{-1} to near -0.5 K d^{-1} . Equatorward of 73°S , the Type II PSC increases the heating, whilst between 60 and 70°S they increase the heating rate by up to 0.3 K d^{-1} to values exceeding 0 K d^{-1} . Type I PSCs have less of an effect: additional cooling at the pole is 0.05 K d^{-1} , while additional heating nearer the equator is about 0.1 K d^{-1} . In the zonal mean, temperatures equatorward of about 65 – 70°S are probably too warm to support PSCs, though they do sometimes occur at these latitudes.

Figure 3(b) reveals that PSCs both increase and decrease the long-wave heating rate for these realistic atmospheric conditions in August, depending on latitude. Long-wave cooling occurs poleward of 73°S and equatorward of 57°S with heating in between. Latitudinal variations in the upwelling long-wave radiation from the troposphere (controlled

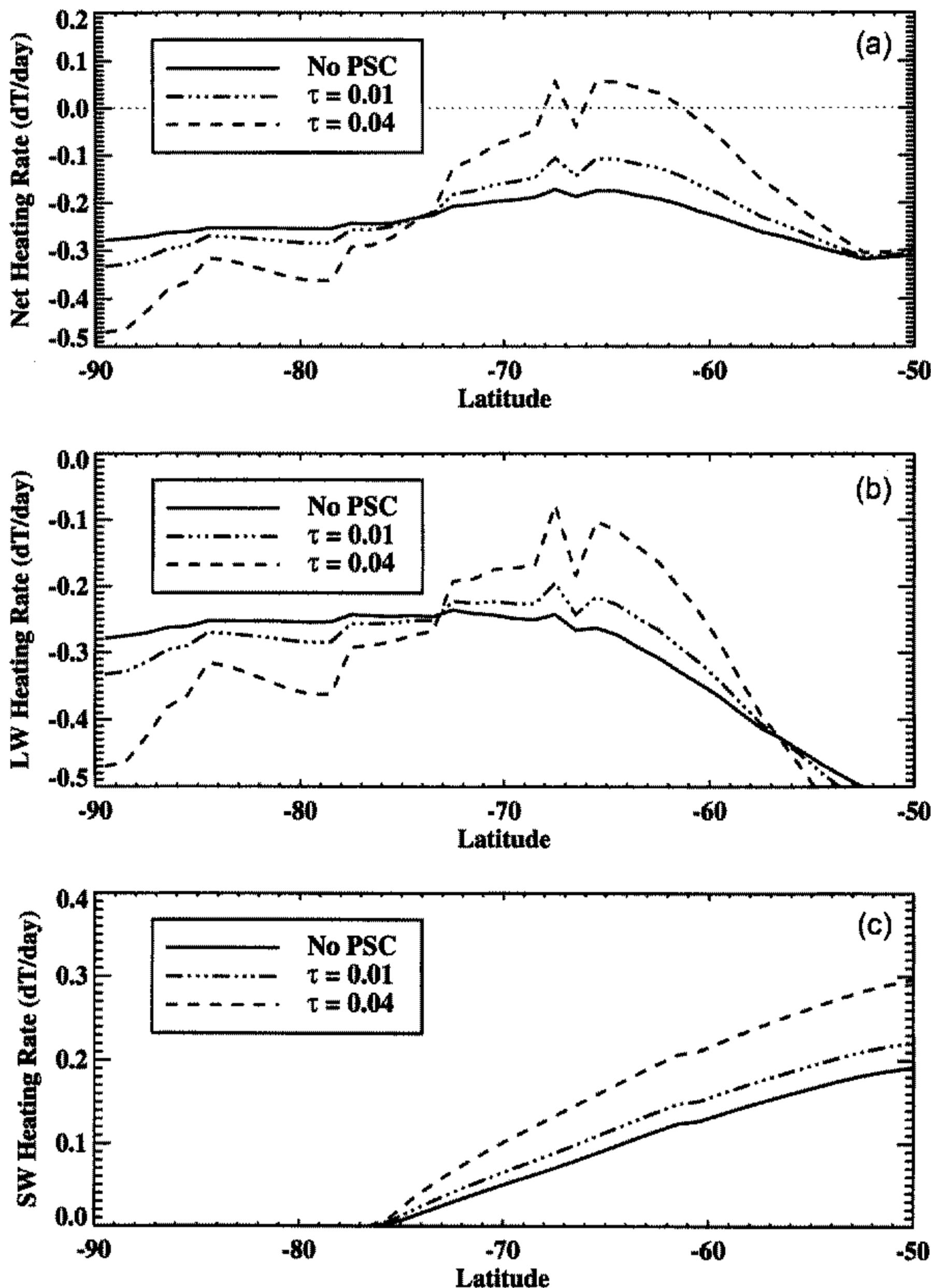


Figure 3. Monthly, zonally averaged (a) net, (b) long-wave, and (c) short-wave heating rates (K d^{-1} , \dot{T}) at polar stratospheric cloud (PSC) level (84 hPa) for August 1994. Heating rates are for profiles with a Type I PSC (optical thickness, $\tau = 0.01$; dot-dashed curve), with a Type II PSC ($\tau = 0.04$; dashed curve), and with no PSC (solid curve). The dotted line in (a) indicates $\dot{T} = 0$.

by the ground temperature and the presence of tropospheric clouds) and the stratospheric temperature are responsible for the changes from cooling to heating. In August, short-wave heating (Fig. 3(c)) is 0 K d^{-1} poleward of 77°S ; equatorward of this, Type II PSCs increase the heating rate by up to 0.1 K d^{-1} , while Type I PSCs increase it by 0.05 K d^{-1} .

September results (Fig. 4) show a similar pattern to August, with PSCs decreasing heating rates at the pole, and increasing the heating at other latitudes. Despite the greater

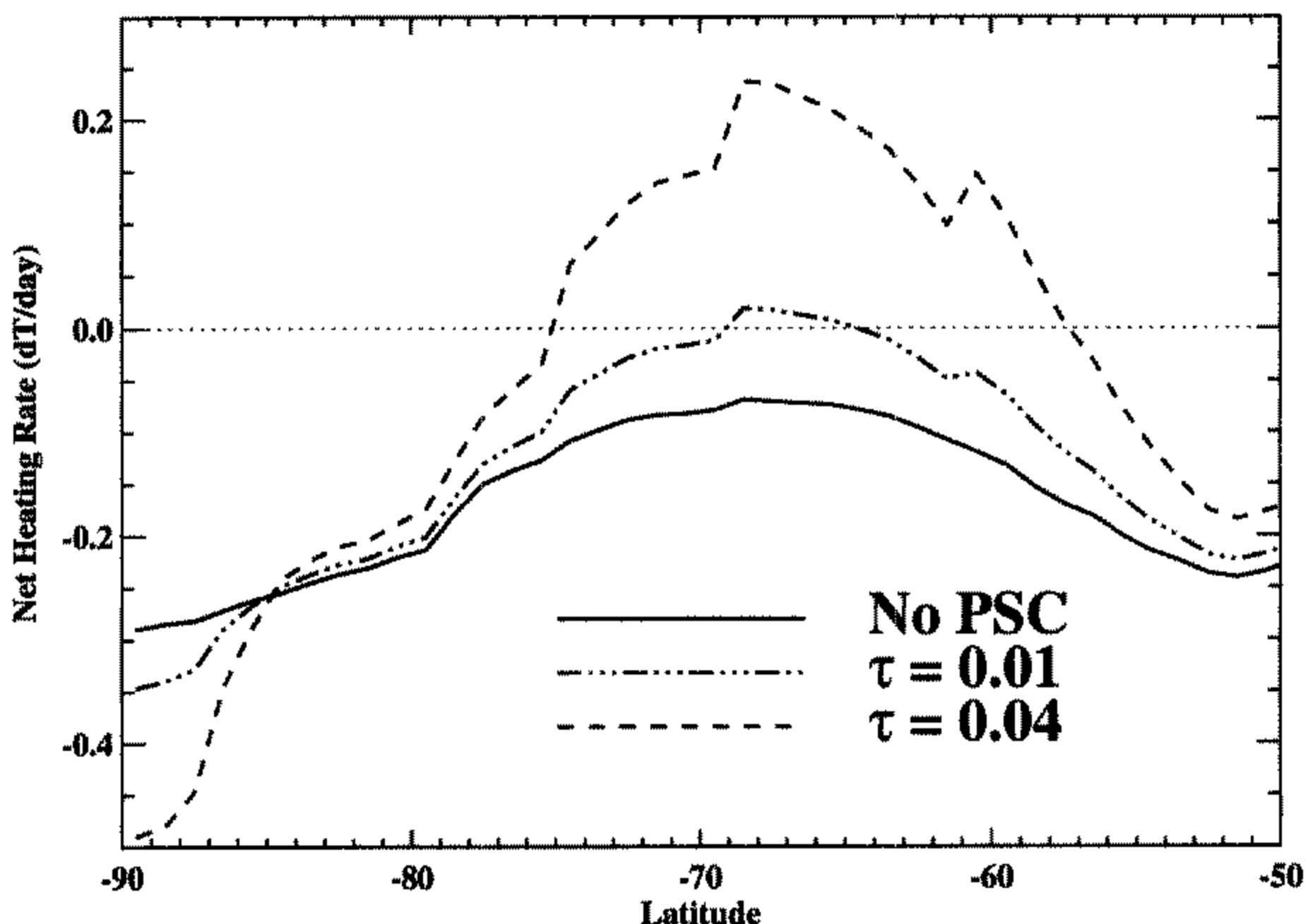


Figure 4. Same as Fig. 3(a) but for September 1994.

short-wave heating in September at most latitudes, heating rates with Type II PSCs are -0.5 K d^{-1} at the pole where short-wave heating remains 0 K d^{-1} . At 68°S the heating rates with PSCs approach 0.25 K d^{-1} . September and August effects for Type I PSCs are similar.

The results from both months show that PSCs have the effect of increasing the latitudinal gradient of the heating rates. Strong descent at the pole is enhanced in the presence of a PSC, while the maximum in heating rates (minimum in the descent rates) seen just inside the vortex edge (about 62°S) increases. The assumption of a cloud fraction of one is more realistic near the pole, and so the probable impact of PSCs is primarily to increase the descent at the pole, and only slightly increase the heating rate near the vortex edge since PSCs are not as frequent there.

This increase in the latitudinal gradient in the presence of PSCs drives the atmosphere away from the view of Schoeberl *et al.* (1992), who proposed that the maximum descent in a 'generic' vortex occurs at the vortex edge. The strong gradient is also more consistent with the behaviour of long-lived trace gases within the vortex. Tuck and Proffitt (1997) discussed poleward decreases of CH_4 (measured by HALOE) and N_2O (measured on board the ER-2) on an isentropic surface; Schoeberl *et al.* (1995) also showed similar results in the lower stratosphere using HALOE CH_4 . This sloping latitudinal gradient was predicted by Danielsen and Houben (1988). Both CH_4 and N_2O decrease with altitude in the stratosphere, and thus in the absence of horizontal mixing lower values of these tracers indicate stronger descent. It is known that while horizontal mixing occurs in the outer vortex and equatorward of the vortex edge, less horizontal mixing is present at the centre of the vortex owing to the lack of wave activity (Tuck *et al.* 1997; Bithell *et al.* 1994). Thus, the lower values of CH_4 and N_2O seen at the latitudes nearest the pole, where less mixing occurs, are indicative of stronger descent there. Increasing tracer values towards the vortex edge are consistent with the minimum

in descent rates computed in this study, if the effect of horizontal mixing is taken into account. In contrast, a maximum descent at the vortex edge, as postulated by Schoeberl *et al.* (1992), would result in a minimum in the tracer values at the vortex edge, or perhaps a decrease in the tracers from pole to lower latitudes. This slope, however, is opposite to that seen in the trace gas behaviour.

The presence of a minimum in descent rates just inside the vortex together with the lack of a maximum in the tracer gas plots suggests that there is significant horizontal mixing, even inside the vortex edge, that eliminates the maximum in trace gas values that would occur in the absence of horizontal mixing. For a given latitudinal gradient in a tracer, the amount of horizontal mixing depends on the degree of descent; more accurate calculations of the radiative heating rate including PSCs would therefore increase our understanding of horizontal mixing.

The increased latitudinal gradient of heating rates by PSCs has another effect related to the strength of the vortex. The stronger latitudinal gradient in heating rates shown by the calculations including a PSC will result in a stronger temperature gradient, which by the thermal-wind equation leads to a higher zonal wind. Thus, inclusion of PSCs when computing heating rates in simulations of Antarctic stratospheric dynamics will produce stronger zonal winds than similar simulations that do not include PSCs.

This effect is enhanced by ozone loss. Many studies, including both modelling (e.g. Kiehl *et al.* 1988) and observations (Randel and Wu 1999), have analysed the positive feedback that ozone reduction would have on the persistence of the vortex. Reduced short-wave heating resulting from ozone loss allows temperatures to remain cold longer and PSCs to persist later in spring. Therefore, prolonged low temperatures associated with ozone loss lead to more frequent PSC occurrence and continued enhancement of the heating-rate gradient.

Figure 5 plots the vertical profile of heating rates with and without a Type II PSC. As can be seen, the vertical extent of the PSC effect is limited to the layer where the PSC occurs, similar to the results of Kinne and Toon (1990). Only very small differences occur in the long-wave heating rate outside of this layer, and virtually no differences occur in the short-wave heating. Thus, the vertical effect of PSCs on the heating rate will be determined by where the PSCs are located in the profile. As discussed above, PSCs have been observed from 13–28 km, and so could have large radiative effects throughout the lower stratosphere. Note that if a PSC with the same optical depth were spread over a larger vertical extent, the heating-rate differences would decrease. As an aside, the existence of a PSC reduces the OLR by less than 5 W m^{-2} , affecting the heating-rate accuracy of the OLR-matching technique by less than 0.02 K d^{-1} .

By using the OLR-matching technique to accurately represent clouds, the effect of tropospheric clouds can be evaluated when assessing PSC impacts. Though the sensitivity of the heating rates in a stratosphere containing a PSC to tropospheric clouds is known (Blanchet 1985; Pollack and McKay 1985; Kinne and Toon 1990), the actual effect has not been quantified owing to a lack of accurate tropospheric cloud information. Hicke and Tuck (1999) used the OLR-matching technique to discuss the actual effect of tropospheric clouds on a clear stratosphere. Here we discuss the actual tropospheric cloud effect on a stratosphere containing a nominal Type II PSC. Figure 6 plots the short- and long-wave heating rates from August 1994 including a PSC both for clear tropospheric skies and for realistic tropospheric clouds. Little difference is seen in the short-wave heating, but large differences occur in the long-wave heating. Long-wave heating rates decrease (cooling rates increase) in the presence of PSCs and tropospheric clouds, compared to results with only PSCs; magnitudes range from 0.05 K d^{-1} at the pole to 0.5 K d^{-1} at 50°S . As discussed in Hicke and Tuck (1999), tropospheric clouds

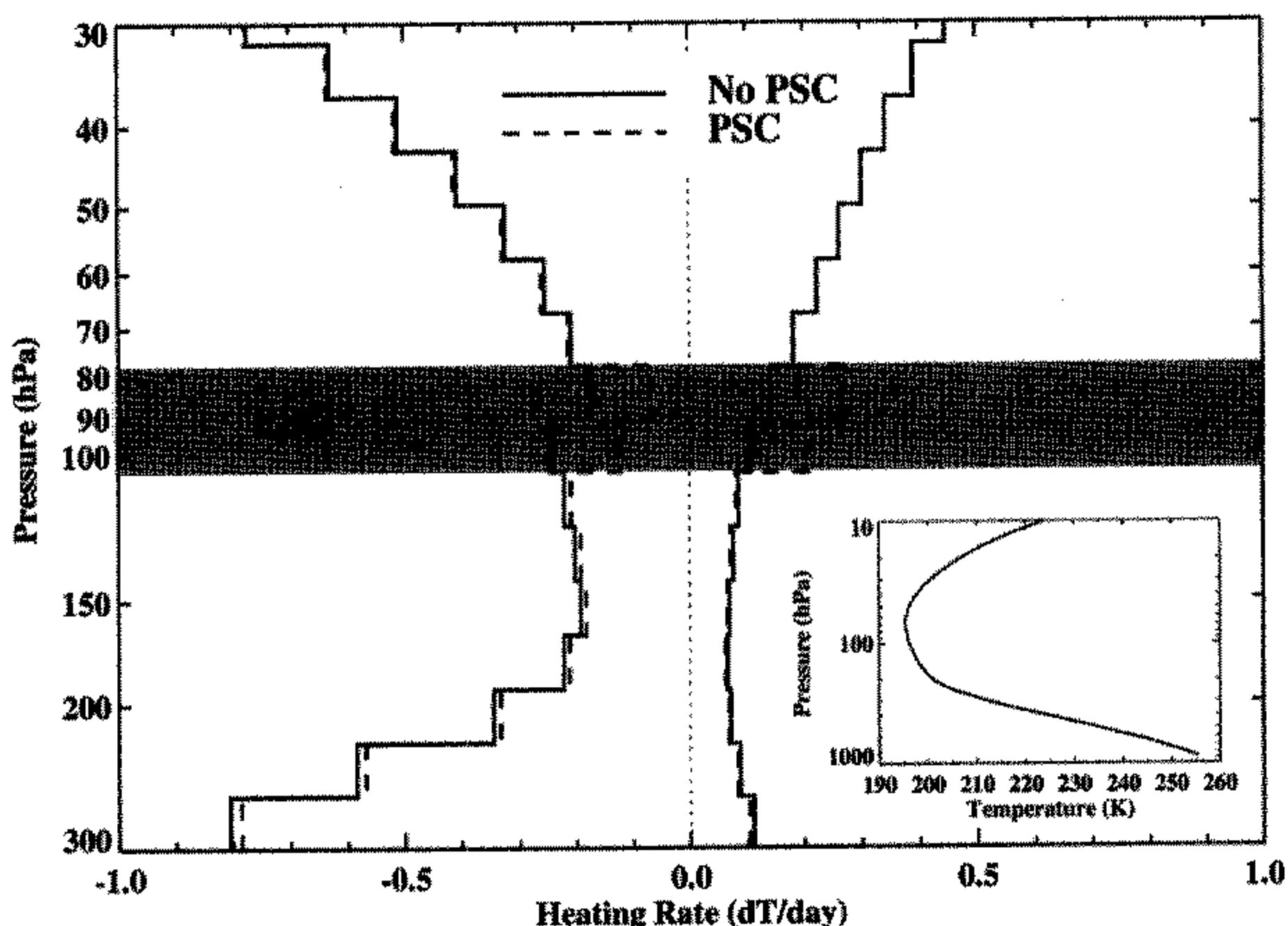


Figure 5. Heating rates for September 1994 at 70.5°S (K d^{-1} , \dot{T}) for a profile with (dashed curve) and without (solid curve) a Type II (see text) polar stratospheric cloud (PSC). Long-wave heating rates are on the left, and short-wave heating rates on the right. The curve inset shows the temperature profile at this location. Outgoing long-wave radiation for this profile is 164 W m^{-2} .

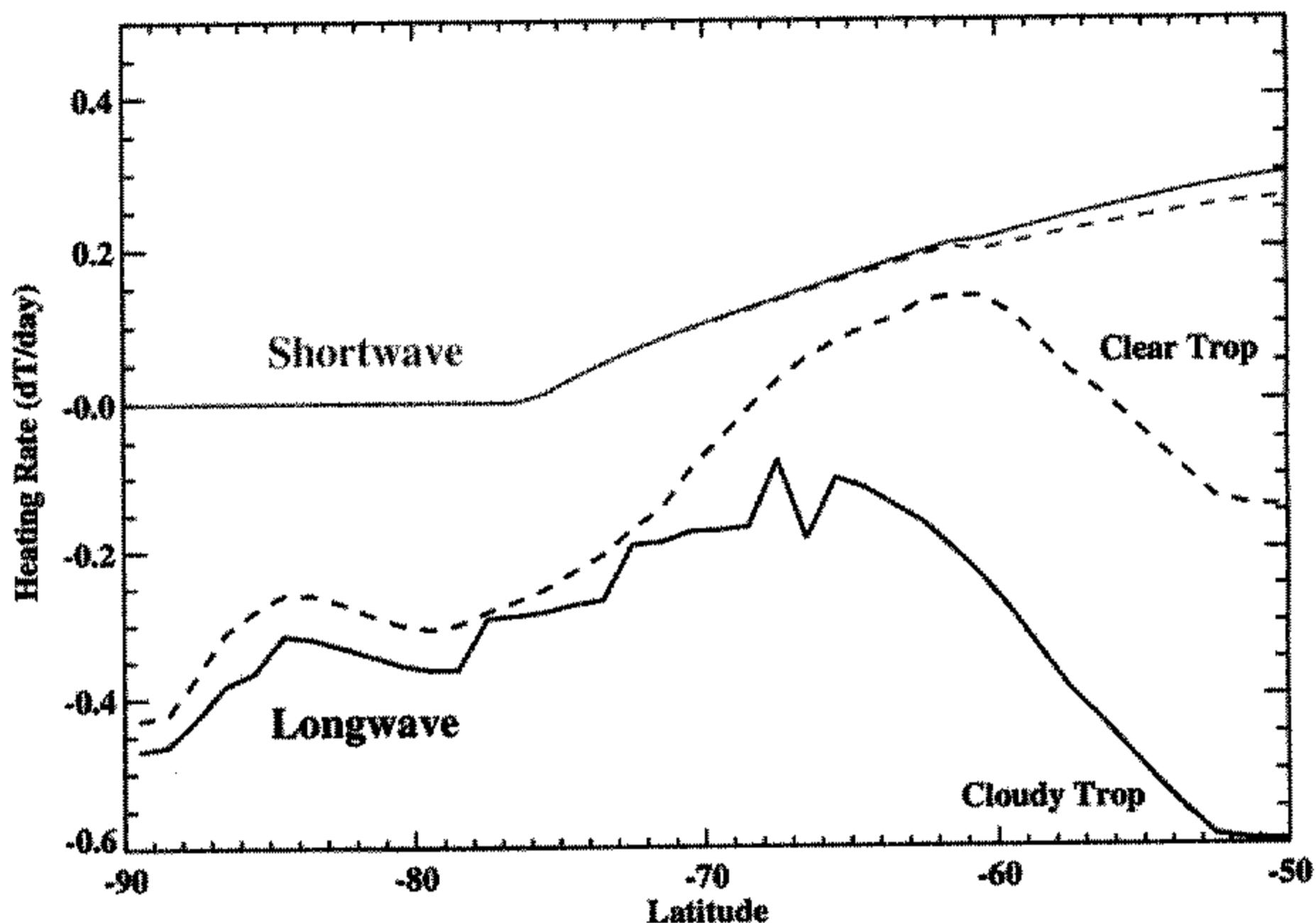


Figure 6. Monthly, zonally averaged short-wave (grey) and long-wave (black) heating rates (K d^{-1} , \dot{T}) for profiles with a Type II (see text) polar stratospheric cloud, with (solid curve) and without (dashed curve) tropospheric clouds for August 1994.

reduce the upwelling radiation into the stratosphere; the presence of a PSC means that the impact of tropospheric clouds is enhanced as a result of the additional absorbing spectral regions as discussed above.

The increasingly large effect of tropospheric clouds at the lower latitudes is a result of two factors. First, the tropospheric cloud fraction increases; little cloud occurs over the interior of the continent at the highest latitudes. Second, the effect of any cloud that is present increases since the ground temperature increases towards the equator, and by placing a relatively cold tropospheric cloud in the profile the upwelling long-wave radiation perturbation increases over a warm, open ocean.

The curve showing results without tropospheric clouds is much smoother than that with tropospheric clouds, and is a result of the increased sensitivity of stratospheric heating rates in the presence of a PSC to upwelling long-wave radiation from the troposphere. By inserting a PSC, the window spectral region can now contribute to the broad-band long-wave stratospheric heating rate, and so changes in ground temperature and cloudiness that result in changes to the upwelling long-wave radiation into the stratosphere will more strongly affect the heating rate. This can also be seen by comparing the curves in Fig. 3(b). Similar results occur for September 1994; Type I PSCs cause smaller impacts. Note that the zonal mean temperatures equatorward of about 60°S are too high to support PSC development.

4. CONCLUSIONS

Polar stratospheric clouds were incorporated into radiative heating-rate calculations that included observations of inputs such as temperature and realistic representations of tropospheric clouds (without resorting to climatologies). Because previous studies (e.g. Kinne and Toon 1990) found large uncertainties of the impact of PSCs based on the characteristics of the lower atmosphere, we eliminated those uncertainties in this study by accurately accounting for these characteristics. Nominal PSCs of visible optical depth 0.01 (Type I) and 0.04 (Type II) and cloud fractions of one were inserted into monthly, zonal mean fields for August and September 1994 in the southern hemisphere. Because the cloud fractions are somewhat too high, particularly in the latitudes of the vortex nearest to the equator where the lower-stratospheric temperatures are warmer, the results presented here should be viewed as an upper bound, but one that probably does not greatly exceed the actual values.

Large perturbations to the heating rate were found in this study, suggesting that PSCs could play an important role in determining the descent rate within the vortex. While heating rates including a Type II PSC of $\dot{T} = -0.5 \text{ K d}^{-1}$ ($\dot{\theta} = -1 \text{ K d}^{-1}$) seen in this study at the pole are similar to those results from previous studies published that report large descent rates (Tuck 1989; Proffitt *et al.* 1989), results for other latitudes suggest that the effect of PSCs is to slow the descent there. Thus, it appears that the addition of PSCs will not increase the radiative cooling rates in a vortex-averaged sense to the large values needed to support large mass flow to the subvortex region. It will, however, lead to a situation in which descent by radiative cooling can carry inner-vortex air down through the $\theta = 400 \text{ K}$ transition to altitudes where isentropic exchange with mid latitudes is freer, while 'peel-off' and the effects of chaotic advection erode the containment in the outer vortex (Bithell *et al.* 1994; Pierce *et al.* 1994; Rosenlof *et al.* 1997; Tuck and Proffitt 1997).

PSCs always increased the short-wave heating, but the long-wave effect was one of either heating or cooling depending on the location. Changes in the upwelling long-wave radiation, governed by the ground temperature and tropospheric clouds, as well

as changes in the stratospheric temperature, result in a latitudinal variation from strong cooling over the pole to strong heating near the vortex edge to cooling again outside the vortex. This study confirms the importance of accurately accounting for tropospheric clouds when assessing PSC impacts that has been seen in previous sensitivity studies (e.g. Kinne and Toon 1990; Rosenfield 1992).

The presence of PSCs strengthens the latitudinal gradient of radiative heating rates discussed by Hicke and Tuck (1999). The PSCs increased the maximum in descent over the pole and decreased the minimum in descent just inside the vortex edge. This view contrasts with the maximum descent proposed by Schoeberl *et al.* (1992) and provides a consistent picture when compared to long-lived trace gas observations in the winter lower stratosphere. The impact of PSCs on the latitudinal heating-rate gradient also strengthens the vortex, an effect enhanced by the ozone hole.

This study provides new information about the impact of PSCs in the presence of realistic tropospheric clouds. However, the lack of knowledge about actual PSC composition and distribution means that these results can be significantly improved upon. A complete evaluation of the radiative effects of PSCs must await the time when satellite retrievals of PSC composition and thickness occur throughout a winter and spring and over all regions of the polar vortex, particularly over the pole where PSCs can have their greatest effect.

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