Radiative Influence of Antarctica on the Polar-Night Vortex

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Abstract

Temperatures over the Antarctic plateau are sharply colder than those over its maritime surroundings. The sharp temperature contrast due to Antarctica is conveyed upward through 9.6-μm absorption by ozone, which shapes the thermal structure in the stratosphere. The radiative impact of Antarctica on the polar stratosphere is investigated in three-dimensional integrations of the nonlinear primitive equations, coupled to a full radiative-transfer calculation that is performed with and without clouds. Cooling associated with Antarctica depresses radiative-equilibrium temperatures by as much as 10 K. This direct radiative influence emerges clearly at high latitudes of the lowermost stratosphere. It is accompanied elsewhere by temperature changes of opposite sign, which result indirectly through adiabatic warming by the induced residual meridional circulation. Collectively, these influences reinforce the polar-night vortex, shift the jet axis poleward, and intensify downward transport over the polar cap by the residual circulation. In this way, radiative forcing from below contributes significantly to the features that distinguish the Antarctic vortex from the Arctic vortex.

1. Introduction

Observed differences between the Antarctic polar-night vortex and its counterpart over the Arctic are widely recognized. The vortex is significantly stronger and colder during austral winter than during boreal winter. These differences, especially in thermal structure, are instrumental in facilitating ozone depletions that occur each spring over Antarctica.

Features distinguishing the Antarctic vortex are thought to follow from weaker planetary waves in the Southern Hemisphere, which leave the stratospheric circulation less disturbed and closer to radiative equilibrium. Smaller topographic forcing, the Andes being the only major orographic feature in the Southern Hemisphere, leads to wave amplitudes that are considerably smaller during austral winter than during boreal winter. Wave drag on the zonal-mean flow is therefore weaker, as is its forcing of the residual mean circulation, which regulates polar temperatures through subsidence and adiabatic warming.

Another feature of the Southern Hemisphere, however, must also be considered. Temperatures at and immediately above the Antarctic surface are considerably colder than those found over the Arctic. They are also sharply colder than those over the maritime surroundings of the Antarctic plateau, behavior that steepens the meridional temperature gradient at polar latitudes. This sharp temperature contrast due to Antarctica influences the stratosphere radiatively. Longwave radiation emitted by the surface at wavelengths of 8–12 μm, within the atmospheric window, passes relatively freely through the troposphere until it is absorbed in the lower stratosphere by the 9.6-μm O₃ band. The sharp temperature contrast due to Antarctica should therefore introduce a similar contrast in longwave heating, which shapes the thermal structure in the lower stratosphere. Cooling by Antarctica acts to deepen the polar trough and thereby reinforce the polar vortex.

We evaluate the radiative impact of Antarctica in three-dimensional integrations of the nonlinear primitive equations, coupled to a full radiative-transfer calculation. Following a description of the calculation, section 3 presents the thermal and dynamical structure that is obtained with Antarctica present. How Antarctica influences this structure is first evaluated by examining the radiative-equilibrium temperature, which is determined with and without radiative forcing by Antarctica in section 4. Section 5 then examines Antarctica’s net radiative impact in the presence of dynamics, specifically, its influence on thermal structure, motion, and the
residual mean circulation. In sections 3–5, calculations are performed under cloud-free conditions to clearly illustrate the key physical effects introduced by the continent. Calculations with clouds are then discussed in section 6. The results imply that radiative forcing from below contributes significantly to the features that distinguish the Antarctic vortex from the Arctic vortex.

2. Numerical framework

The primitive equations are solved in isentropic coordinates with the spectral model described in Callaghan et al. (1999). The dynamical model extends from the $\theta = 375$ K isentropic surface upward through the mesosphere, terminating in a deep sponge layer that achieves the radiation condition. Rossby wave and gravity wave dynamics project naturally onto the Hough basis used in the spectral expansions (Kasahara 1977, 1978). Gravity wave breaking is parameterized using Rayleigh friction (Holton and Wehrbein 1980a,b). Forced by observed structure in the upper troposphere, the 3D model provides a respectable simulation of the dynamical and chemical structure characteristic of the wintertime middle atmosphere (Fusco and Salby 1999). In particular, it reproduces the major features of the observed polar-night vortex, the residual mean circulation, and the distribution of ozone.

For the present purpose, the integration is coupled to a multispectral radiative transfer calculation using the radiative transfer package of the NCAR Community Climate Model version 2 (CCM2) (Hack et al. 1993). Below 50 km, longwave radiative transfer is calculated using the band model of Kiehl and Briegleb (1991). At higher altitudes, nonlocal thermodynamic equilibrium is accounted for as in Garcia et al. (1992) through Newtonian cooling, which, in the global mean, balances shortwave heating (Schoeberl and Strobel 1978; Holton and Wehrbein 1980a,b).

The radiative calculation is performed from the surface upward, utilizing 3D distributions of absorbers that are prescribed from observations. Tropospheric properties like surface temperature, air temperature, humidity, cloudiness, and geopotential height are specified as mean distributions, derived from the International Satellite Cloud Climatology Project (ISCCP) C2 climatology, TIROS-N Operational Vertical Sounder (TOVS) retrievals, and Met Office meteorological analyses. At high latitudes, where ISCCP climatologies are unreliable, cloudiness is based on estimates from Berlyand and Stroquina (1980) and Sasamori et al. (1972). ISCCP/TOVS supplies two relevant temperature datasets: clear-sky brightness temperature for surface temperature, and air temperatures for the model troposphere. In the stratosphere, $O_3$ is obtained from MLS retrievals and NO$_2$ from CLAES. Carbon dioxide is fixed at 340 ppmv throughout. With these properties specified, the radiative calculation yields upwelling and downwelling fluxes and net radiative heating rates, which are then passed to the dynamical calculation.

3. Simulated behavior under wintertime conditions

The model is forced by climatological structure in the upper troposphere, inclusive of planetary wave activity, and with observed distributions of absorbers, both under July conditions. Figure 1 shows the time- and zonal-mean thermal structure recovered from the simulation, which is in good agreement with observed structure (Randel 1992). Temperatures decrease to a minimum of $\sim 185$ K over the South Pole near 25 km. Temperatures of 200 K and colder at the tropical tropopause are likewise consistent with observed structure, as is the separated stratopause (Hitchman et al. 1989; Kanzawa 1989), which is displaced to about 60 km over the winter pole. Corresponding zonal-mean winds (not shown), which approach 100 m s$^{-1}$ in the polar-night jet, are also in good agreement with climatology.

Figure 2 shows the residual mean vertical velocity $\mathbf{\nabla} \cdot \mathbf{u}$, which, in isentropic coordinates, follows directly from the (density-weighted) zonal-mean diabatic velocity $\mathbf{u}$ (Andrews et al. 1987). Air sinks diabatically over the winter extratropics, while it rises diabatically in the Tropics and subtropics of the summer hemisphere. Both branches of the residual circulation intensify upward, diabatic subsidence reaching 2.5 mm s$^{-1}$ near 40 km. The mass overturning implied in Fig. 2 is in good agreement with that deduced from observations of chemical tracers and calculations of radiative cooling (Fischer et al. 1993; Kiehl and Solomon 1986; Manney et al. 1994;
Fig. 2. Residual mean vertical velocity $\bar{w}$ ($10^{-5}$ m s$^{-1}$) calculated under Jul conditions.

Fig. 3. Radiative-equilibrium temperature (K) under Jul conditions.

perform a purely radiative calculation to obtain $T_{re}$ by solving the equation

$$\frac{\partial T}{\partial t} = \frac{1}{c_p} (q_{LW} + q_{SW}), \quad (1)$$

which is marched through an annual cycle. Here $q_{LW}$ and $q_{SW}$ are, respectively, the longwave and shortwave heating rates, which are computed as described in section 2. The calculation is performed from the $\theta = 375$ K isentropic surface upward through the mesosphere, underlain with a prescribed troposphere and surface.

The resulting radiative-equilibrium temperature for August is shown in Fig. 3. Consistent with the detailed calculations of Fels (1985), it reaches exceedingly low values over the summer pole, far colder than observed temperatures (Randel 1992). The sharp gradient of radiative-equilibrium temperature evident across the polar-night terminator is also largely absent in climatology. Observed temperatures over the winter pole are significantly warmer than $T_{re}$, even over the South Pole. They reflect adiabatic warming through the residual mean circulation which, in the stratosphere, is driven by planetary waves.

Planetary waves are only one influence that regulates polar temperatures, indirectly through the induced residual mean circulation. Polar temperatures are also influenced by Antarctica directly through radiative transfer. The cold Antarctic surface introduces less upward LW flux into the stratosphere than its warmer maritime surroundings. Absorption of upwelling 9.6-µm radiation by ozone leads to a colder polar stratosphere above Antarctica, relative to lower latitudes. This influence should be reflected in the radiative-equilibrium temperature.

Rosenfeld et al. 1987; Rosenfeld et al. 1994; Schoeberl et al. 1992). Discernible in the high-latitude structure of $\bar{w}$ is a downward displacement of contours poleward of 72°S, particularly below 25 km. It defines a region of intensified subsidence and is a manifestation of Antarctica’s radiative influence, as will be shown below.

4. Radiative-equilibrium temperature

Some insight into the structure emerging from the coupled dynamical–radiative calculation comes from the radiative-equilibrium temperature $T_{re}$, toward which radiative transfer drives the circulation. For the deep temperature structure characteristic of middle atmosphere planetary waves, and outside the lower stratosphere, radiative transfer is described approximately by the cool-to-space approximation (Rodgers and Walshaw 1966). Upon linearization, this reduces to Newtonian cooling. Radiative heating is then proportional to the temperature departure from $T_{re}$, with a spatially dependent cooling rate that varies cubically with temperature (e.g., Andrews et al. 1987). It is therefore instructive to examine $T_{re}$ and how it is influenced by the presence of Antarctica.

In principle, $T_{re}$ could be obtained from the coupled dynamical-radiative integration by omitting dynamical forcing terms. However, computational considerations make this impractical. Without wave drag and the accompanying mean circulation, the polar vortex achieves very large velocities that require a very short time step to maintain numerical stability. As an alternative, we
a. Elimination of Antarctica

Antarctica’s radiative influence is gauged by comparing the July radiative-equilibrium temperature (Fig. 3) against that of a companion calculation in which Antarctica is removed. To eliminate the continent, we extrapolate surface temperature toward the pole from the surrounding ocean, employing observed temperature gradients near 60°S. (These gradients are typical of much of the maritime region surrounding Antarctica, and are similar to Northern Hemisphere zonal mean gradients in January.) Antarctica’s elevation is also eliminated, with temperatures extrapolated at sea level. Zonal-mean surface temperatures with and without Antarctica are shown in Fig. 4. The presence of Antarctica is quite evident in the abrupt change of surface temperature gradient near 60°S (solid). Removing the continent (dashed) therefore leads to elevated temperatures over the pole, by as much as 23 K.

From the Stefan-Boltzmann law, the differences in surface temperature evident in Fig. 4 imply that upwelling longwave fluxes in the presence of Antarctica are smaller, by as much as 50%. Longwave heating at wavelengths within the atmospheric window (e.g., in the 9.6-μm band of ozone) should then reflect a commensurate reduction, which in turn should be reflected in a depression of radiative-equilibrium temperature.

Antarctica’s elevation and cold surface also influence temperatures in the overlying troposphere. In the lowermost troposphere, air temperatures drop as the pole is approached, reflecting the even larger drop in surface temperature. Lower-tropospheric temperatures over the continent are coupled to the surface through vertical transport, but are mitigated by horizontal transport that mixes in warmer maritime air from the surroundings.

Removing the continent therefore also implies changing the troposphere’s thermal structure. We estimate this change by again extrapolating poleward from maritime regions, using observed tropospheric temperature gradients near 60°S. As in Fig. 4, Antarctica steepens the meridional tropospheric temperature gradient, which is left more uniform in its absence. In the lowermost stratosphere, removing Antarctica raises tropospheric temperatures at the pole more than 10 K (not shown). Temperature changes in the upper troposphere are small, less than 2 K above 8 km. This implies that most of the ~50% difference in longwave flux associated with Antarctica passes through the troposphere to reach the stratosphere, where it is absorbed in the 9.6-μm O₃ band.

b. Modified radiative-equilibrium temperature

With Antarctica eliminated, we now recompute $\bar{T}_{re}$ and compare the result to the July radiative-equilibrium temperature in Fig. 3. Figure 5 shows the difference in radiative-equilibrium temperature, that is, the change $\Delta T_{re}$ introduced by Antarctica. The continent’s radiative influence, which depresses $\bar{T}_{re}$, is substantial, implying that the lower stratosphere is efficiently coupled to the surface by longwave radiative exchange in the 9.6-μm band of O₃. The depression of radiative-equilibrium temperature reaches ~9.6 K near 20 km and 80°S. Introduced primarily by the change in surface temperature, it diminishes to ~2 K near the stratopause and is zero equatorward of 63°S. Although strongest in the lower stratosphere, the depression of radiative-equilibrium temperature extends upward throughout the polar stratosphere and lower mesosphere.

The structure of $\Delta T_{re}$ results from two effects. First, the underlying troposphere is largely transparent to upwelling longwave radiation emitted by the surface at wavelengths within the atmospheric window. This leads
to absorption in the 9.6-μm O₃ band, which couples the lower stratosphere to the thermal structure below. Sharp horizontal variations of surface temperature then introduce similar variations in longwave heating, which are manifested in $\Delta T_{re}$. Second, the O₃ optical depth varies rapidly with height in the lower stratosphere. This controls the net heating rate, which is then largest at altitudes where ozone is concentrated.

Antarctica’s contribution to radiative-equilibrium temperature during other months has a structure similar to that during July. The temperature depression varies through an annual cycle (Fig. 6) with a broad minimum near $-9.6$ K that persists from July through October. The radiative influence of Antarctica is therefore substantial throughout much of the year. In addition, because the annual minimum in Fig. 6 is broad, its structure is well approximated by its mean over the winter season. Dynamical simulations performed under July conditions should therefore capture the salient features of the continent’s radiative influence.

5. Radiative influence in the presence of dynamics

To quantify the influence that Antarctica actually exerts on the wintertime stratosphere, we now repeat the coupled dynamical-radiative simulation under July conditions (section 3), but with Antarctica removed. This parallels the development in section 4. However, while accounting for radiative forcing by Antarctica, which cools the polar stratosphere diabatically, this calculation also accounts for dynamical forcing, which warms the polar stratosphere adiabatically through subsidence in the induced residual circulation. The behavior recovered is then compared against the reference state presented in section 3 and interpreted in terms of the radiative-equilibrium thermal structure of section 4.

a. Thermal structure

Figure 7 shows the temperature difference $\Delta T$ associated with Antarctica, which may be compared against the radiative-equilibrium temperature difference in Fig. 5. Unlike $\Delta T_{re}$, the temperature difference in Fig. 7 is not limited to polar latitudes or even to a single sign. The continent’s direct radiative effect is found over the pole in the lowermost stratosphere, where $\Delta T$ reaches $-6.5$ K, more than half of the change anticipated from the change in radiative-equilibrium temperature. (The limiting behavior, which requires extending the integration longer than the winter season, yields a temperature depression of $\Delta T = -8$ K, even closer to $\Delta T_{re}$.) The discrepancy between $\Delta T$ and $\Delta T_{re}$ in the lowermost stratosphere results from adiabatic warming associated with the residual circulation (Fig. 2), which opposes the direct radiative effect of Antarctica and is likewise modified by its presence. Outside of the lowermost polar stratosphere, this indirect effect prevails.

Higher and equatorward of the temperature depression in Fig. 7, the temperature change actually has the opposite sign ($\Delta T > 0$) even though the direct radiative influence of Antarctica cools the polar stratosphere (as reflected in $\Delta T_{re} < 0$). Increased temperatures above the polar O₃ maximum near 18 km ($\Delta T > 0$) result from enhanced subsidence in the induced residual circulation, as will be seen shortly. The same applies to the temperature increase in the lowermost stratosphere equatorward of the continent. On the other hand, the temperature depression near the stratopause and 40°S results from adiabatic cooling, which is associated with diminished sinking motion (enhanced rising motion) that compensates enhanced sinking motion over the winter pole. A weak influence is evident even over the summer pole.
poleward by 10°–15°. Oppositely signed wind changes located poleward and equatorward of the continent’s edge (Fig. 8) also increase horizontal wind shear, which then increases vorticity and tends to situate the vortex edge over this region.

Antarctica also deepens the cyclonic minimum of potential vorticity (PV) over the South Pole (Fig. 9). The PV difference is negative (the same sign as ambient PV) poleward of 60°S. In this region, it intensifies total PV by more than 25%. The radiative influence of Antarctica thus provides a source of PV not available to the Arctic vortex. Since the corresponding wind field difference is almost zonal, Antarctica’s influence acts to reinforce the polar symmetry of the vortex (e.g., against eddy advection). Equatorward of the continent, the change in PV has opposite sign, so Antarctica weakens vorticity there. Together, these oppositely signed PV changes reflect a poleward shift of the vortex edge, along with the accompanying jet axis. The steepened gradient of PV that results also reinforces the vortex against eddy mixing by planetary waves, better isolating air in its interior.

c. Residual mean circulation

The depression of polar temperature $\Delta T$ falls short of that anticipated from the change in radiative-equilibrium temperature $\Delta T_{\text{eq}}$. In addition, outside the region of strong 9.6-$\mu$m absorption in the lowermost polar stratosphere, $\Delta T$ is actually positive, not negative. This discrepancy is accounted for by mean vertical motion, specifically, adiabatic warming, which accompanies subsidence in the induced residual circulation, and adiabatic cooling, which accompanies ascent. This influence is expressed conveniently in isentropic coordinates through the vertical velocity $\dot{\theta}$.

Figure 10 shows the percentage change in $\dot{\theta}$ induced by Antarctica. Inside the vortex, the continent enhances diabatic descent in the lowermost stratosphere by nearly 20%. The largest percentage change occurs near the $\text{O}_1$ heating maximum where it reinforces residual mean motion. This change in descent rate is consistent with estimates based on the Newtonian cooling approximation and the change in radiative-equilibrium temperature (Fig. 5). For example, the maximum change in $\dot{T}_{\text{res}}$ (near 80°S and 18-km altitude) is nearly 10 K. Incorporating this into the departure from radiative equilibrium with and without Antarctica gives $(\dot{T} - \dot{T}_{\text{res}})_{\text{Ant}} \approx 19$ K and $(\dot{T} - \dot{T}_{\text{res}})_{\text{NoAnt}} \approx 14$ K, which implies that $\Delta \dot{T}/\dot{T} \approx 26\%$, close to the 20% change recovered in the coupled dynamical-radiative integration.

The preceding analysis focused on the percentage change of diabatic descent rate in the lowermost polar stratosphere, the region where Antarctica’s radiative influence is strongest. After decreasing above the ozone layer, the percentage change in $\dot{\theta}$ increases again in the upper stratosphere, as well as equatorward of the continent. The absolute change of $\dot{\theta}$ is plotted in Fig. 11. It clearly reflects the induced residual circulation, which
accounts for the positively signed temperature differences situated above and equatorward of the lowermost polar stratosphere (Fig. 7), where radiative cooling by Antarctica is strong. Poleward of 60°S, Antarctica leads to enhanced subsidence—not just inside the ozone layer but at all altitudes (Fig. 11). Induced subsidence intensifies upward, approaching \(-14\) K day\(^{-1}\) near the stratopause. Absolute vertical motion (Fig. 2) below 20 km is strongest poleward of 72°S. Above 20 km the strongest diabatic descent occurs at the periphery of the vortex, consistent with the trajectory calculations of Manney et al. (1994). Changes above the O\(_3\) layer follow adiabatically through the induced residual motion, which must satisfy continuity. Air subsiding inside the lowermost stratosphere (\(\Delta \dot{\theta} < 0\)), where 9.6-\(\mu\)m absorption is strong, is replaced by air drawn downward from overhead. Mass continuity also requires that this subsiding motion be compensated by poleward flow, which in turn is compensated by rising motion equatorward of the continent (\(\Delta \dot{\theta} > 0\)).

These differences in \(\dot{\theta}\) describe the residual circulation induced by Antarctica, specifically, through its depression of \(T_{\text{re}}\) in the polar stratosphere relative to the maritime surroundings. Adiabatic warming associated with enhanced subsidence offsets less than half of the diabatic cooling in the lowermost stratosphere, but it prevails above the ozone layer. Equatorward of Antarctica, adiabatic cooling associated with enhanced rising motion introduces a temperature difference of the opposite sign. In combination, these temperature tendencies reinforce the vortex and shift the jet axis poleward.

Changes of residual mean circulation reflect changes of the Eliassen–Palm flux divergence \(\nabla \cdot \mathbf{F}\) (Fig. 12). Antarctica’s radiative influence alters wave absorption throughout the winter stratosphere, especially at high latitudes, where \(T_{\text{re}}\) is depressed. Poleward of 55°S, the change in \(\nabla \cdot \mathbf{F}\) is chiefly negative, so it reinforces existing wave drag. Increased westward acceleration experienced by the mean flow is then acted upon by a Coriolis force that deflects air increasingly southward. Air drifting southward (reflected in intensified \(\tau^*\) that balances \(\nabla \cdot \mathbf{F}\) in the zonal-mean equations) converges over the pole and sinks. Compressional warming that attends this sinking motion then leads to the increased

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**Fig. 9.** Differences in Ertel potential vorticity (K m\(^2\) kg\(^{-1}\) s\(^{-1}\)) and horizontal velocity (m s\(^{-1}\)) due to Antarctica under Jul conditions. The figure is plotted on the \(\theta = 823\) K isentropic surface, near 30 km.
temperatures found above the region of strong cooling by Antarctica (Fig. 7).

Intensified $\nabla \cdot \mathbf{F}$ in Fig. 12 reflects intensified absorption of planetary wave activity. The depression of radiative equilibrium temperature by Antarctica acts to strengthen the vortex, but it also accelerates diabatic effects that dissipate wave activity. Since $T_w$ is reduced over Antarctica, but not at lower latitudes, air displaced southward into the polar cap finds itself farther from radiative equilibrium. It therefore experiences intensified cooling, which accelerates the destruction of eddy PV and dissipation of wave activity. As $\nabla \cdot \mathbf{F}$ is balanced by $f\vec{u}^*$ in the momentum budget, which in turn is balanced by $\vec{w}^*$ to maintain continuity, Antarctica leads ultimately to intensified subsidence over the polar cap (Fig. 11).

6. Cloud influences

The results of previous sections follow from coupled dynamical–radiative simulations performed under cloud-free conditions, in order to isolate the key physical effects introduced by Antarctica. These are modified by cloudiness. Total cloud fraction over Antarctica is greatest near the continent’s edge, falling off steadily as the pole is approached. In July, typical cloud cover is 70% at 60°S and 30% at the pole (Liou 1992).

The meridional variation of cloud fraction therefore mirrors the poleward decrease in Antarctica’s surface temperature. Thus, the continent’s strongest radiative effects are communicated to the stratosphere through the least-cloudy region of the polar troposphere. Between the pole and the Antarctic coast, radiative coupling between the troposphere and stratosphere through 9.6-μm absorption involves a smaller surface contribution and a progressively larger contribution from clouds.

![Fig. 10. Percentage change in $\theta$ due to Antarctica under Jul conditions.](image1)

![Fig. 11. Absolute change in $\frac{K}{day}$ due to Antarctica under Jul conditions.](image2)

![Fig. 12. Change in density-weighted Eliassen–Palm flux divergence $L(\sigma)\nabla \cdot \mathbf{F}$ (m s$^{-1}$ day$^{-1}$) due to Antarctica under Jul conditions. Here $\sigma$ plays the role of density in isentropic coordinates.](image3)
Calculating Antarctica’s radiative influence under cloudy conditions requires an estimate of the cloud fraction in the continent’s absence. In the Arctic, during northern winter, cloud cover also extends to the pole, decreasing from about 65% at 60°N to 50% at 90°N (Liou 1992). With Antarctica removed, a similar cloud fraction is applied at southern latitudes during July. Cloud climatologies allow estimates of both high and low cloud fraction and cloud height. Cloud cover (assumed optically thick) is dominated by low cloud fraction, its temperature variation reflecting the strong drop of surface temperature toward the pole.

With clouds present, Antarctica’s influence reduces upwelling fluxes by as much as 45% (not shown), nearly the same as the clear-sky change (section 4a). A significant radiative influence is therefore implied even under cloudy conditions.

This is confirmed through coupled dynamical–radiative calculations that include clouds. These preserve the qualitative relationships developed above under clear-sky conditions. Sixty-five to eighty percent of the clear-sky influence on temperature, wind, and diabatic descent is preserved in the cloudy simulations. Antarctica’s radiative influence is therefore a robust component of the middle atmosphere circulation.

7. Conclusions

Weak planetary waves in the Southern Hemisphere shape the stratospheric circulation during austral winter. However, so too does the radiative impact of the Antarctic plateau, which is conveyed upward through 9.6-μm absorption by ozone. Thermal and dynamical structure of the lower stratosphere are both influenced significantly by Antarctica. Temperature, wind, and residual mean motion are all shaped by the depression of radiative-equilibrium temperature over the pole. Antarctica’s contribution to θ in the lower stratosphere accounts for about one-half of the observed difference from the northern winter Arctic vortex, making its radiative influence a significant factor. Higher, Antarctica’s influence wanes, the contribution from planetary waves eventually prevailing.

In modifying radiative-equilibrium temperatures in the lower stratosphere, Antarctica also alters θ and the downward branch of the Brewer-Dobson circulation. This has important consequences for ozone, particularly for springtime depletions over the South Pole. Intensified subsidence enhances the transfer of ozone into the polar stratosphere from its chemical source region above and equatorward. Because Antarctica’s radiative impact is greatest in the lowermost stratosphere, these changes contribute to shaping the ozone column abundance, which is concentrated at the same altitudes.

Antarctica’s influence also has implications for ozone depletion through its regulation of polar temperature. By depressing polar temperatures in the lower stratosphere, Antarctica provides conditions favorable for the formation of polar stratospheric clouds (PSCs). Specifically, radiative cooling maintains lower-stratospheric temperatures across much of Antarctica below 195 K, the threshold for type I PSC formation (WMO 1995). In addition, temperatures below 188 K, the threshold for type II PSCs, expand downward to altitudes where water vapor mixing ratios are more favorable for cloud formation. Jointly with dynamical changes that lead to better isolation of polar air, these influences make possible the conditions that favor heterogeneous chemical processes that culminate each spring in ozone depletion.

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