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Maintenance of polar stratospheric clouds in a moist stratosphere

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CPD

3, 935–960, 2007

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

EGU

Abstract

Previous work has shown that polar stratospheric clouds (PSCs) could have acted to substantially warm high latitude regions during past warm climates such as the Eocene (55 Ma). Using a simple model of stratospheric water vapor transport and polar stratospheric cloud (PSC) formation, we investigate the dependence of PSC optical depth on tropopause temperature, cloud microphysical parameters, stratospheric overturning, and tropospheric methane. We show that PSC radiative effects can help slow removal of water from the stratosphere via self-heating. However, we also show that the ability of PSCs to have a substantial impact on climate depends strongly on the PSC particle number density and the strength of the overturning circulation. Thus even a large source of stratospheric water vapor (e.g. from methane oxidation) will not result in substantial PSC radiative effects unless PSC ice crystal number density is high, and stratospheric overturning (which modulates polar stratospheric temperatures) is low. These results are supported by analysis of a series of runs of the NCAR WACCM model with methane concentrations varying up to one thousand times present levels.

1 Introduction

The fundamental challenge to climate dynamics posed by the paleoclimate record of the Eocene, when polar winter temperatures were well above freezing, but tropical temperatures were not much warmer than at present, remains unresolved (Huber and Sloan, 2001). Sloan and Pollard (1998) and Kirk-Davidoff et al. (2002), hereinafter referred to as KSA) investigated the possible role of polar stratospheric clouds (PSCs) in maintaining the extreme polar warmth of the Eocene climate. Sloan and Pollard (1998) showed that an optically thick (optical depth $\tau=1$) cloud inserted into a GCM stratosphere substantially warmed polar surface temperatures. KSA examined the plausibility of a dynamical feedback mechanism whereby reduced pole-to-equator temperature difference could yield reduced stratospheric overturning, increased tropical tropopause

CPD

3, 935–960, 2007

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

EGU

temperature, a moister stratosphere, increased PSC optical depth, polar warming and thus further reduced pole-to-equator temperature difference. They showed that this feedback cycle led to clouds formed under plausible physical constraints (reasonable cloud particle number density, cloud water content, and explicitly calculated optical depth) with sufficient optical depth (though generally less than unity) to warm the polar surface.

KSA left many details of their proposed feedback cycle untested. They did not take into account either the reduction of water vapor transport to the polar cap that would accompany reduced overturning circulation, or the loss of stratospheric water vapor due to sedimentation of PSC ice particles. Here we explore these aspects of the stratospheric water vapor budget using the moderate-complexity model described in KSA, modified to account for transport of tracers by the stratospheric mean circulation, and to include loss of water by sedimentation of cloud particles. We test the sensitivity of PSC optical depth to assumptions about the ice crystal physics, cloud-radiation interactions, tropospheric temperature distribution and stratospheric overturning circulation strength. In addition, we compare these results to those of simulations using the NCAR Whole-Atmosphere Community Climate Model (WACCM), which were performed to investigate the radiative and chemical role of increased methane in the Permian atmosphere (Lamarque et al., 2006).

The microphysics of PSCs are understood only approximately, so that extrapolation to conditions of substantially high stratospheric water vapor is difficult. In the modern atmosphere, PSCs form every winter in both the Arctic and Antarctic stratosphere. In the Arctic they consist primarily of Nitric Acid Trihydrate (NAT) crystals (type I PSCs), and are mostly found in cold regions formed by gravity waves in the lee of mountain ranges. These clouds are able to substantially reduce stratospheric nitrogen content over the course of the winter, but do not remove much water (Drdla et al., 2002). Colder temperatures in the Antarctic winter allow formation of water ice clouds (type II PSCs) by freezing of sulfuric acid droplets. By the end of the winter, sedimentation of the ice crystals forming these clouds results in substantial dehydration of the polar

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

vortex. These PSCs typically have number densities of $0.01\text{--}1\text{ cm}^{-3}$ and crystal sizes of a few microns (Peter, 1997).

The processes controlling the number density of PSCs and of the liquid sulphate aerosol are not well understood, and so it is hard to constrain these numbers in paleoclimate studies. Our approach therefor will be to analyze the dependence on PSC radiative properties by assuming a wide range of possible number densities, so that we can determine how the various still unknown features of the past stratosphere might constrain the role of PSCs in warming the poles. PSC properties are not typically comprehensively treated in climate models. When they are specifically predicted for studies of stratospheric chemistry, their radiative interactions are often assumed away, while climate prediction models typically use empirically designed cirrus parameterizations derived from tropospheric observations throughout the depth of the atmosphere. Since the distribution of potential cloud condensation nuclei is quite different in the stratosphere and troposphere (Brock et al., 1995), it is not clear that this approach is well justified.

In this paper we will address the following questions:

- For which particle number densities is substantial infrared radiative heating by PSCs possible?
- Does radiative heating of PSCs themselves substantially affect their abundance?
- In a low-temperature gradient world, are thick PSCs possible without a reduction in stratospheric overturning?
- In a low-overturning world, can sufficient water vapor be provided to the poles to match sedimentation and produce substantial PSCs?
- Can additional methane alone, without changes to overturning circulation cause substantial increases in PSC optical depth?

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

2 Model Description

The model used to predict the stratospheric water vapor budget and polar stratospheric cloud formation under varying assumptions of particle number density and stratospheric overturning is described in detail in Kirk-Davidoff and Lindzen (2000) and in KSA. In the calculations described below, the model is run with prescribed, annually varying surface temperature, to focus on the behavior of the PSCs.

The model is two dimensional in pressure and latitude, with a meridional resolution of three degrees latitude, and a vertical resolution of 15 hPa in pressure. Tropospheric temperatures are set to lie along moist adiabats in the tropics, while an assumed fixed distribution of potential vorticity results is used to determine the temperature in the mid-latitude troposphere, given the surface temperature, and the surface static stability. Tropospheric relative humidity is fixed. Temperatures in the stratosphere are allowed to come into radiative equilibrium through a thermodynamic equation, and the tropopause height is chosen to minimize the temperature discontinuity at the tropopause. The radiative heating terms are calculated using the schemes of Chou et al. (2001) and Chou and Suarez (1999), and the stratospheric overturning is represented by a term of the form:

$$Q_{BD} = 0.8 \text{ K d}^{-1} \frac{\Delta T_{pe}}{45 \text{ K}} \cos(\pi \sin(\phi)) \quad (1)$$

where Q_{BD} is the heating due to the stratospheric overturning (Brewer-Dobson) circulation, ΔT_{pe} is the pole-to-equator temperature difference in the present hemisphere, and ϕ is the latitude. The two constants are chosen so that for the present climate conditions, the mean stratospheric cooling due to the overturning averages to 0.5 K over the tropical region. As discussed in KSA, the dependence of the stratospheric overturning circulation on the surface temperature gradient is based on the idea that tropospheric temperature gradients should control baroclinic wave growth rates, and that tropospheric waves propagating into the stratosphere and breaking there are the cause of upwelling in the tropical stratosphere and downwelling in the polar strato-

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

sphere (Kawamoto and Shiotani, 2000).

The surface static stability varies linearly with latitude so that it is continuous across the edge of the tropics at 30 degrees latitude in each hemisphere, and approaches a value at the poles which is adjusted so that the isentrope leaving the surface at the equator intersects the tropopause at the poles. This has the effect of reducing polar static stability as the pole-to-equator temperature difference decreases.

Cloud fraction and optical depth are fixed in the troposphere. In the stratosphere they are calculated as follows. Ice particle number density is taken to be an adjustable parameter, but is held fixed for the duration of each model run. Cloud ice mixing ratio set equal to excess of water vapor mixing ratio over saturation mixing ratio. Excesses can occur due to local cooling or to transport of water vapor. The effective radius of stratospheric cloud ice particles is derived from number density and cloud ice mixing ratio, assuming spherical particles, so that

$$R = \left(\frac{3}{4\pi N \rho_i} r_i \right)^{1/3} \quad (2)$$

where R is the effective radius, N is the particle number density, r_i is the cloud ice mixing ratio, and ρ_i is the density of ice.

Water loss by cloud particle sedimentation is determined by assuming that particles fall at the Stokes velocity, with no reevaporation:

$$v_s = \frac{2}{9} \frac{\rho_i g}{\nu \rho} R^2 \quad (3)$$

where v_s is the Stokes velocity, g is the acceleration of gravity, ν is the kinematic viscosity of air, and ρ is the air density.

Cloud water infrared optical depth is calculated in the Chou et al. (2001) code by assuming spherical particles, and using a ten band look-up table for IR extinction coefficients:

$$\tau(l) = P_i * \left(a_1(l) + \frac{a_2(l)}{R^{a_3(l)}} \right) \quad (4)$$

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

where $\tau(l)$ is the optical depth for frequency band l , P_i is the liquid water path, equal to $r_i\rho$, and $a_1(l)$, $a_2(l)$ and $a_3(l)$ are empirical parameters.

Figure 1 shows the relationships between cloud water mixing ratio and cloud particle size, cloud water sedimentation loss rate for a 15 hPa grid box, and cloud water optical depth (for the 1100–1215 cm^{-1} band) that result from these relationships.

Transport of water and methane in the stratosphere are controlled by a simple semi-lagrangian differencing scheme, such that transport occurs only in the meridional dimension, and only in the poleward direction. At the equator, water vapor throughout the stratosphere is set to its value at the tropopause, while methane is set to its tropospheric value. As with the warming or cooling due to the overturning circulation, the magnitude scales linearly with the pole-to-equator temperature difference, so that for modern conditions water and methane are transported poleward with a velocity of 3 degrees latitude per month, yielding an “age of air” of about 3 years at the poles. This is fast compared to observed ages of air of 4 to 5 years, but agrees with typical model results (Scheele et al., 2005), including results for the WACCM cases we discuss here. For the Eocene case, the overturning circulation speed is cut by a factor of three, yielding an “age of air” of 9 years at the pole. We also perform sensitivity experiments showing results for “Modern” overturning with “Eocene” surface temperatures, and vice versa. Methane is converted to water vapor at a rate of 2% per month, which results in a conversion of about 2/3 of methane by the time air reaches the poles for modern conditions. Although the model is run for both hemispheres, annual mean boundary conditions and results are essentially symmetric about the equator, so we present only northern hemisphere results in the following figures.

3 Results from the simple model

Figures 2–7 show results for several experiments with the simple model. The “Modern” runs have an annual mean ΔT_{pe} of 46 K, with the temperature at the equator being 302 K, and the poles varying from 236 K to 276 K over the seasonal cycle. The

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

“Eocene” runs have an annual mean ΔT_{pe} of 15 K, with the temperature at the equator again 302 K, and the polar temperatures varying from 273 K to 293 K. In the base runs, PSC particle number densities were set to 10 cm^{-3} . Additional runs were performed with particle number densities of 0.01, 0.1, and 1 cm^{-3} . Runs were performed with widely varying methane concentrations, and with varying stratospheric overturning circulations strengths. Runs were also performed with “Eocene” overturning and “Modern” surface temperatures, and with “Modern” overturning for “Eocene” surface temperatures. Annual mean PSC optical depths reflect a seasonal cycle during which PSCs are typical absent except in winter, so that the annual mean optical depth is about four times smaller than winter-time mean values. For the “Eocene” control case, maximum PSC optical depths were 0.22, corresponding to a cloud water mixing ratio of 0.4 ppm

Figure 2 shows results for “Eocene” conditions with varying overturning circulation strength and for varying levels of tropospheric methane. The top panel shows annual mean infrared optical depth of simulated PSCs as a function of latitude. Optically thick PSCs are only present when the overturning circulation is reduced along with the pole-to-equator temperature gradient, because otherwise, with fast (“Modern”) overturning, temperature at the pole are too warm to allow condensation, even with water vapor mixing ratios of 15 ppmv (bottom panel). Fast overturning also results in colder temperature at the tropical tropopause (190 K, versus 203 K for “Eocene” overturning). This in turn results in much lower water vapor concentrations, if tropospheric methane is held fixed. The bottom panel shows that sedimentation from PSCs reduces water vapor concentrations towards the poles when PSCs are present.

Figure 3 shows results for varying overturning circulation and tropospheric methane mixing ratio for “Modern” conditions. For this model, tropospheric temperature conditions do not have much direct (radiative) impact on stratospheric temperatures, so the results are very similar to the results for “Eocene” conditions. Optically deep PSCs arise when sufficient stratospheric water vapor is present (due to either reduced stratospheric overturning, warming the tropical tropopause, or increased tropospheric

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

methane, or both), and temperatures in the polar regions are cold enough (for a given amount of water vapor) to form this clouds. For “Modern” stratospheric overturning values, this requires extremely high methane concentrations- over 100 times present values.

Figure 4 shows results for “Eocene” conditions with a range of assumed particle number densities. Optical depths are much higher for higher number densities, for three reasons. Sedimentation of cloud particles is smaller for higher number densities, since particles are smaller and have slower terminal velocities, optical depth is increased because the surface area to volume ratio of small particles is larger than that of large particles, and finally due to PSC radiative effects. Note that for all number densities at or below 1 cm^{-3} , water vapor concentrations are substantially lower than for a high number density (10 cm^{-3}). This is so both because of the higher water losses by sedimentation with higher number density and because wintertime temperatures for clouds with high PSC number densities and high optical depths are substantially higher, so less water vapor is lost by condensation and sedimentation. These factors are disentangled in Figs. 5 and 6, which explore the impact of PSC radiative forcing on the model results. Figure 5 shows that when PSC radiative effects are turned off, PSCs are shifted to lower latitudes, and water vapor concentrations are substantially reduced. Figure 6 shows the time series of monthly mean 75 hPa temperature and water vapor mixing ratio for the Eocene control case, and the case with PSC radiation turned off. When PSC radiative effects are operating, temperatures do not fall as low in winter, because the PSC raise the radiative equilibrium temperature of the stratosphere by increasing the emissivity in regions where there is strong upwelling infrared radiation. The warmer temperatures mean that water vapor mixing ratio is not reduced so much via condensation and sedimentation. The higher water vapor concentrations result in cooler temperatures for much of the year, when the energy balance in the stratosphere is between heating by ozone absorption of solar radiation, and infrared cooling, partially via water vapor. Thus the annual mean temperatures shown in Fig. 4 are only slightly cooler for high number densities.

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Figure 7 shows the two dimensional distribution of cloud water mixing ratio for two different number densities. It helps to explain why temperature in Fig. 4b depends so strongly on PSC particle number density. For low number densities (Fig. 7b), the radiative heating effect of PSCs is small, and the pattern of cloud water mixing ratio in latitude and pressure is uniform. At higher number density, the tendency PSC to raise local temperatures causes a patchy pattern of cloud water mixing ratio: temperatures hover near the frost point, and when temperatures dip below the frost point, large amount of water can condense on the large number of particles, without much loss by sedimentation. Temperatures in the grid square with clouds rise, causing evaporation. This cycle of condensation and evaporation leads to a patchy distribution of cloud water for high number densities, as shown in Fig. 7a.

We have shown that low overturning circulation is necessary in our simple model to allow temperatures low enough to form optically thick PSCs. Figure 8 addresses the question of how the reduced stratospheric water vapor and methane transport associated with reduced overturning affects the water vapor concentrations and PSC optical depth. Both runs shown here are for “Eocene” conditions, but the curves labeled “Fast Transport” have water vapor and methane transport rates corresponding to “Modern” conditions (three times greater than “Eocene” conditions), while the cooling and warming due to vertical motions in the stratosphere are held fixed at “Eocene” levels. As expected, panel c of this figure shows that water vapor concentrations are larger for the “Fast Transport” case, as are PSC optical depths. These effects should be contrasted with the results shown in Figs. 2 and 3, where both the transport and dynamic heating and cooling associated with the overturning circulation are changed- in that case stronger stratospheric overturning is clearly associated with much reduced PSC optical depth, due to increased high latitude temperatures. Thus, overturning changes do affect PSC optical depth via changes in water vapor supply to the polar regions, but this effect is dwarfed by the effect of overturning changes on tropopause temperature (and thus stratospheric water vapor concentration) and on polar night temperatures. For both of these latter mechanisms, increased overturning strongly reduces PSC op-

PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

tical depth.

4 Results from WACCM experiments

In this section we describe relevant results from a series of experiments performed using the NCAR WACCM, intended to simulate the impact of a large release of methane at the end of the Permian. The model and boundary conditions for these experiments are described in detail in Lamarque et al. (2006). Here we focus on the influence of changing tropospheric methane concentrations on stratospheric water vapor and cloud fraction. Figure 9 shows cloud ice mixing ratio, temperature and water vapor mixing ratio, all at the model level corresponding to about 72 hPa. As expected, increasing tropospheric methane yields increasing water vapor. As in the simple model, methane oxidizes to water vapor as air ages, resulting in an increase in water vapor from the tropics to the mid-latitudes. Beyond the mid-latitudes, water vapor is lost due to cloud formation and precipitation formation. In WACCM, ice crystal size is a function of temperature only, with stratospheric temperatures yielding crystals sizes of about 30 μm for temperature of 210 K, much larger than in our simple model. In addition, beyond about 3 ppmm of cloud ice, conversion to snow takes place, which results in a very rapid loss of cloud ice. These factors combine to prevent PSC optical depth from reaching significant levels, despite the large water vapor source due to methane oxidation. Thus Fig. 10 shows very smooth concentrations of cloud ice, without the patchiness evident in the simple model results, that occurs due to PSC self-heating.

Figure 11 shows plots of methane concentration versus age of air (the number of years since a parcel in the stratosphere entered the stratosphere via the tropical tropopause). The nearly monotonic relationship between age of air and methane concentration, and between age of air and latitude (not shown) explains why the simple models trivial transport scheme (air moves at a constant rate from the tropics to the poles) can approximately reproduce the patterns of methane, water vapor and cloud ice seen in WACCM, if ice parameters are adjusted so that PSCs have little infrared

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

optical depth.

5 Discussion and conclusions

Our simple modeling results demonstrate that PSCs of optical depth sufficient to cause significant surface warming require high particle number densities (exceeding 1 cm^{-3}), and either weak stratospheric overturning or extremely high tropospheric methane concentrations.

As speculated in KSA, we demonstrate here that in our simple model optically thick PSC's self-heating can substantially reduce water loss via sedimentation. Optically thick PSC warm the stratosphere locally, so that model temperature tend to hover around the frost point. Thus optically thick PSCs can evaporate before substantially sedimentation takes place. This provides some basis for arguing that in a moister stratosphere the fractional rate of loss via sedimentation might not be substantially larger than in the present, helping explain sustained wintertime high latitude warmth.

We have shown that In the absence of reduced overturning circulation (which tends to cool the polar stratosphere), very high levels of water vapor (whether from methane oxidation or some other source) are required to allow optically thick PSCs, even for very high number densities.

Our results further demonstrate that the reduction of water vapor transport associated with weak overturning does not prevent formation of optically thick PSCs, for sufficiently high particle number density. PSC particle number densities of 10 cm^{-3} have been observed, but are uncommon in the present atmosphere, where much lower densities are typical (Peter, 1997). Thus, for optically thick PSCs to have been part of the solution to Eocene polar warmth, there must have been some change in ice-formation microphysics. One can speculate that in a much moister stratosphere, more favorable condensation nuclei would be rapidly removed, leaving only the background sulphate aerosol, with a number density of approximately 6 cm^{-3} (McLinden et al., 1999; Fortin et al., 2003).

PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Our understanding of the factors controlling PSC particle number density is so poor that we have little basis for preferring its treatment in either the simple model or in WACCM, but our results from the simple model indicate a very strong sensitivity of the radiative impact of PSCs to particle number density. This provides a strong motivation for improved understanding of the physical mechanisms controlling ice particle nucleation and growth in the stratosphere.

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PSCs in a moist stratosphere

D. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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CPD

3, 935–960, 2007

PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

EGU

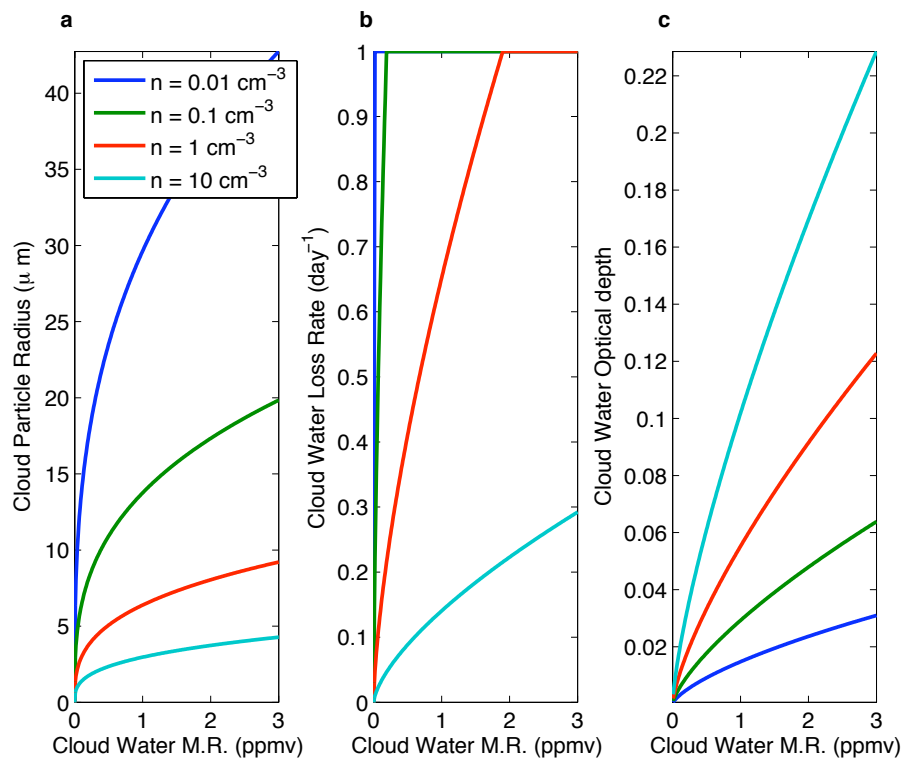
PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 1. Parameterized relationships between cloud water mixing ratio and **(a)** particle radius, **(b)** cloud water loss rate due to sedimentation and **(c)** cloud water optical depth for a 10 km deep region with uniform cloud water mixing ratio.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

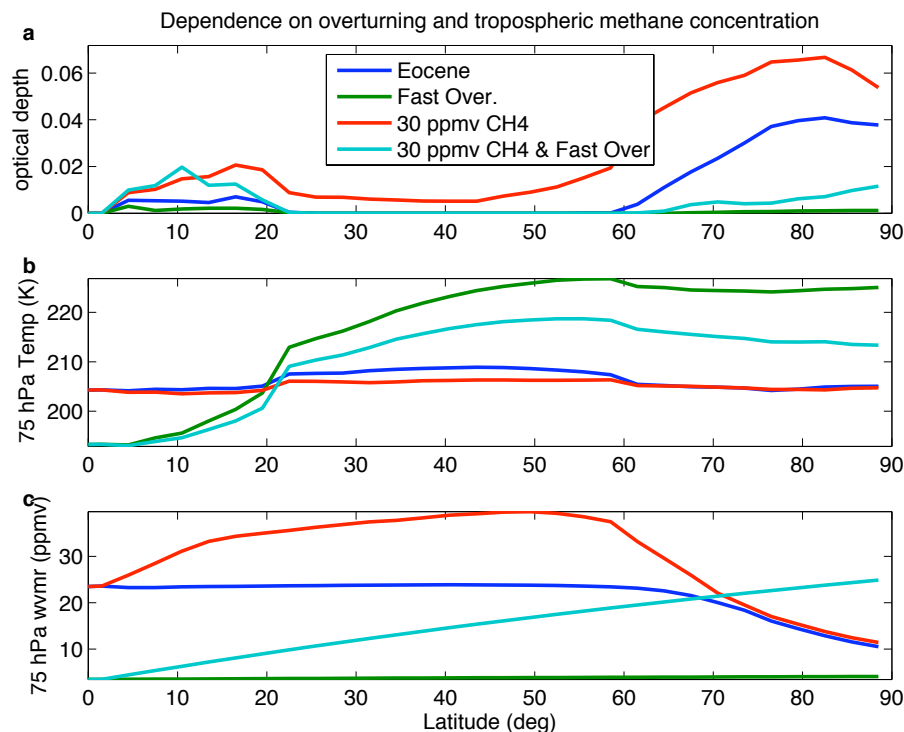
PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 2. Two-dimensional model results for total PSC optical depth **(a)**, 75 hPa temperature **(b)** and 75 hPa water vapor volume mixing ratio **(c)** for model experiments with varying tropospheric methane concentrations and varying overturning circulation strength. In all cases the surface temperature and static stability distributions are set to Eocene values (low meridional temperature gradient, low surface static stability gradient).

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

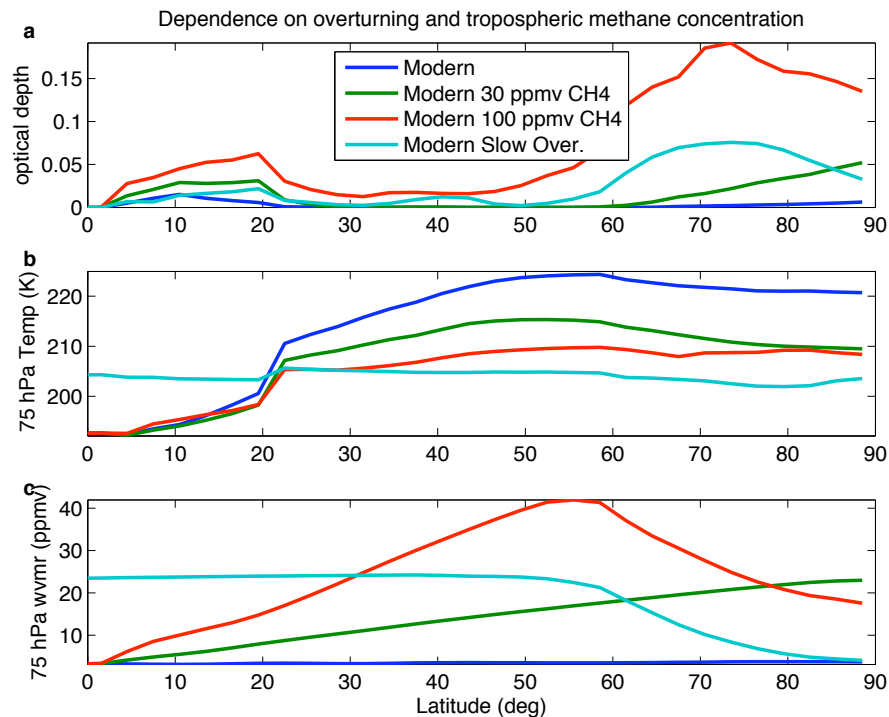
PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 3. As for Fig. 2, for the Modern case, for two cases with Modern surface temperature but elevated methane concentrations, and for the Modern case with slow (Eocene-level) overturning.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

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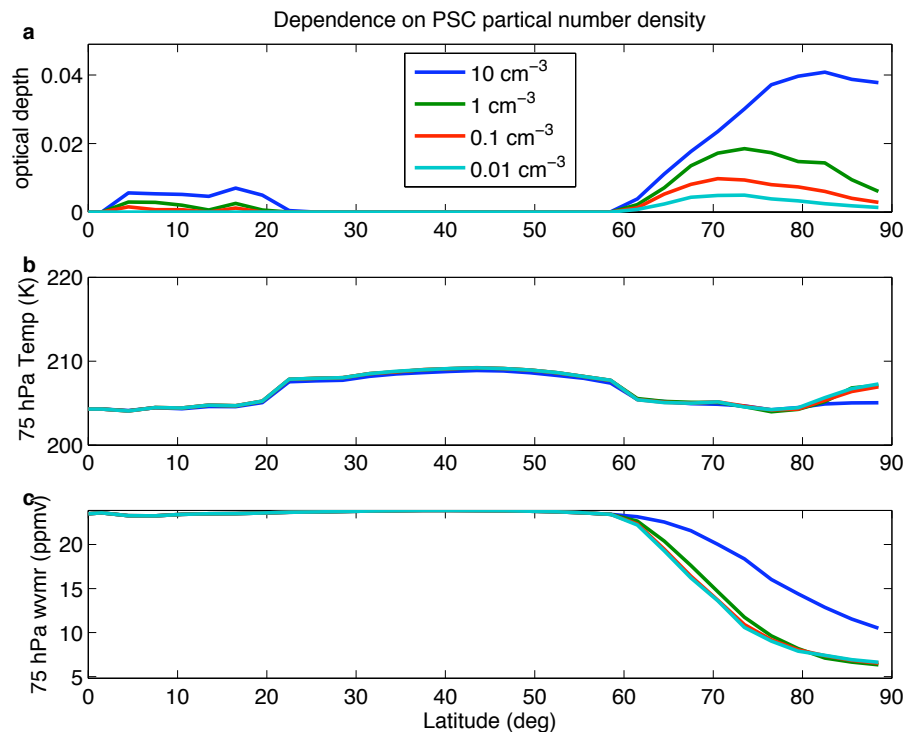
Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque**Fig. 4.** As for Fig. 2, for four different values of the PSC particle number density.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

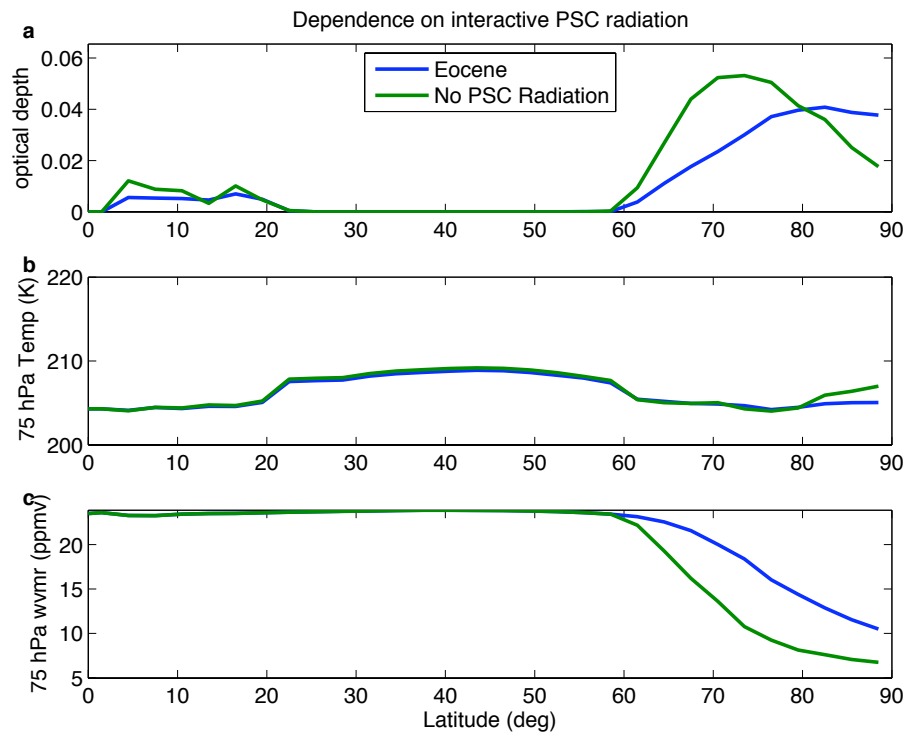
**PSCs in a moist
stratosphere**D. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 5. As for Fig. 2, for the control Eocene case, and for a case with PSC radiative effects turned off.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

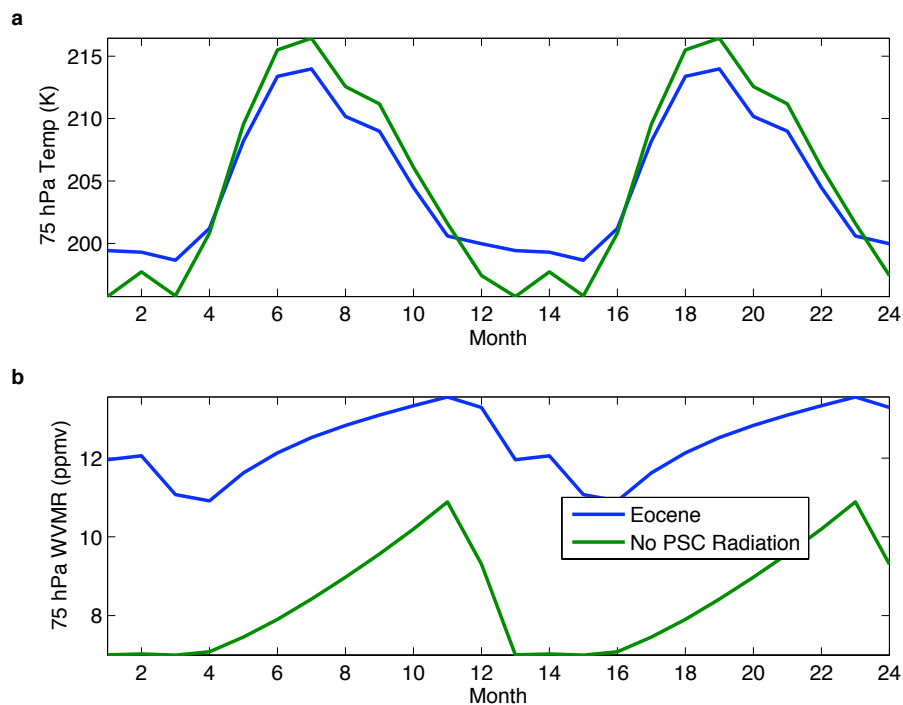
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J.-F. Lamarque

Fig. 6. Seasonal cycle of **(a)** 75 hPa temperature and **(b)** 75 hPa water vapor for the control Eocene case, and for a case with PSC radiative effects turned off.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

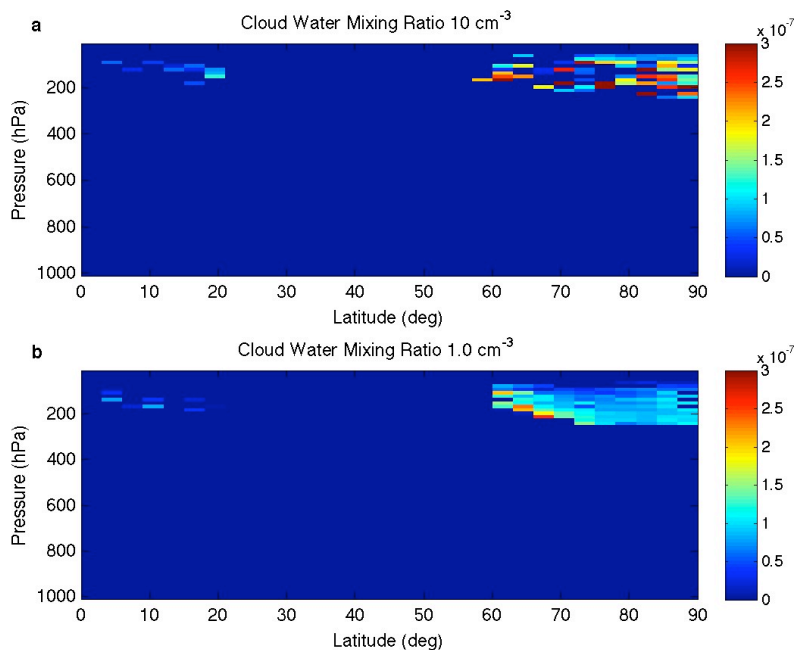
PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 7. Dependence of PSC uniformity on particle number density. Panel (a) shows August cloud water mixing ratio for PSC particle number density equal to 10 per cubic centimeter. Panel (b) shows August cloud water mixing ratio for PSC particle number density equal to 0.1 per cubic centimeter.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

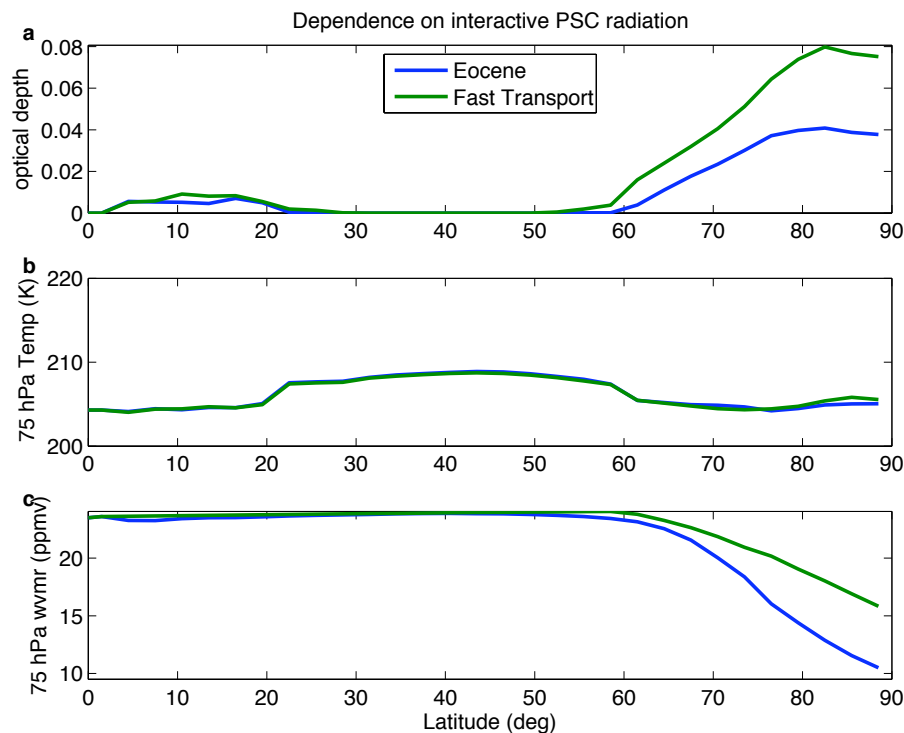
PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 8. As for Fig. 2, for the Eocene case, and for a the case of Modern transport.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

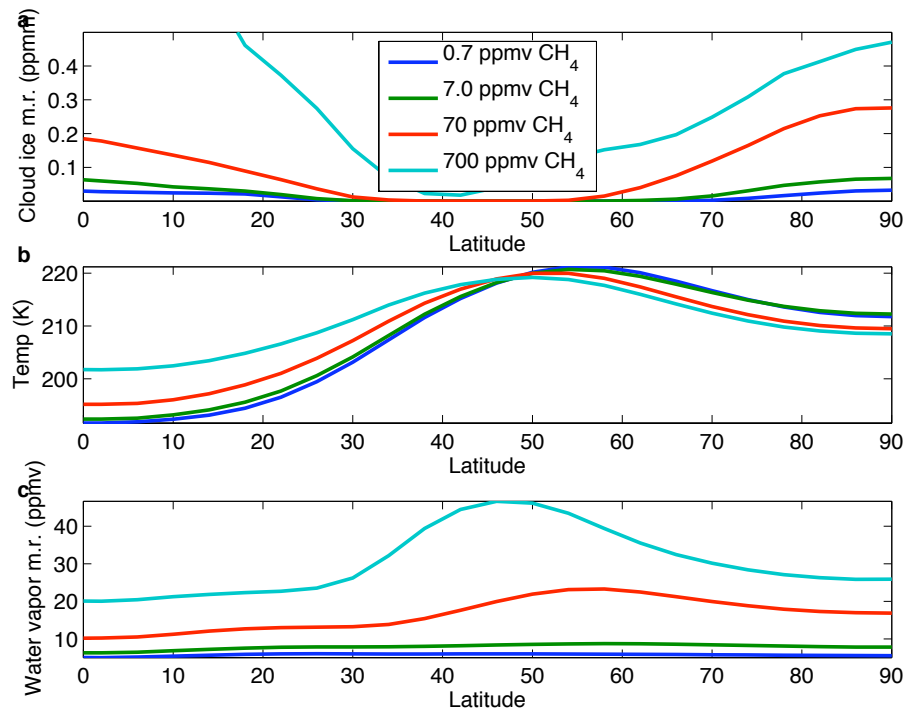
PSCs in a moist
stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 9. Results from three WACCM runs with Permian boundary conditions and varying tropospheric methane concentrations. **(a)** 75 hPa cloud water mass mixing ratio. **(b)** 75 hPa Temperature (K). **(c)** 75 hPa water vapor volume mixing ratio.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

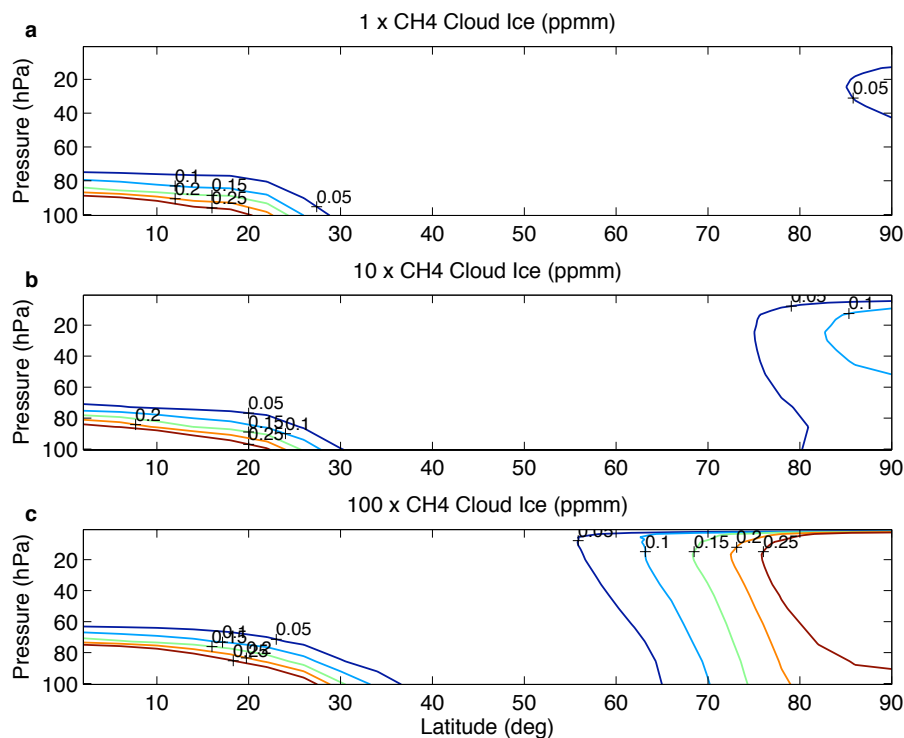
**PSCs in a moist
stratosphere**D. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 10. Contours of stratospheric cloud ice mixing ratio for four WACCM runs with Permian boundary conditions, with tropospheric methane concentrations of 1, 10, 100 and 1000 times pre-industrial concentrations of 700 ppmv. Contour interval is 0.05 ppmm of cloud ice.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

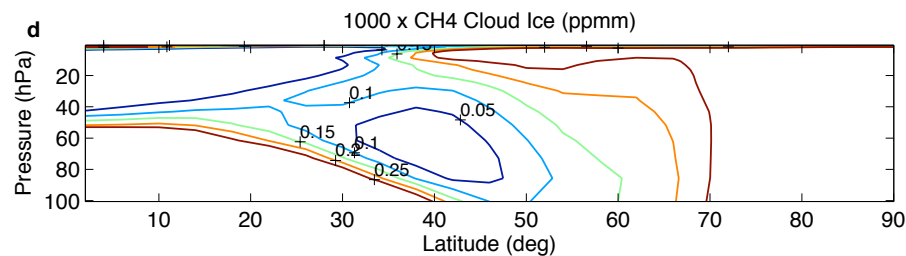
Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

PSCs in a moist stratosphereD. B. Kirk-Davidoff and
J.-F. Lamarque**Fig. 10.** Continued.[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

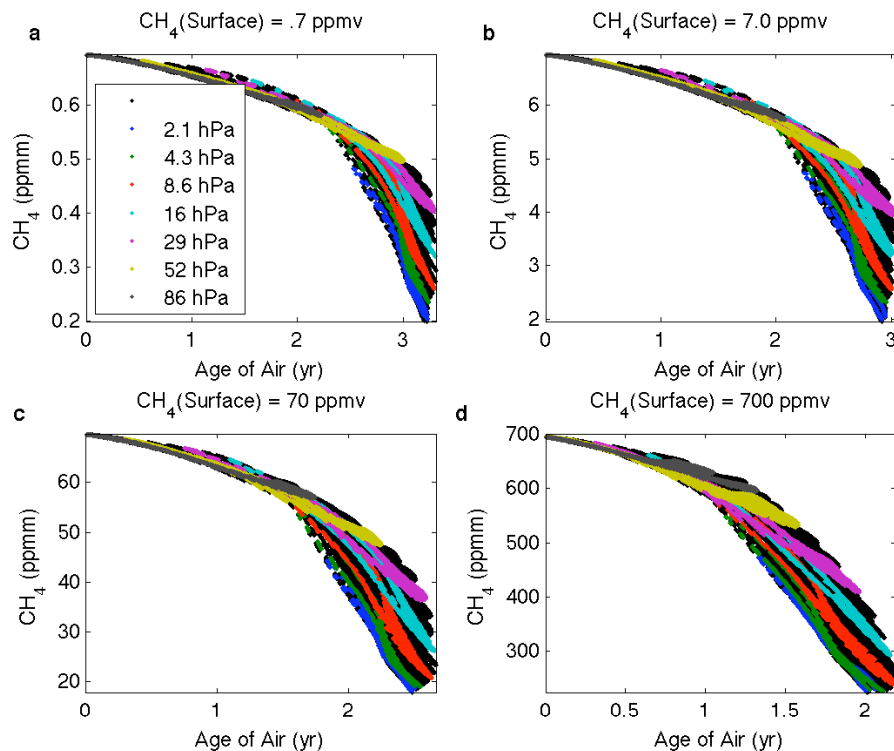
**PSCs in a moist
stratosphere**D. B. Kirk-Davidoff and
J.-F. Lamarque

Fig. 11. Mass mixing ratio of Methane (CH_4) plotted against age of air. Different model levels (labeled with their average pressure level) are denoted by different colors, as labeled in the legend.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)