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Simulations of desert dust and sea-salt aerosols in Antarctica with a general circulation model of the atmosphere

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ABSTRACT

A coupled aerosol/climate model, elaborated on the basis of a general circulation model of the atmosphere, is used to study the features of desert dust and sea-salt aerosols in the Antarctic region. Some of the observed seasonal characteristics of the two tracers are well simulated, and they are interpreted in terms of their relations with components of the atmospheric circulation. The model exhibits a strong influence of the boundary layer stability on the aerosols' vertical distributions near the surface. Observation at surface level may therefore be very misleading with respect to concentrations and seasonal variability higher in the atmosphere. An ice age experiment is run with the same aerosol/climate model, but fails to reproduce an expected large increase of dust and sea-salt concentrations in surface snow. The simulated enhancements of the production rates and atmospheric transport efficiency are weak. Changes in the distribution of the sources of dust, ignored in the experiment, could therefore have largely contributed to the last ice age dust increase. Sea-salt results suggest that the aerosol/climate model shortcomings are nevertheless not confined to source uncertainties, unless the complexity of the sources of sea-salt aerosol is also overlooked, for instance with respect to their relations with partial sea-ice cover.

1. Introduction

The relations between aerosols and climate are complex and poorly understood. Aerosols can affect the climate system (Hansen and Lacis, 1990) by interaction with the transfer of radiation through the atmosphere, with the formation of clouds and their radiative properties, and with the chemical and biological balance of the global environment. Aerosols are also very sensitive to climate and climate changes. Their life-time is typically a few days or a few weeks, which is similar to the time constant of the variability of meteorological features such as large scale turbulence, winds or precipitation. Also, because of such a short life-time, their distributions are very inhomogeneous and therefore sensitive to slower changes (e.g., seasonal or longer) of the general circulation. At last, aerosol production may depend on short and long term, natural or anthropogenic modifications of the environment.

For these reasons, aerosols can be used as tracers of climate and environment changes on a wide range of time scales. Those aerosols which can be transported over long distances and can therefore be observed in regions remote from their sources are particularly good tracers of the large scale circulation. Species that encounter less physical or chemical alteration during their transport than others are all the more appropriate for analyses focusing on circulation features. This is to some extent the case for mineral dust windblown over arid continental regions (desert dust) and for sea-salt aerosol produced by bubbling and foaming at the surface of the ocean.

Desert dust and sea-salt are complementary tracers in the sense that the former has continental sources while the later is of marine origin. However, the production of both aerosols is highly dependent on the surface wind stress. Their mixing and transport in the atmosphere and their deposition are a priori controlled by the same processes.

The same processes may not act with the same efficiency on the two species though. For instance, sea-salt particles, which are hygroscopic, are probably more sensitive than desert dust to deposition induced by liquid precipitation.

Time series of observed atmospheric concentrations of the two species in Antarctica (Cunningham and Zoller, 1981; Tuncel et al., 1989; Wagenbach et al., 1988) indicate that they are driven by, and therefore track, different features of the atmospheric circulation. Although data for sea-salt taken at different periods at the South Pole do not seem to agree (fig. 5 of Cunningham and Zoller, 1981 versus fig. 2 of Tuncel et al., 1989), the seasonal cycles of desert dust and sea-salt are clearly not in phase, and they are possibly in phase opposition (Cunningham and Zoller, 1981). Moreover, the seasonal cycle in surface snow, which is related to the cycle of deposition, may not have the same amplitude or phase as the cycle in the surface atmosphere (Thompson et al., 1975; Legrand, 1985; Petit, personal communication). At last, glacial-interglacial climate transitions, on time scales of several thousands of years, are associated with large changes of the concentration of the two species in the ice (Legrand et al., 1988; Lorius et al., 1984). These changes are much larger for desert dust than for sea-salt.

A modeling approach is used here to analyze the climatic parameters that drive the observed cycles of dust and sea-salt in Antarctica. A coupled aerosol-climate model is built by implementing aerosol parameterizations into an existing climate model (Genthon, 1992). Whenever the model fails to simulate the observed aerosol characteristics, possible shortcomings of either the aerosol or the climate part of the model are considered. Aerosol modeling uses simple concepts and empirical relations. Climate modeling is certainly more robust, but because aerosols are so sensitive, even minor problems with the simulated circulation can result in significant problems with the simulated aerosol distributions. Global climate models have exhibited deficiencies over Antarctica (Schlesinger, 1984), and the use of aerosol tracers may actually help analyze where and what some of these deficiencies are. It may also help for climates for which little data are available, such as the ice ages, and for the simulation of which aerosols data are in fact some of the strongest constraints.

When the model succeeds to simulate the observation, it is then relatively easy (i.e., easier than in the real world) to unravel to what climate variables and mechanisms the aerosols respond dominantly. It is likely that other atmospheric species, the cycle of which is in addition affected by chemical reactions, will respond in part to the same climate variables. The simpler aerosols used here as inert tracers help isolate these variables. At last, the model yields an extensive picture of the aerosol distribution in the atmosphere which, both in time and in space, goes well beyond the available local and sporadic observation. The model also suggests strategies of sampling that would provide a more complete and meaningful information on the aerosol atmospheric structure than surface concentration measurements alone.

2. Aerosol-climate model description

The base climate model is the GISS (Goddard Institute for Space Studies) Model II (Hansen et al., 1983, H83 from now on) General Circulation Model (GCM). The fact that the climate is simulated *in line* with the aerosol, unlike most other global tracer models, allows for a better account of many climate processes of importance for aerosols control: turbulence, convection, precipitation, etc. It would also allow the simulation of feedbacks of the simulated aerosol distributions on climate, although this is not considered here. On the other hand, running the climate along with the aerosol parameterizations makes computing costs correspondingly high. As a consequence, so far only the medium resolution version of the GCM (latitude × longitude = 7.82° × 10°, 9 layers in the vertical) could be used. The southern-most grid point is centered at the South Pole, and extends to the latitude 86° S. The next three rows containing 36 grid points each cover Antarctica entirely (Fig. 1).

Parameterizations for the sources, aerosol production over sources, turbulent transfer through the boundary layer (including turbulent dry deposition over non-source regions), vertical mixing through the depth of the atmosphere by convection, large scale transport by the general circulation, gravitational settling and wet removal have been implemented in the GCM. In the real world, these processes act at fairly small time scales (less than a day). The period at which the

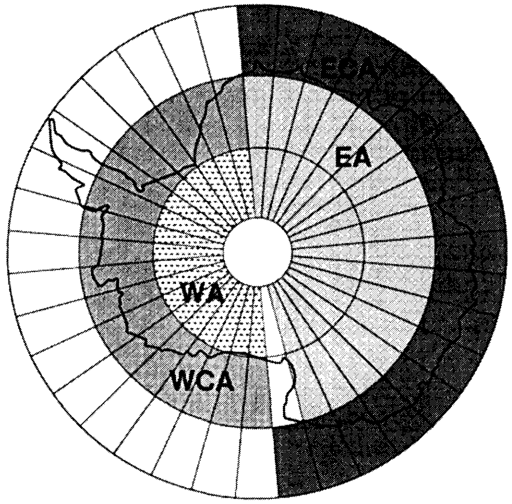


Fig. 1. Model grid over Antarctica. SP = South Pole. WA = west Antarctica (inland) = grid points 1 to 18 of row 2. EA = east Antarctica (inland) = grid points 19 to 35 of rows 2 and 3. WCA = west Coast of Antarctica = grid points 1 to 18 of row 3. ECA = east Coast of Antarctica = grid points 19 to 36 of row 4.

parameterizations are called in the model, which is the period at which the relevant climate variables are updated, is one hour or less. The reader is referred to Genthon (1992) for a more detailed description of these parameterizations and a presentation of global scale results. Because it is of crucial importance for the fate of aerosols in Antarctica, a limited description of the formulations used for aerosol production and deposition, including the treatment of the boundary layer, is given here.

2.1. *Boundary layer turbulent transfer parameterization*

The boundary layer over most of Antarctica is rather peculiar compared to the rest of the world: it is generally stable because of a temperature inversion particularly steep during winter. Transfer through the boundary layer is therefore strongly inhibited. This is of importance for long range transported species, since transport occurs in the interior of the atmosphere and not in the boundary layer. For such species, sedimentation is weak, and therefore what is observed at the surface is mostly what could diffuse through the stable boundary layer. It should consequently be less than above in the bulk atmosphere.

A two step parameterization is used to simulate this transfer. An intermediate level, called "surface level," is defined a few meters above the ground. The diffusion from the first dynamic level of the climate model (level 1) to the surface level (level s) uses a traditional κ-diffusion expression:

$$F_{1s} = \rho\kappa \frac{Q_1 - Q_s}{Z_{1s}} + G_{1s}. \tag{1a}$$

The flux between level s and the ground (level g) is parameterized with a bulk aerodynamic expression:

$$F_{sg} = \rho C_d W_s (Q_s - Q_g) + G_{sg}. \tag{1b}$$

ρ and W_s are the density and the velocity of air at surface level, computed by the GCM. C_d and κ, the drag and turbulent diffusion coefficients, are also provided by the GCM interactively. Z_{1s} is the altitude of level 1 above ground ($\sim$150–200 m), and G is a Stokes gravitational term. Q_i is the aerosol concentration at level i ($i = 1$, s or g). Q_g is set to 0: the aerosol is immediately and irreversibly captured when it reaches the ground.

The surface fluxes of heat, moisture and momentum in the climate model use a similar two-step parameterization (H83). The climate model does simulate a temperature inversion across the flux layer (between the surface and level 1) in Antarctica. The strength and the seasonal cycle of this inversion at the South Pole (Fig. 2) is compatible with observation, which support that

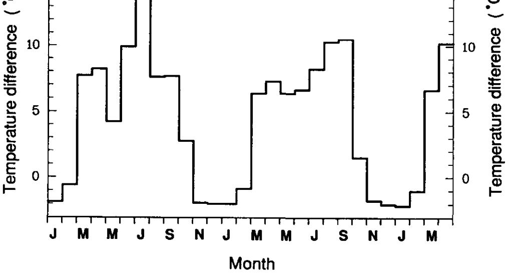


Fig. 2. Difference of temperature between the first dynamic level of the model and the surface level at the South Pole.

surface turbulent parameters are reasonably well simulated. Note that this is only part of the total inversion strength, that is the temperature difference between the warmest model layer (generally higher than level 1) and the ground (Fig. 4, to follow).

2.2. *Aerosol production*

Over sources (deserts for dust, free ocean for sea-salt), the parameterization for the injection of aerosol in the atmosphere also uses eq. (1). For dust, the concentration at ground level is empirically parameterized as:

$$Q_g = CW_s^3, \tag{2}$$

based on the work of Gillette (1981), and Nickling and Gillies (1990). For sea-salt, the concentration at the surface level (level s) is directly parameterized using:

$$Q_s = e^{aW_s + b}, \tag{3}$$

with $a = 0.16$ and $b = 1.45$ for W_s less than $15\,\mathrm{m\,s^{-1}}$, and $a = 0.13$ and $b = 1.89$ otherwise (Erickson et al., 1986). For sea-salt, only (1a) is necessary to compute the flux into the interior of the atmosphere. The size distribution for the marine aerosol is given by (Erickson and Duce, 1988):

$$\mathrm{mmr} = 0.422\,W_s + 2.12, \tag{4}$$

where mmr is the mean mass radius of a log-normal distribution with geometric standard deviation 3. For dust, C in Eq. (2) is adjusted a posteriori to fit an observed averaged distribution at sea (Prospero, 1981).

2.3. *Wet deposition*

In each grid box where condensation occurs (condensation scavenging) or which is swept through by precipitation from above (collision scavenging), the quantity of aerosol scavenged is expressed by:

$$\Delta Q = Q_i(1 - e^{-\alpha P}). \tag{5}$$

Q_i is the mass of aerosol initially in the grid box, P is the quantity of precipitation involved ($\mathrm{kg\,m^{-2}}$), and α is an adjustable parameter. α is high for nucleation scavenging by both liquid and solid precipitation for sea-salt, but it is high only for solid precipitation for dust. Collision removal efficiency is low for particles of size around 1 μm, and comparatively high for larger and smaller particles (Radke et al., 1980). Values of α used in the present simulation are given in Table 1.

Table 1. *Scavenging coefficient* α ($m^2\,kg^{-1}$). α_l/α_2: α_l *is for liquid precipitation* ($T > -10°C$), α_2 *is for solid precipitation*

Aerosol	Particle	α nucleation	α collision
desert dust	P1 ($R = 0.2\,\mu$m)	0.0/2.3	0.7/0.7
	P2 ($R = 1.0\,\mu$m)	0.0/2.3	0.2/0.2
	P3 ($R = 5.0\,\mu$m)	0.0/2.3	2.3/2.3
sea-salt	P1 ($R = 1.0\,\mu$m)	2.3/2.3	0.2/0.2
	P2 ($R = 5.0\,\mu$m)	2.3/2.3	2.3/2.3
	P3 ($R = 25.0\,\mu$m)	2.3/2.3	2.3/2.3

3. Simulation results for modern climate

The aerosol-climate model was run for 27 months, starting on December 1st with an aerosol free atmosphere, and using boundary conditions (sea surface temperatures, insolation, etc.) of modern climate. The size distribution of the long range transported fraction of the two aerosols is schematically represented by 3 independent size bins. Radii are 0.2, 1 and 5 μm for desert dust, and 1, 5 and 25 μm for sea-salt (25 μm sea-salt particles are insignificant in Antarctica, but they are very significant at sea and therefore on a global scale). It appears that one month is a long enough spin-up time for complete dispersion of the aerosol in the atmosphere. Therefore, time series are shown starting from the first month of January. Seasonal averages are computed using all available triads of months relevant to each season (e.g., winter = average of the two series of June-July-August available from last 26 months of the run). Here, all seasons are southern hemisphere seasons.

Time series at the South Pole describe the evolution of the aerosols in the polar grid box only (row 1 of the model). Coastal diagnostics are averaged over the western-most 18 grid boxes of row 3 for west Antarctica, and the eastern-most 18 grid boxes of row 4 for east Antarctica (Fig. 1).

H83 provide a thorough analysis of the global climate simulated by the GISS GCM at the resolution used here. Observations used as reference are not always the most appropriate for Antarctica (for which region comparatively little data are available). Before aerosols results are presented, a quick re-analysis of the climate model in and around Antarctica is performed. Some additional diagnostics likely to be of importance for aerosols distributions in Antarctica are also presented.

3.1. GCM simulated climate of the Antarctic region

The atmospheric surface temperature over the Southern ocean is very strongly constrained by the prescribed sea surface temperature. It is therefore not surprising that it comes fairly well out of the model (H83, Fig. 21). On the other hand, the surface temperature over the Antarctic continent is too warm by about 5° on an annual mean (reliable observations are presented in Radok et al. (1986) for instance). Temperature in Antarctica is a delicate balance between heat advection, for a large part by the mean meridional circulation (Nakamura and Oort, 1988), and radiation loss. H83 (Fig. 44) present simulated transports of energy to the high southern latitudes which are correct over Antarctica, although too strong at lower latitudes. However, observations of winds

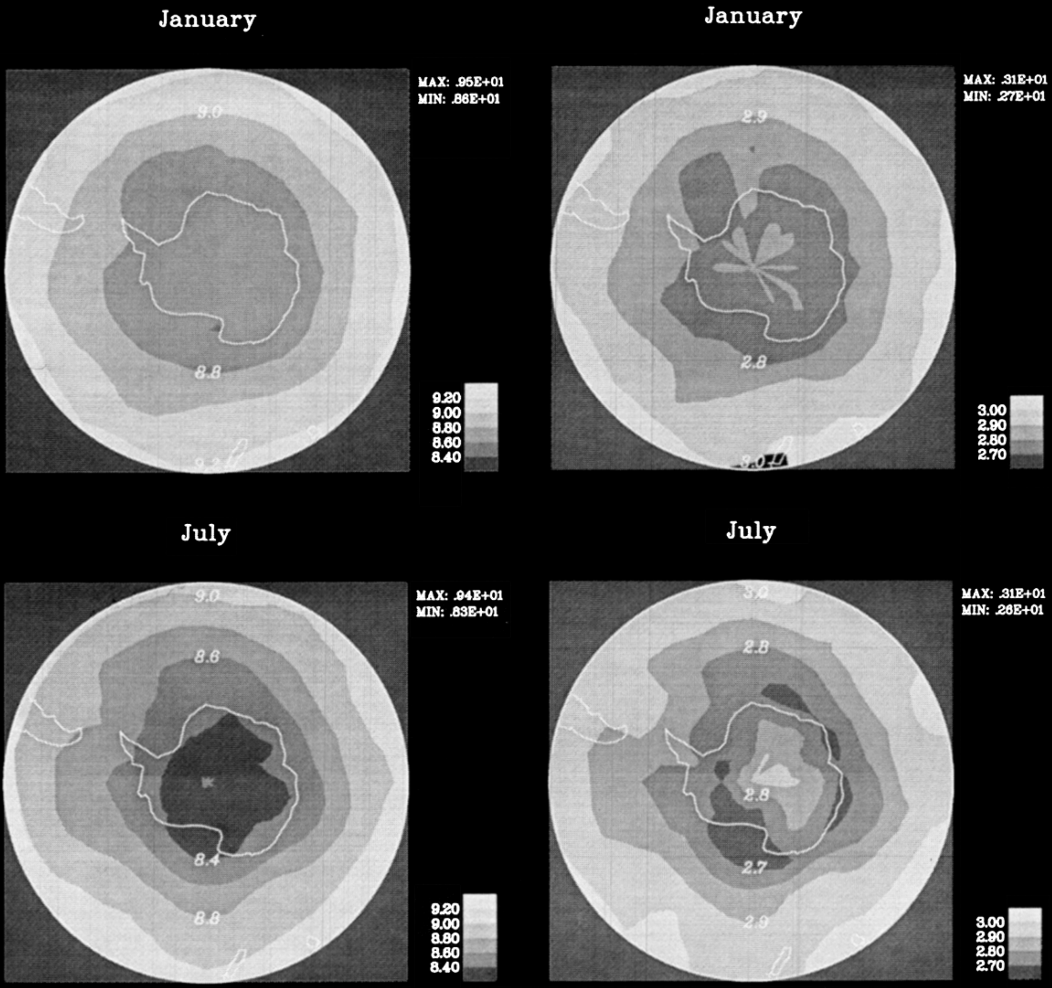


Fig. 3. Geopotential topography (km) at the 300 (left) and 700 (right) mb levels in January (upper plots) and July (lower plots).

and therefore transports over the Antarctic region are scarce and doubtful (Nakamura and Oort, 1988). Observations of surface albedo are also questionable, but data used as reference by H83 (Fig. 23) suggest that albedo is somewhat low in the model. If true, this may explain some of the temperature discrepancy.

Due to the elevated Antarctic topography, the atmospheric circulation in and around Antarctica is usually analyzed from diagnostics of the geopotential height at low pressure levels, such as 300 mb or 100 mb (Schwerdtfeger, 1984). The structure of the geopotential at the 300 mb level, presented on Fig. 3 for January and July, looks correct (although the mean is about 200 m too high everywhere). The strong deepening of the Antarctic depression during winter, which reflects an enhancement of the circulation intensity, is well reproduced by the model. Fig. 3 also shows the geopotential height at the 700 mb level, which is very close to the surface of the ice sheet in the interior of the continent. The model simulates a series of lows around the continent which looks particularly reasonable in winter.

Over Antarctica, the model surface winds (Fig. 30 of H83) may be compared, at least for direction, to the wintertime streamline estimate of Parish and Bromwich (1987). The low model resolution does not allow a detailed account of the ice sheet topography. As a consequence, the drainage of the downslope streams which results in very strong but localized coastal currents is not simulated. On the other hand, in the continental interior, the general wind directions seem reasonable. On a larger scale, for our purposes, the surface wind is more important for its intensity (over source regions) than for its direction. The wind is generally much stronger in the southern hemisphere during local winter than during summer. Simulated precipitation looks reasonable over the southern ocean (Fig. 28 of H83). Over Antarctica, in spite of some disagreements between the available observations (Giovinetto and Bull, 1987), precipitation is probably too strong in the

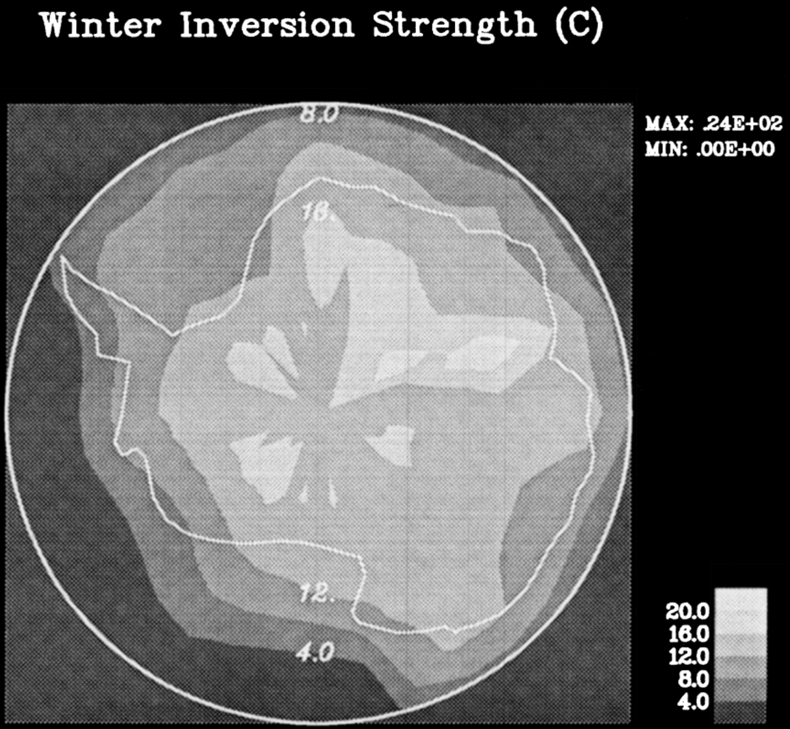


Fig. 4. Winter inversion (temperature difference between the warmest model layer and the surface, the unit is °C).

model by a factor of two. This might be related to the atmosphere being too warm and therefore carrying too much water vapor.

At last, the surface temperature inversion has been suspected (and will be confirmed) to affect significantly the transfer of particles through the boundary layer to the surface level. Schwerdtfeger (1984) reports inversions as strong as 20°C at the South Pole, and even higher values on the east Antarctic plateau. Fig. 4 presents the simulated winter inversion over Antarctica, that is the temperature gradient between the warmest layer of the model and the surface. A large part, but not all, of the inversion is confined between the first atmospheric level and the surface (Fig. 2). The simulated inversion is probably too weak in the continent interior (15° instead of 20° at the South Pole) while it is possibly too strong in coastal regions. One can nevertheless expect that a great deal of the consequences of the real inversion will be captured by the model.

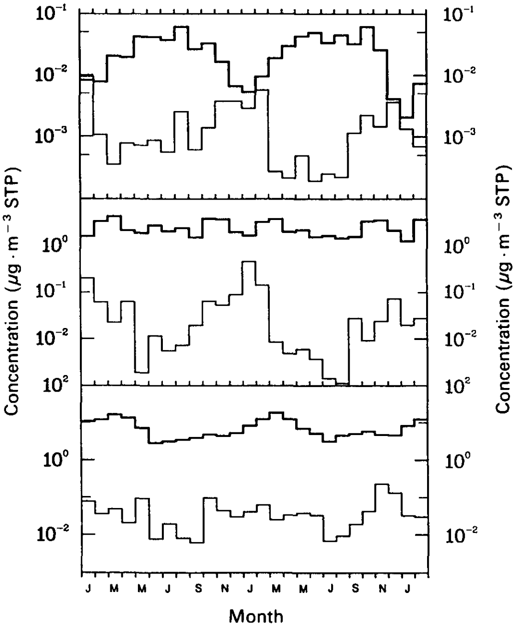


Fig. 5. Desert dust (thin) and sea-salt (thick) atmospheric concentration at surface level at the South Pole (upper) and coastal west (middle) and east (lower plot) Antarctica.

3.2. Atmospheric aerosols concentrations

Fig. 5 presents the time series of desert dust and sea-salt in the atmosphere at the surface level, i.e., near the bottom of the boundary layer. This diagnostic should be representative of actual measurements of dust and sea-salt atmospheric concentrations, which are made typically a few meters above the ground. Table 2 presents the annual, winter and summer mean values. The model simulates a very different behavior for the two aerosols. Desert dust shows high concentrations during late spring and summer at the South Pole and at both coasts, and in fact all over Antarctica. The cycle is in agreement with observations, but mean values appear too low at South Pole and possibly too large at the coasts (Cunningham and Zoller, 1981; Wagenbach et al., 1988). Sea-salt shows highs during winter at the South Pole, in agreement with some observations (Cunningham and Zoller, 1981). At the coast, no cycle is simulated in west Antarctica, and a weak seasonal cycle with highs in fall is simulated in east Antarctica. Wagenbach et al. (1988)'s observations at G. Von Neumayer station (70°37′S, 8°22′W) show a weak and irregular seasonal cycle, but with higher values more systematically during fall or winter.

Table 2. *Atmospheric concentrations at surface level: SP = South Pole, WCA = west coast of Antarctica, ECA = east coast of Antarctica*

Surface concentration (ng.m^{-3}STP)	Desert dust			Sea-salt		
	SP	WCA	ECA	SP	WCA	ECA
annual	1.5	48	46	28	2400	7500
winter	0.77	5.1	14	44	1900	3800
summer	3.0	140	55	5.8	2300	8600

Table 3 presents annual, winter and summer mean concentrations vertically averaged over the 3 lowest atmospheric layers of the model (up to ~2000 m above surface), above the flux layer. At the South Pole, the seasonal cycle of desert dust is, if anything, opposite to the cycle at surface level. On the other hand, coastal sites still show higher concentrations during summer, but the amplitude of the cycle is much weaker than at the surface. Things are just opposite for sea-salt: at the South Pole, the same cycle is found above the boundary layer as at the surface, but with a larger amplitude and larger averaged values. Over the coast of east Antarctica, the cycle of sea-salt is inverted and has highs during winter rather than summer. At last, concentrations over the coastal west Antarctica, which shows no cycle at the surface, varies higher in the atmosphere in phase with the other regions of Antarctica.

It is therefore clear that the surface is a particular region of the atmosphere of Antarctica and that, if the model is right, surface level data are not typical of the composition in the bulk atmosphere. This is clearly due to the presence of an inversion layer which prevents homogenization of the air composition down to the surface. The steepest part of the temperature inversion in the model occurs between the surface level and the first dynamic level, which is therefore the region of strongest mixing inhibition. Fig. 2 shows that at the South Pole, this inversion can be as large as 15°C on a monthly average in winter (the total inversion reaches almost 20°C), while it is close to 0°C in summer. The model also simulates deep winter inversions (up to 8°C) between the surface and level 1 at the coast.

The cycles of aerosol production and of efficiency of the atmospheric transport also contribute to modulate concentrations in Antarctica. It should actually drive the vertically averaged aerosol loads. Fig. 6 shows the winter and summer production of desert dust and sea-salt in the model, which mostly respond to winds intensity. The seasonal contrast of production is higher for sea-salt than for dust. Combined with stronger meridional transport and in spite of wet removal being more efficient due to larger precipitation rates, higher production drives higher sea-salt concentrations in the winter atmosphere of Antarctica. The cycle of stability of the boundary layer then affects the sea-salt concentrations at the surface, compared to higher levels in the atmosphere: It somewhat reduces the amplitude of the cycle at the South Pole, yet it does not control it (it has a strong effect on mean concentrations though). On the other hand, it has a large enough impact in coastal regions to cancel the effect of production (dominated by local production, with a cycle amplitude weaker than in regions of highest production to which South Pole may be more sensitive) in west Antarctica and to reverse the cycle in east Antarctica.

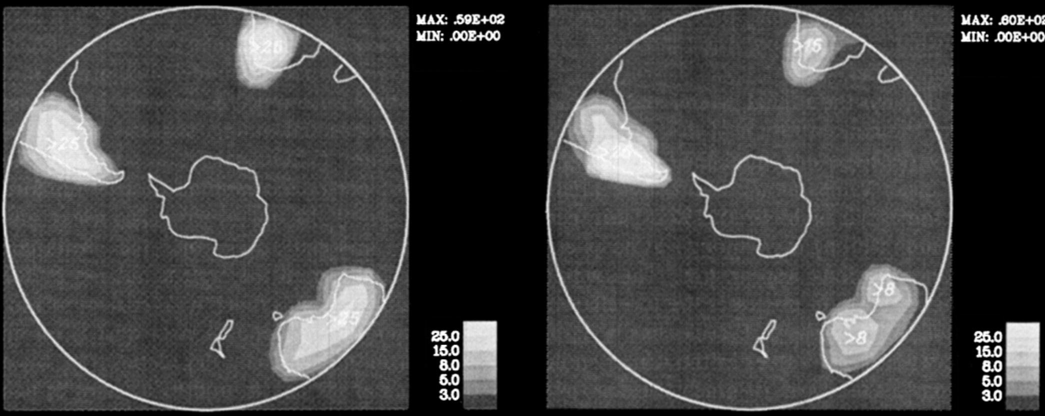
Production and other remote effects also impact significantly on desert dust in the atmosphere of Antarctica. They drive higher summer concentrations in the low level atmosphere at the coasts. Inland however, because the dust that cannot reach the surface accumulates above the boundary layer, concentrations are higher there in winter. The cycle of surface concentration at the South Pole is therefore primarily driven by the cycle of the surface atmosphere stability, and the concentrations above the boundary layer are also affected although in the opposite direction.

Note that some simulated aerosol features would improve if, as discussed in the presentation of the GCM produced climate, the strength of the inversion in winter were weaker in coastal regions and stronger at the South Pole. For instance, coastal sea-salt concentrations at surface level could

Table 3. *Low level (up to ~2000 m above surface) atmospheric concentrations*

Low level concentration ($ng.m^{-3}STP$)	Desert dust			Sea-salt		
	SP	WCA	ECA	SP	WCA	ECA
annual	27	300	300	310	4200	6300
winter	41	150	170	580	6200	7600
summer	20	510	370	25	1400	3200

Dust Austral Winter Dust Austral Summer



Sea–Salt Austral Winter Sea–Salt Austral Summer

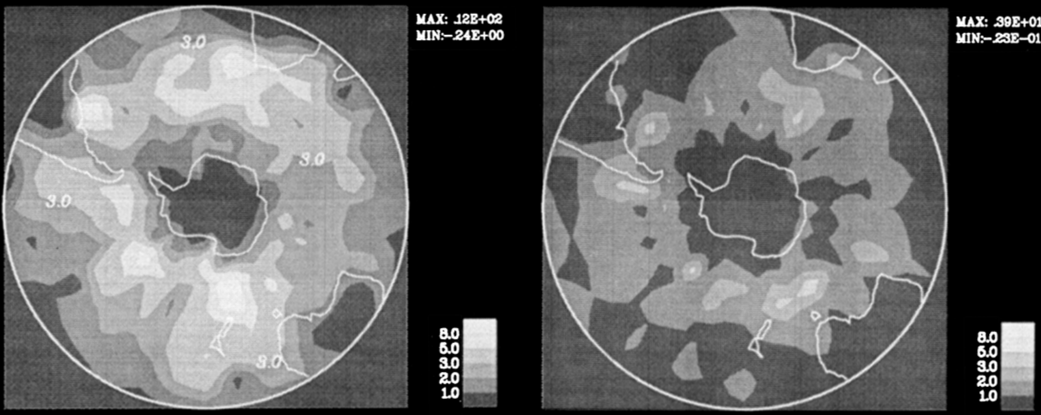


Fig. 6. Production (mg m^{-2} h^{-1}) of desert dust (upper plots) and sea-salt (lower plots) aerosols during austral winter (left) and summer (right). Modern continents boundaries are for localization purposes, but the fractional treatment of the surface in the model (Hansen et al., 1983) allows production to occur beyond these boundaries.

then be clearly higher in winter, even in west Antarctica. It may also make simulations fit better with recent observations by Tuncel et al. (1989) of higher sea-salt concentrations at the South Pole in spring rather than in winter (but it would conversely make the agreement with Cunningham and Zoller (1981)'s observations looser).

3.3. *Concentrations in surface snow*

For many species, and in particular for desert dust and sea-salt, the snow-covered surface is a natural sampler of the Antarctic atmosphere. Measurements can be easier there than in the atmosphere, and historical records can be obtained by sampling snow pits and ice cores. However, the atmospheric composition can be deduced from snow or ice measurements only if the air-to-surface transfer function is known. This function involves mainly the processes of dry deposition and precipitation induced deposition (so-called "wet" deposition, although this is not very appropriate for solid precipitation). The fraction that deposits through dry process records information of atmospheric concentrations close to the surface, while the wet deposited fraction is more relevant of concentrations in the interior of the atmosphere. Which process dominates controls what information the snow or ice composition carries.

Table 4 displays the annual, winter and summer averages of the concentration of desert dust and sea-salt in surface snow as simulated by the model. For dust, mean values at the South Pole are, if anything, somewhat high (Briat, 1974; Petit, personal communication) even though atmospheric surface concentrations appear low. This would suggest that deposition processes are too efficient in the model. The seasonal cycle has a large amplitude with higher values in the winter snow. At the coast, the cycle is much weaker and opposite. Annual mean values are an order of magnitude larger than at the Pole, which is consistent with higher concentrations in the atmosphere. Sampling by Mumford and Peel (1982) at 70°53′S and 64°57′W in the Antarctic Peninsula plateau may be representative of coastal values. They find that the number of insoluble microparticles per unit mass of snow is roughly 5 to 15 times larger than observed at the South Pole (Mosley-Thompson and Thompson, 1982; Petit, personal communication), which would be consistent with model results. However, Boutron (1971) and Briat (1974) report rather homogeneous distributions of the terrigeneous components in snow over Antarctica, except very close to the coast.

Because measurements are difficult due to the low quantities involved, and because reprocessing of the surface snow can alter the seasonal information, attempts to recover seasonality are of limited reliability in central Antarctica. The observed seasonal amplitude of dust in snow at the South Pole, which might be smoothed by post-deposition processes ignored in the model, is smaller than simulated (Petit, personal communication). However, at Byrd (80°15′S, 119°31′W), Thompson et al. (1975) observe cycles which they attribute to seasonal variations, with an amplitude compatible with what the model simulates. Comparison with the isotopic record, which gives an approximate seasonal timing, suggests that highs occur in winter (Thompson et al., 1975) or in spring (Petit, personal communication). Mumford and Peel (1982) do not observe any clear and systematic seasonal cycle in the Antarctic Peninsula, and indeed the cycle simulated by the model at either coast is much weaker than inland.

In the model, wet deposition of dust is on an annual mean higher than dry deposition by a factor of ten at the coasts, and by a factor of two at the South Pole. Wet deposition still dominates at the coasts on a seasonal basis, but at the South Pole dry deposition is larger than wet removal by a factor of three in summer. This is understandable in view of the much higher surface concentrations during the warm season. It is more difficult to understand why wet deposition is an order of magnitude larger in winter than in summer. Dust concentration above the inversion layer, where much of the condensation occurs, is higher in winter but only but a factor of two (Table 3). Precipitation is lower in winter by a factor of two. The size distribution of particles changes with season with relatively more large particles (P3 in Table 1) in winter, and P3 particles are more efficiently collected by hydrometeors. However, large particles at South Pole at any time contribute for at best only one tenth the total dust concentration, in agreement with observations from central Antarctica (Mounier, 1988).

It is possible that in winter dust and moisture are advected to central Antarctica together, so that precipitation events correlate better with high quan-

Table 4. *Concentrations in surface snow = ratio of total aerosol deposition over precipitation*

Concentration in snow ($ng\,g^{-1}H_2O$)	Desert dust			Sea-salt		
	SP	WCA	ECA	SP	WCA	ECA
annual	60	540	560	1600	15000	25000
winter	110	360	370	3600	20000	29000
summer	28	680	500	67	7400	15000

tities of dust in the atmosphere than in summer. This hypothesis cannot be checked now because model diagnostics were saved on a monthly mean basis to limit storage burden. A re-run specifically made to address this question, with a higher storage frequency, would be necessary.

Annual mean sea-salt concentrations in surface snow at South Pole are 30 × higher than those for dust. This does not look unreasonable if one considers that atmospheric concentrations themselves are typically an order of magnitude higher, and that more heavy particles which sediment more easily can reach the Pole since the sources are less distant. Indeed, on an annual mean for sea-salt, dry deposition is almost as efficient as wet deposition.

However, observed concentrations of sea-salt in surface snow are much lower than what the model simulates. At South Pole, Legrand (1985) reports values of the order of a few tens of ng g^{-1}H$_2$O, which is at least one order of magnitude less than what the model produces. Moreover, the model simulates concentrations at the coast which are one order of magnitude higher than at South Pole, while observations by Legrand (1985), Aristarain et al. (1982) or Mumford and Peel (1982) suggest that values in sites slightly remote from the coast are barely higher than at the South Pole. In the model, most coastal grid boxes contain a large fractions of ocean (Fig. 1). Simulated atmospheric concentration and deposition values, which are averaged over entire grid boxes, are therefore very much influenced by the oceanic environment and may not be representative of regions even just a little inland. The grid box that contains the German coastal station G. V. Neumayer, which belongs to row 3 of the model, is mostly over land but contains a small fraction of ocean. The concentrations simulated in this particular grid box are in very good agreement with the observations by Wagenbach et al. (1988), for annual mean concentrations both in the surface snow ($\sim$16 mg kg^{-1}H$_2$O) and in the surface atmosphere ($\sim$1.5 μg m^{-3}).

Seasonal cycles of sea-salt in surface snow simulated by the model are questionable to the extend that mean concentrations seem strongly overestimated. However, the model does simulate a very large cycle at the South Pole, and a weaker one at the coasts, in all cases with highs during winter. Legrand (1985) reports a seasonal cycle, and approximate timing obtained from the water isotope record suggests that highs occur during late winter. The amplitude observed by Legrand is much smaller than what the model simulates, but as for dust, the signal may be smoothed by reworking of snow after deposition. Mumford & Peel (1982) report a seasonal cycle in the Antarctic Peninsula, close to the coast, which is weaker and less systematic than what Legrand observes at the South Pole. The model also simulates a weaker and fuzzier cycle in coastal regions, although it is still stronger than observed. In the grid box that contains the G. V. Neumayer station, the model simulates a clear seasonal cycle with highs in late fall and winter. This is again in good agreement with data from Wagenbach et al. (1988).

At the South Pole, in the model, the annual mean rates of dry and wet deposition of sea-salt are approximately equal. As for dust however, dry deposition dominates in summer even though precipitation is more important and atmospheric surface concentrations are lower. Again the process of wet deposition appears more efficient in winter, possibly because of a correlation between the occurrences of high aerosol concentrations and precipitation.

Altogether, it seems that the model simulates rates of deposition which are too high over Antarctica. Pourchet et al. (1983) estimate that the dry fallout of radioactive particles is 15 to 25% of total deposition in coastal regions, and 40 to 60% in Central Antarctica. However, Legrand et al. (1988) find that deposition of sea-salt in Central Antarctica occurs mostly by dry processes. Wet deposition in the model may therefore be too strong compared to dry deposition. It is not excluded that in fact both depositions are overestimated. The simulated total deposition looks more reasonable on a global scale (Genthon, 1992), and it therefore appears that the problem is confined to the high latitudes, possibly for wet removal in relation with the particularly low rates of precipitation in such regions.

A complementary study of the sensitivity of the model to weaker rates of aerosol deposition is thus desirable. Meanwhile, one can speculate that, by increasing the species mean life time, less deposition over Antarctica would make the spatial horizontal distributions more homogeneous, hence in better agreement with some observations. Overestimated dry or wet deposition probably affect the vertical distribution in the low levels of the atmosphere in opposite directions. Dry deposition removes aerosol from surface level and therefore, if too large, accentuates the gradient of concentration through the boundary layer. On the other hand, if wet deposition is too efficient, too much aerosol is removed from the top of the inversion, where the air is warmer and moister and where therefore most of the precipitation originates from. The same precipitation could also scavenge some aerosol below while sweeping through the boundary layer, and predicting the impact of changing the wet removal efficiency is not straightforward.

Beyond these uncertainties, the model suggests that the aerosol system in Antarctica has more complexity than sampling in surface snow and surface atmosphere alone can address. Atmospheric surface sampling is not representative of the aerosol distribution through the depth of the atmosphere. Sampling of surface snow provides a mixture of information on surface and bulk atmospheric concentrations which cannot easily be deconvolved. Moreover, the model suggests that seasonal changes of the relative efficiency of dry and wet removal makes information at time scales lower than a year even more difficult to unravel.

4. Simulation results for an ice-age climate

The same aerosol-climate model was used to simulate global aerosol distributions for an ice-age climate. Proxi-data from an ice age, such as aerosols concentration in polar ice, often provide strong signals which are nevertheless hard to translate in terms of climate and environment changes. GCMs have been used in some instances to study and refine over speculations of the response of such proxis to climate changes (e.g., Joussaume et al., 1989). In turn, the simulated proxis put direct constraints on the quality of the modeled circulation.

For this ice age experiment, the sources of sea-salt aerosol were changed only to the extent that the free ocean surface is reduced due to sea level lowering and increased sea-ice cover. It is likely that the aridity of the planet changed and on the averaged increased during the last ice-age, and that as a consequence the regions of production of desert dust moved and expanded very significantly (Sarnthein, 1978; Kutzbach, 1990). Little of the details of these changes are well known though. The same fractional distribution of dust sources over continents was used for both modern and ice-age experiments, so that mostly (but not quite only, see next) the response of dust distributions to glacial-interglacial changes of the components of the circulation is tested here.

As for the modern experiment, the model was run for 27 months, but with the boundary climate conditions of the last glacial maximum (CLIMAP 1981). An expansion of the surface of the continents due to lowering the sea level is taken into account. For grid points in which the continental fraction increases, the dust source area increases in proportion. This has a rather minor effect on the overall sources extent and distribution. No new ice age dust source is set for grid points that did not contain any source in the modern experiment, even if their continental fraction becomes positive or increases in the ice age run. Because most of the information about ice-age aerosols in Antarctica was obtained from deep ice cores drilled inland (Byrd (80°01′S, 119°31′W) in west Antarctica, and Dome C (74°39′S, 124°31′E) and Vostok (78°28′S, 106°5′E) in east Antarctica (Lorius et al., 1984)), diagnostics are averaged over the western-most 18 grid points of row 2 of the model for the west Antarctic plateau, and the next 17 grid points of row 2 and 3 for east Antarctic plateau (Fig. 1). Time resolution in deep ice cores does not generally reach the season, due to the very low accumulation rates, and aerosol results are shown as annual means averaged over the last 24 months of the run.

4.1. Simulated climate change for an ice age

Hansen et al. (1984) and Rind (1986) present a rather extensive comparison of the modern and ice age climates simulated with the GISS model at the resolution used here. Again, these analyses are not particularly focused on the southern latitudes and Antarctica. Results for this particular region are summarized here and additional diagnostics are presented.

Surface temperature is globally 3.6°C lower for an ice age, but it is ~5°C lower in Antarctica (Hansen et al., 1984, Fig. 9). This is about half the cooling estimated from ice core isotopic data on the plateau (Jouzel et al., 1989). In the high southern latitudes, there is little change in the intensity of the eddy transport of energy (Rind, 1986, Fig. 12–13). Because the available energy (for instance sensible heat) at one given latitude decreases (since it is cooler), this may indicate some enhancement of the energy transport efficiency of the atmosphere. However, the very moderate change in the geopotential topography at 300 mb (Fig. 7) suggests that there is little increase in the intensity of the circulation (see also Figs. 7 and 9 of Rind (1986), for latitudes higher than 45° S).

We will see, from diagnostics of aerosol production, that the intensity of the surface wind is not much larger in the simulated ice age, at least over source regions for aerosols. Precipitations in the southern hemisphere are somewhat less intense (Rind, 1986, Fig. 2), but one would not expect such a small change to affect very significantly the aerosol transport efficiency of the atmosphere. There are some evidences that over the east Antarctic plateau, the accumulation of ice was about half its rate nowadays (Jouzel et al., 1989), which is close to what the GCM simulates in this region.

Fig. 8 shows the ice age minus modern climate change of the strength of the surface inversion over the Antarctic region in winter. The inversion becomes weaker near the South Pole, remains unchanged or increases in coastal regions, and increases very significantly further away, where sea ice builds up during the ice age. As mentioned above, all the information available on ice age aerosols is from the east and west Antarctic plateaux. The simulated change in the inversion strength is minor there, and would affect little the way the boundary layer controls vertical distribution and deposition processes. On the other hand, it may prevent some of the dry deposition of the aerosols before they reach the continent, and therefore contribute to enhance the transport efficiency to Antarctica.

4.2. *Aerosols results for an ice age*

The only direct information ice-cores can provide is the concentration in surface snow at the

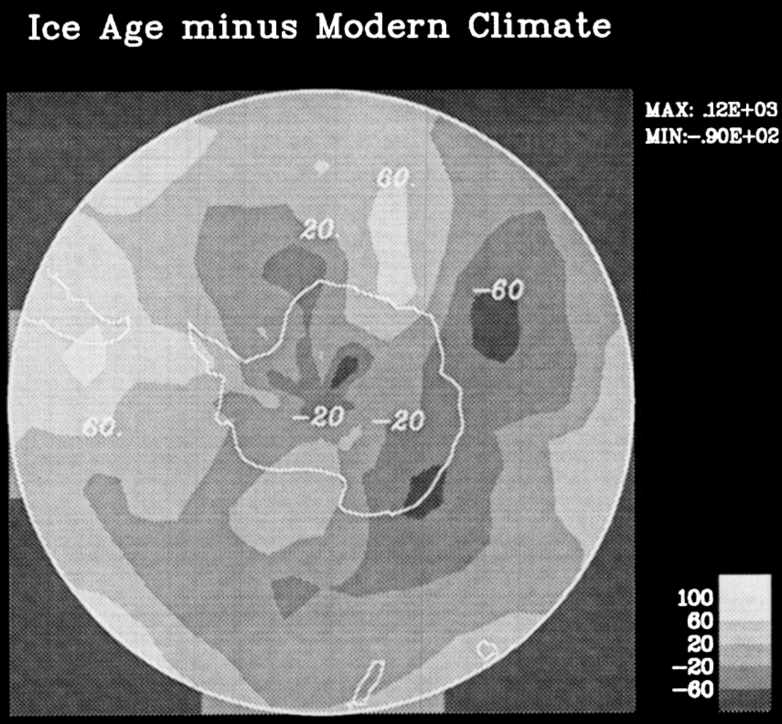


Fig. 7. Ice age minus modern climate 300 mb geopotential topography (m).

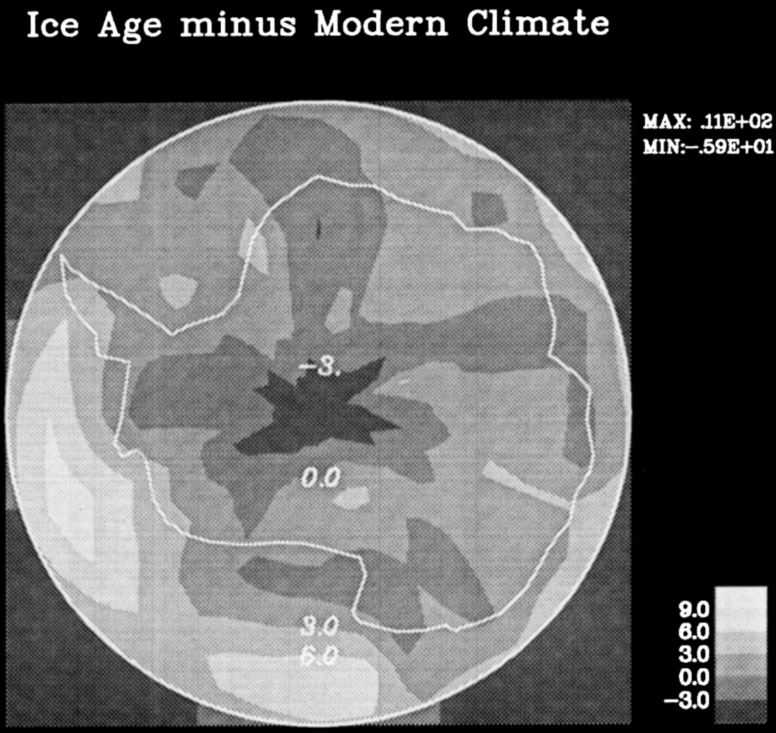


Fig. 8. Ice age minus modern climate strength of the winter inversion (°C).

time of deposition. It is admitted that this information is well preserved over time in the ice for many species, and in particular for components of desert dust and sea-salt aerosols. Sampling of these components reveal that in ice from the last ice age, desert dust and sea-salt concentrations are much larger than what is observed in recent ice or surface snow (Lorius et al., 1984).

Table 5 shows the ice age over current climate ratios of annual mean concentrations of dust and sea-salt in the atmosphere and in surface snow, as simulated by the model over the west and east Antarctic plateaux. The model barely simulates an increase of dust in snow. Elemental determinations and particle counting indicate an ice age increase by a factor of about 4 in Byrd ice core (Lorius et al., 1984; Thompson and Mosley-Thompson, 1981), and a factor of about 20 in Vostok and Dome C ice cores (Lorius et al., 1984; Petit et al., 1990). Model simulated concentrations are therefore much too low, which is not totally surprising since the probable enhancement of dust production in response to an increased continental aridity during an ice age is not taken into account. The model suggests that a relatively minor fraction of the observed higher concentrations in ice from east Antarctica is linked to changes of atmospheric circulation components only.

Table 5. *Ice age over modern climate ratio of annual mean concentration of aerosol in the surface and low level atmosphere and in the surface snow; WA = west Antarctica plateau, EA = east Antarctica plateau*

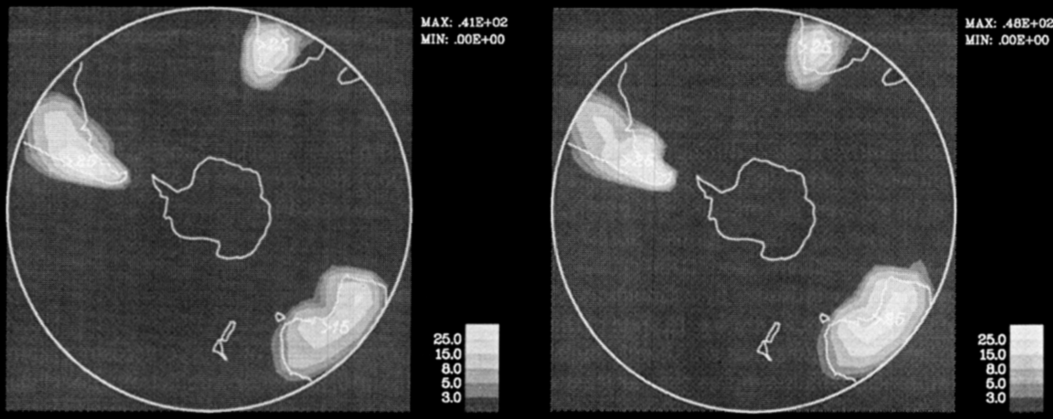
Ratio (ice age)/ (modern)	Desert dust		Sea-salt	
	WA	EA	WA	EA
surface atm. concentration	1.2	2.0	0.70	1.2
low level atm. concentration	0.85	1.9	0.55	1.0
concentration in snow	1.1	2.2	0.70	1.3

Desert dust concentrations in the atmosphere are also generally higher during an ice age, by a factor similar to the factor of increase in the surface snow. This suggests that there is no significant local enhancement of deposition processes efficiency. Production of dust (Fig. 9) is not much higher in the ice age experiment than in the modern run. Averaged over the Southern sources, it increases very little, more in response to continental expansion than to stronger surface winds. The increased transport efficiency of the glacial atmosphere, associated with a stronger meridional eddy transport and weaker precipitation scavenging, accounts therefore for increasing dust concentrations in Antarctica by a factor of less than two. If the model is right, expansion or migration of dust sources, in association with corresponding changes of global aridity and lowering of sea level, must account for 95% (east Antarctica) to 100% (west Antarctica) of the observed increased ice age concentrations.

The same conclusion is reached by Joussaume (1990) who, with a similar approach, also fail to simulate any significant increase of dust in

Dust Modern Climate

Dust Ice Age Climate



Sea-Salt Modern Climate

Sea-Salt Ice Age Climate

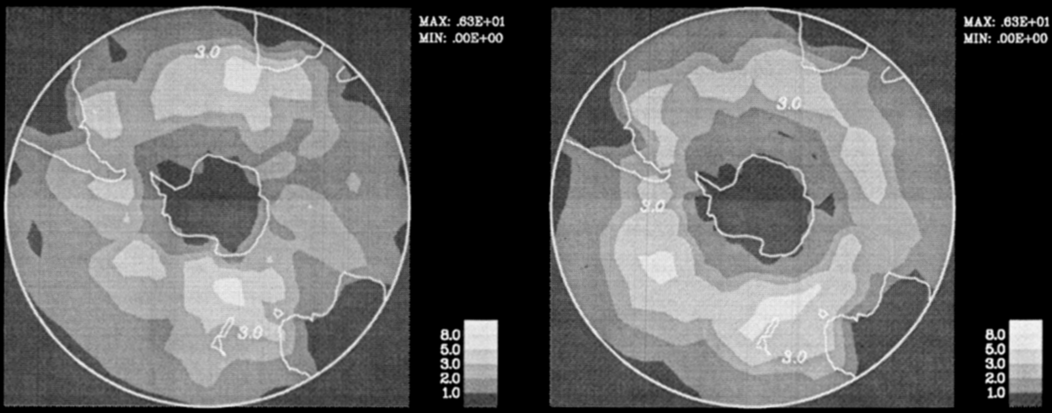


Fig. 9. Annual mean production (mg m^{-2} h^{-1}) of desert dust (upper plots) and sea-salt (lower plots) for modern (left) and ice age (right) climates.

Antarctica for an ice age. Sensitivity experiments by Gillette (1992) with a zonally averaged dust/climate model confirm that changes in source regions is probably the single most important parameter which can explain the observed enhanced dust deposition in polar regions. It would be interesting to run with the present model similar sensitivity experiments in which sources of dust are arbitrary moved or expanded.

The ice age experiment results in concentrations of sea-salt in the surface snow which are lower than for modern climate in west Antarctica, and barely higher in east Antarctica (Table 5). Observations in ice cores (Lorius et al., 1984) indicate an increase everywhere. However, this increase is more moderate for sea-salt, with a factor of 2–3 in Byrd ice core (west Antarctica) and 4–5 in Vostok and Dome C ice cores (east Antarctica), than for dust. Simulated concentrations are therefore about 4 × too low everywhere. Production of sea-salt aerosol over the Southern Ocean is not significantly higher during an ice age (Fig. 9), showing again that surface winds and other boundary layer parameters are relatively unchanged. However, the regions closest to Antarctica where significant production occurs are moved toward lower latitudes by as much as 15° because of the expansion of sea-ice. In spite of this, the atmospheric concentrations are slightly higher in east Antarctica, which support that at least in this region the simulated transport efficiency of the ice age atmosphere increases.

Even though the model fails to simulate the observed increased concentrations in surface snow for both dust and sea-salt, it is not as far off for the later. If sea-salt characterizes a problem with either the airborne aerosol (i.e., not considering the source definition component) or the climate part of the model, which would affect dust similarly, then the contribution of changes of the sources of dust during an ice age still comes out as very significant (e.g., source changes must still contribute for at least half the concentration increase in east Antarctica). Definition of the sources of sea-salt may not be as straightforward as it seems though, and the complex role of leads in sea-ice and of polynias, totally overlooked here, may be critical for all kinds of exchanges between the sea surface and the atmosphere (Andreas et al., 1984). This is all the more potentially important that leads and polynias may occur near the coast of Antarctica, in response to katabatic winds for instance, and they are thus not affected by the expansion of sea-ice during an ice age.

It is also possible that the modeled circulation is not quite as good as desirable for simulating such highly sensitive tracers. It has been pointed out that for energy, which is not as sensitive because it is more smoothly distributed, GCM produced transports can be off, even for modern climate, by as much as a factor of 2 (Stone and Risbey, 1990). That atmospheric meridional transports and surface winds barely increase during an ice age is unexpected in view of the higher baroclinicity of the atmosphere. However, we have found in the previous section that the parameterizations for airborne aerosol in the model, which have a much more empirical basis than the climate counterpart, sometimes provide questionable results. This is the case for simulated removal efficiencies for instance, and because precipitation regimes change during an ice age (an other point that the climate model may not simulate well), changes of removal efficiency along the transport to Antarctica may be incorrectly accounted for.

5. Conclusions

The implementation of simple aerosol parameterizations into a climate prediction model has helped address questions that observations of desert dust and sea-salt in Antarctica raise about the current and past interactions between climate and aerosols. The present study brings some elements of answer to these questions and some interesting conclusions, but it also points to model components that appear most sensitive and should deserve further attention.

Among the most striking results, the model indicates that the way the Antarctic atmosphere is monitored nowadays, by sampling only the surface atmosphere and the deposited snow, does not bring all the information necessary to understand these interactions. The vertical structure of the aerosol distribution is a particularly important parameter, and what is seen at the surface, for instance in terms of seasonal cycle, may be opposite to what is happening above and therefore in the column as a whole. Because the model simulates that dust and sea-salt have different seasonal production and vertical distribution, they were very appropriate tracers together to point to this problem, and in general for such a modeling approach of the aerosol/climate system.

Observations of a very large increase of dust deposition in Antarctica during the last ice age have raised the hypotheses that (i) the source regions expanded and migrated, due to the extension of the continental margins and/or changes of continental aridity, (ii) the production of dust over sources increased due to increased wind stress, and/or (iii) the transport efficiency of the atmosphere toward the high latitudes increased. To the extent that the model is right, results presented here show that (ii) and (iii) contributed in fact little, and (i) is left as the single most likely interpretation.

Although more moderate than for dust, the ice age changes for sea-salt in Antarctica are nevertheless quite significant. The sea-salt aerosol is therefore a tracer complementary to dust for paleoclimate experiments as well. Because it is generally felt that sources of sea-salt are simple and well understood, only hypotheses (ii) and (iii) have generally been considered to interpret the enhanced deposition of sea-salt during an ice age. In the model presented here, these mechanisms cannot overcome the negative contribution of the source parameter if source regions follow the ice age expansion of sea-ice around Antarctica. As a consequence, the model as a whole does not simulate an increased sea-salt deposition in Antarctica. Either the climate or the aerosol part of the model (or the two) are wrong enough to misrepresent some processes essential for explaining the paleoclimate observations. It is possible that the complexity of the actual sources of sea-salt and of the way fractional cover of sea-ice impacts on them were overlooked, here and in previous interpretations of the data.

The results presented here are a starting point to modeling the relations between aerosols and climate, and they should in the future be complemented with sensitivity experiments that test for the most tentative aspects of the aerosol model. Deposition and sources of both dust and sea-salt are high priority candidates for such sensitivity experiments. There are several conceptually simple studies, some mentioned in the text, for which the set up cost (but, unfortunately, not the computing cost) would be moderate and which would most likely bring further insight in the problem.

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