PLANKTONIC DIMETHYLSULFIDE AND CLOUD ALBEDO: AN ESTIMATE OF THE FEEDBACK RESPONSE

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Abstract. Partial control of climate by the biosphere may be possible through a chain of processes that ultimately links marine plankton production of dimethylsulfide (DMS) with changes in cloud albedo (Charlson *et al.*, 1987). Changes in cloud optical properties can have profound impacts on atmospheric radiation transfer and, hence, the surface environment. In this study, we have developed a simple model that incorporates empirically based parameterizations to account for the biological control of cloud droplet concentration in a first attempt to estimate the strength of the DMS-cloud albedo feedback mechanism. We find that the feedback reduces the global climatic response to imposed perturbations in solar insolation by less than 7%. Likewise, it modifies the strength of other feedbacks affecting surface insolation over oceans by roughly the same amount. This suggests that the DMS-cloud albedo mechanism will be unable to substantially reduce climate sensitivity, although these results should be confirmed with less idealized models when more is known about the net production of DMS by the marine biosphere and its relation to aerosol/cloud microphysics and climate.

1. Introduction

In many respects our understanding of the role of clouds in the climate system and in climatic change is wholly inadequate. Through their effect on radiation, clouds can strongly influence the heating and cooling of the planet. As climates have varied in the past, cloud properties may also have varied, but in ways that are poorly known. Only in the last decade have exploratory studies begun to test hypotheses as to whether changes in cloud distributions and cloud radiative properties act as positive or negative feedbacks to climatic perturbations (e.g. Somerville and Remer, 1984; Wetherald and Manabe, 1988; Roeckner, 1988; Mitchell *et al.*, 1989).

In this study we focus on a feedback mechanism involving a metabolic product of marine plankton, dimethylsulfide (DMS), and its effects on the scattering of solar radiation by clouds. According to a hypothesis proposed by Charlson *et al.* (1987), changes in environmental conditions affect the production by marine plankton of dimethylsulfide, which is a major source of cloud-condensation nuclei (CCN) in the marine troposphere. As a source of CCN, DMS may have significant

Climatic Change **18**: 1–15, 1991.

effects on cloud optical properties, and therefore climate. They argue that through this process, which we will call the DMS-cloud albedo feedback, partial biological regulation of climate may be possible.

Charlson *et al.* (1987) considered the parameter dependence of the feedback and estimated the climatic response to prescribed changes in CCN. However, they were unable to estimate the strength of the feedback in a climate change scenario because they had no way of describing the response of the biosphere to changes in surface climatic conditions. Measurements of seasonal and regional variations in the atmospheric flux of DMS, condensation nuclei concentrations, and climatic conditions by Bates *et al.* (1987) have now made this possible.

In contrast to many of the better-known climate feedbacks (e.g. temperature-water vapor feedback, ice-albedo feedback), the DMS-cloud albedo feedback, as currently understood, does not directly depend on changes in temperature. According to Bates *et al.* (1987), the feedback appears to involve a relationship between surface insolation, DMS flux, CCN concentration, and cloud albedo. Thus the feedback does not become active in direct response to a change in temperature, but only in response to modifications in surface insolation. The feedback may therefore be directly initiated by 'external' forcing that perturbs surface insolation over the oceans, such as changes in atmospheric aerosol concentration (including volcanic aerosol, arctic haze, and tropospheric dust), variations in Earth's orbital parameters, and changes in the solar constant itself.

Additionally, and perhaps more importantly, the DMS-cloud albedo feedback will modify the climatic response to *any* kind of external forcing, as long as a part of the climatic response involves changes in surface insolation. For an increase in atmospheric CO_2, for example, part of the climatic response is likely to be a change in cloud distribution and cloud liquid water content, which clearly affects surface insolation. Consequently, although changes in atmospheric CO_2 do not appreciably affect surface insolation directly, the DMS-cloud albedo feedback could still be an important component of the climatic response. If the changes in cloud distribution and cloud radiative properties are considered to be a feedback (as opposed to a direct response of climate to changes in forcing), then the DMS-cloud albedo feedback might be described as a 'feedback on a feedback'. Even so, it should not be presumed that the DMS-cloud albedo feedback is necessarily of secondary importance, since it operates on feedbacks that may by themselves alter climatic responses by as much as 100% (e.g. Somerville and Remer, 1984; Schlesinger and Mitchell, 1987; Hansen *et al.*, 1984).

To estimate the strength of the DMS-cloud albedo feedback, we have formulated a simple model based upon empirical relationships between surface insolation, DMS flux from the ocean to the atmosphere, and CCN concentration. Section 2 describes this model and the various parameterizations used in it. To gauge the direct effects of the feedback to changes in external forcing, we have introduced a perturbation to the solar constant and calculated the changes in surface insolation with and without the DMS-cloud albedo feedback in Section 3. In Section 4, we have considered the importance of the feedback in relation to how it may modify the effects of other cloud-related climate feedbacks. In particular, we discuss how this feedback may operate during a hypothetical CO_2 warming scenario. We then conclude with a summary and review of the limitations of this exploratory study.

2. Model Description

Figure 1 illustrates the components and processes represented in our model of DMS-cloud albedo feedback. An imposed increase in surface insolation produces an increase in the flux of DMS to the atmosphere according to a relationship reported by Bates *et al.* (1987). The oxidation products of DMS contribute an increased concentration of CCN and, under the proper conditions, an increased density of cloud droplets. As the number density of cloud droplets increases, the ability of the cloud to scatter solar radiation is enhanced, thus reducing the imposed change in surface insolation. According to this model, the DMS-cloud albedo feedback is therefore negative.

Each of the processes outlined above can be described mathematically, and together these comprise our model as described in the next several subsections. It is of some interest to note that unlike many climate feedbacks, the DMS-cloud albedo feedback, as presently understood, does not involve temperature. There is

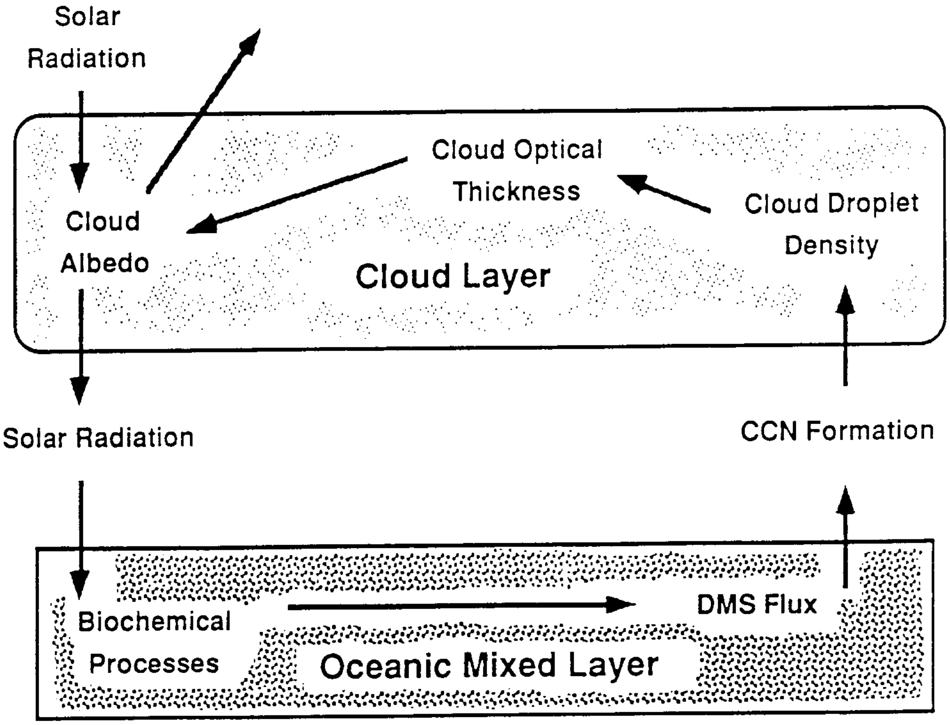

Fig. 1. A schematic representation of DMS-cloud albedo feedback.

therefore no need to consider longwave radiation. The strength of the DMS-cloud albedo feedback can be measured in terms of changes in surface insolation. It is natural then to consider first the processes directly affecting surface insolation.

2.1. *Solar Radiation*

To represent the solar radiation at the surface, we assume that absorption of solar radiation within the atmosphere is negligible so that the net downward radiation at the surface is

$$Q = (1 - \alpha_p)\,\frac{\mu S_o}{2}\,. \tag{1}$$

Here μ is the cosine of the solar zenith angle, S_o the solar constant, and α_p is the planetary albedo which, accounting for multiple reflections between clouds and surface, can be expressed as

$$\alpha_p = \alpha_c + \frac{(1 - \alpha_c)^2\alpha_s}{1 - \alpha_c\alpha_s}\,, \tag{2}$$

where α_s is the surface albedo and α_c the cloud albedo. An expression for the cloud albedo can be derived from the two-stream approximation for conservative scattering (Lacis and Hansen, 1974),

$$\alpha_c = \frac{\sqrt{3}(1 - g)\delta}{2 + \sqrt{3}(1 - g)\delta}\,,$$

which, for an asymmetry factor $g = 0.85$ appropriate for cloud droplets, reduces to

$$\alpha_c = \frac{\delta}{7.7 + \delta}\,, \tag{3}$$

where δ is the optical thickness of the cloud. This relationship between cloud albedo and optical thickness is presented in Figure 2. Although (3) was chosen because it is a particularly simple formulation, we have found that our calculations are fairly insensitive to this parameterization. This conclusion was reached after trying alternative relationships between cloud albedo and optical thickness (e.g. Slingo, 1989; Wiscombe, 1977).

2.2. *Cloud Radiative Properties*

The most important effects of clouds on solar radiation can be accounted for by their optical thickness, δ. For a given liquid water content, the ability of a cloud to scatter solar radiation is largely determined by the total droplet surface area (Twomey, 1977). According to Mie theory, when the circumference of cloud droplets is much larger than the wavelength of incident radiation, the extinction efficien-

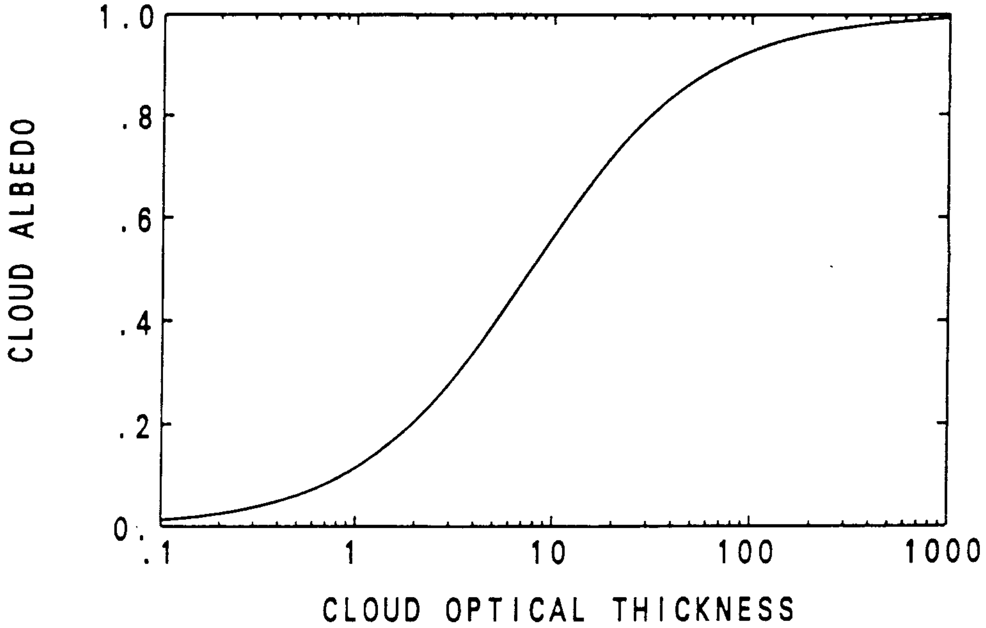

Fig. 2. Cloud albedo dependence on cloud optical thickness, δ. For the nonabsorbing cloud considered here, the albedo increases from 0 to 1 as the cloud optical thickness increases. The slope of the curve (rate of change of cloud albedo with $\log(\delta)$) is largest for an optical thickness of about 8.

cy approaches the asymptotic value of two. Therefore, the extinction optical thickness of a cloud can be expressed as (Stephens, 1978)

$$\delta = \frac{3wh}{2\rho r_e}, \tag{4}$$

where w is the liquid water concentration, h the cloud thickness, r_e the effective cloud droplet radius, and ρ the density of water. The effective radius for a distribution of cloud droplets is defined as the ratio of the mean cubed droplet radius to the mean square droplet radius (Hansen and Travis, 1974).

Given the cloud water concentration and the effective radius of the droplets, the effective number density of cloud droplets, N, can be defined such that

$$w = \frac{4}{3}\pi r_e^3 \rho N. \tag{5}$$

According to this definition, a collection of N cloud droplets per unit volume, each with a radius r_e, has the same water concentration as the actual polydisperse distribution of droplets found in the cloud.

By substituting (5) into (4), we obtain an expression for cloud optical thickness as a function of N, h, and w:

$$\delta = h \left[\frac{9\pi w^2 N}{2\rho^2} \right]^{1/3} \tag{6}$$

To isolate the effects of DMS-cloud albedo feedback, the cloud thickness and cloud liquid water concentration are held constant as N responds to changes in CCN concentration caused by changes in the DMS flux to the atmosphere.

2.3. *Cloud Droplet Number Density*

The next link required by the DMS-cloud albedo feedback is a relationship governing the cloud droplet number density. It is believed that most natural CCN in marine air consist of sulfate aerosols. Sub-micrometer sulfate aerosols are not emitted directly from land or ocean surfaces, but are created in the atmosphere by gas to particle conversion. DMS is the most important source of atmospheric sulfur over the oceans: over land, however, sulfur is emitted as part of a variety of compounds (Charlson *et al.*, 1987). The oxidation of DMS produces, among other products, sub-micrometer sulfate aerosols, which directly contribute to the backscatter of solar radiation, and thus induce a small cooling effect (Shaw, 1983). Charlson *et al.* (1987) suggested, however, that the more important effect of sulfate aerosols would be through their role as sources of CCN, rather than as non-nucleated aerosols.

In the marine boundary layer, condensation nuclei concentrations are generally between 50 and 300 cm^{-3}. Bates *et al.* (1987) have shown that there is an apparently linear relationship (with a correlation coefficient of 0.90) between nuclei concentrations and the calculated DMS flux, such that

$$N_{cn} = 76.9 F_{\mathrm{DMS}} - 16.9, \tag{7}$$

where N_{cn} is the condensation nuclei concentration (cm^{-3}) and F_{DMS} the DMS flux to the atmosphere (micro-mole $\mathrm{m}^{-2}\ \mathrm{d}^{-1}$).

For the sake of simplicity, we shall assume that when a cloud forms, each condensation nucleus becomes activated so that the cloud droplet concentration, N, is equivalent to the condensation nuclei concentration:

$$N = N_{cn}. \tag{8}$$

Pruppacher and Klett (1978) indicate that in fact no simple relationship between N and N_{cn} holds under all conditions. The simplifying assumption given above, however, will provide an upper bound on the strength of the feedback being studied here.

2.4. *Flux of DMS to the Atmosphere*

The final link required by the DMS-cloud albedo feedback mechanism is a relationship between climatic variables and the flux of DMS to the atmosphere. It has long been established that marine plankton are a source of dimethylsulfde, $(CH_3)_2S$ (Lovelock *et al.*, 1972). Although the biological function of DMS remains unclear, it is known to be produced by phytoplankton and by grazing zooplankton in the surface waters of the ocean (Dacey and Wakeham, 1986). Despite its biological origin, DMS concentration in surface sea water, and hence its atmospheric flux, is weakly correlated with plankton productivity. While appearing to be independent of plankton productivity, DMS concentrations are highest in the warmest, most saline, and most illuminated regions (Andreae, 1986).

Charlson *et al.* (1987) speculate that there are two reasons for the lack of a strong correlation between DMS concentration and plankton productivity. First, the concentration of DMS in the surface waters is not only dependent on rates of production, but on rates of removal as well. DMS is removed from the ocean by three processes: atmospheric diffusion, photochemical dissociation, and microbial digestion. A second reason for the lack of a simple relationship between plankton productivity and DMS concentration may be attributed to the large variation (over three orders of magnitude) of DMS production rates in different species (Keller, private communication).

Because oceanic DMS concentrations and rates of atmospheric diffusion are weakly correlated with productivity, efforts have been made to find empirical relationships between DMS flux and climatic variables. In particular measurements have been made attempting to relate DMS flux, sea surface temperature, and surface insolation, although other factors may also be important (e.g. salinity, wind stress, etc.).

Recent measurements exhibit only a weak correlation between DMS flux and sea surface temperature. According to Bates *et al.* (1987), a linear regression between calculated DMS flux and sea surface temperature gives a correlation coefficient, r, of only 0.44. Therefore, any existing albedo feedback mechanism does not have a simple, linear association with temperature.

It has been found that the relative change in seasonal surface insolation has a high correlation ($r = 0.93$) with relative seasonal variations in DMS flux (Bates *et al.*, 1987). Their work also indicates that there is a strong correlation ($r = 0.90$) between the daily input of solar radiative energy to the oceans and calculated DMS flux to the atmosphere, such that

$$F_{\mathrm{DMS}} = 0.0151\,Q + 0.50, \tag{9}$$

where Q is the daily average solar radiation ($\mathrm{W\ m^{-2}}$) incident on the ocean surface (estimated under the assumption that the atmospheric transmission coefficient is 0.8).

3. Model Sensitivity to Imposed Changes in Insolation

Our model is now complete and can be used to determine the strength of the DMS-cloud albedo feedback. The model will be applied first to a hypothetical, ocean-covered and completely overcast Earth, which will yield an upper bound on the strength of the DMS-cloud albedo feedback. The true strength of the feedback will then be roughly estimated by considering the fraction of Earth that is actually ocean covered and overcast.

The strength of the feedback can be characterized by the normalized difference between absorbed solar radiation with and without the feedback,

$$f_{\mathrm{DMS}} = 1 - \frac{\Delta Q_o}{\Delta Q}, \tag{10}$$

where ΔQ is the equilibrium increase in solar radiation absorbed by the surface when a small increase in the solar constant is imposed. The subscript 'o' identifies the 'no feedback' case, which is obtained simply by holding the optical depth of the clouds fixed in (1)–(3).

To evaluate the strength of the feedback, we expand the change in solar radiation using the chain rule of differentiation,

$$\begin{aligned} \Delta Q &= \Delta Q_o + \frac{\partial Q}{\partial \delta} \Delta \delta \\ &= \Delta Q_o + \frac{\partial Q}{\partial \delta} \frac{\partial \delta}{\partial N} \Delta N \\ &= \Delta Q_o + \frac{\partial Q}{\partial \delta} \frac{\partial \delta}{\partial N} \frac{\partial N}{\partial Q} \Delta Q, \end{aligned} \tag{11}$$

where we have assumed the cloud liquid water content does not change. Substitution of (11) into (10) yields the DMS-cloud albedo feedback strength,

$$f_{\mathrm{DMS}} = \frac{\partial Q}{\partial \delta} \frac{\partial \delta}{\partial N} \frac{\partial N}{\partial Q}. \tag{12}$$

From (1)–(3)

$$\frac{\partial Q}{\partial \delta} = \frac{\partial Q}{\partial \alpha_p} \frac{\partial \alpha_p}{\partial \alpha_c} \frac{\partial \alpha_c}{\partial \delta} = -\frac{\mu S_o}{2} \left(\frac{1 - \alpha_s}{1 - \alpha_s \alpha_c}\right)^2 \frac{\partial \alpha_c}{\partial \delta} \simeq -\frac{\mu S_o}{2} \frac{\alpha_c(1 - \alpha_c)}{\delta}, \tag{13}$$

where we have assumed $2\alpha_s(1 - \alpha_c)/(1 - \alpha_s) \ll 1$. From (6)

$$\frac{\partial \delta}{\partial N} = \frac{\delta}{3N}. \tag{14}$$

From (7), (8), and (9),

$$N = aQ + b, \tag{15}$$

so that

$$\frac{\partial N}{\partial Q} = a, \tag{16}$$

where the constants a and b have the empirical values 1.16 m^2 W^{-1} cm^{-3} and 21.2 cm^{-3}, respectively. Combining (12), (13), (14), and (16), the DMS-cloud albedo feedback strength reduces to

$$f_{\text{DMS}} \simeq -\frac{a\mu S_o}{2}\,\frac{\alpha_c(1-\alpha_c)}{3N}. \tag{17}$$

In order to evaluate the magnitude of f_{DMS}, we prescribe the solar zenith angle ($\mu = 0.5$), the surface albedo ($\alpha_s = 0.05$), and the nominal value of the solar constant ($S_o = 1370$ W m^{-2}). Given these parameters and the empirical constants a and b, the feedback strength is entirely determined by the cloud optical depth. Because the optical depth of marine stratiform clouds ranges roughly from 3 to 30 (Stephens, 1978), it is instructive to explore the dependence of the strength of the feedback on δ. For a given optical depth, the surface solar radiation and cloud albedo are calculated from (1)–(3). The feedback parameter is then evaluated from (17) after using (15) to determine N.

Figure 3 shows how the strength of the DMS-cloud albedo feedback varies with cloud optical thickness. As is clear from (17), the shape of the curve is controlled mainly by the quantity, $\alpha_c(1 - \alpha_c)$, which by (3) is equivalent to $\delta\partial\alpha_c/\partial\delta$, i.e. the slope of the curve shown in Figure 2. According to (3), the slope of the curve is a maximum at an optical depth of 7.7 ($\alpha_c = 0.5$), and the slope approaches zero for both large and small optical thicknesses. The DMS-cloud albedo feedback would

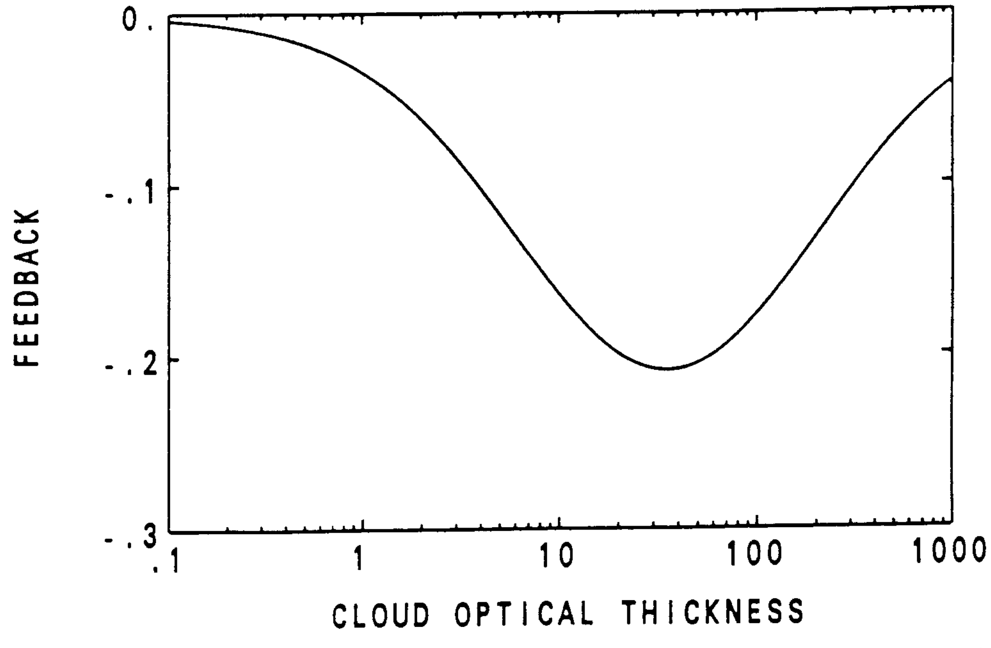

Fig. 3. DMS-cloud albedo feedback strength dependence on cloud optical thickness for a cloud covered ocean. Most marine stratiform clouds have optical thicknesses in the range of 3 to 30.

also be strongest at an optical depth of 7.7 if it were not for the dependence of f_{DMS} on the cloud droplet number density. According to (15), optically thicker clouds lead to a reduction in N which according to (17) strengthens the feedback. Thus the maximum strength of the feedback is shifted slightly toward higher cloud optical depths.

The maximum magnitude of the feedback shown in Figure 3 is about 0.2. This should be considered an upper bound for global average climatic response because the feedback should only be active over cloud covered ocean regions. Since approximately 1/3 of the Earth actually meets these conditions, the globally averaged magnitude of the feedback is likely less than 0.07. According to the definition of f_{DMS}, the net change in solar forcing is scaled by the factor $1/(1 - f_{\mathrm{DMS}})$ when the feedback is active. Thus the feedback reduces the imposed change in solar forcing by less than 7%.

4. DMS-Cloud Albedo Feedback Effects on Climatic Response

In the previous section, we analyzed how the DMS-cloud albedo feedback can *directly* reduce the effects of changes in the solar radiative forcing of the climate system. We now consider the role of this feedback in modifying the climatic response to *any* forcing, whether or not it *directly* affects surface insolation. As an illustrative example, we shall estimate the importance of the DMS-cloud albedo feedback on the climatic effects of doubling the atmospheric concentration of carbon dioxide. In this case the DMS-cloud albedo feedback is activated primarily by other cloud feedbacks that affect surface insolation.

Schlesinger and Mitchell (1987) have developed a formalism for characterizing the strength of various feedbacks. They show that the equilibrium temperature response of climate (ΔT) is related to changes in forcing (i.e., perturbations in the radiative fluxes at the tropopause, ΔR_{ext}) by a coefficient that depends on the gain of the climate system (G_o) and the feedbacks (f):

$$\Delta T = \frac{G_o}{1-f}\,\Delta R_{\mathrm{ext}}, \tag{18}$$

where

$$G_o = -\left(\frac{\partial R}{\partial T}\right)^{-1} \tag{19}$$

and

$$f = G_o \sum \frac{\partial R}{\partial I_j}\,\frac{\mathrm{d}I_j}{\mathrm{d}T}\,. \tag{20}$$

The gain is a measure of how the net radiation, R, would depend on temperature if all other aspects of the climate system remained unchanged. The feedback strength parameter, f, represents the sum of all climate feedbacks (i.e. the effects on temperature of changes in the climatic state other than temperature itself). It is presumed that all the internal variables of the climate system can be written ultimately as functions of temperature alone.

Radiative transfer models have shown that an instantaneous doubling of atmospheric CO_2 concentration reduces the outgoing radiation at the tropopause by about 4 W m^{-2} (Ramanathan *et al.*, 1979). This is the forcing, ΔR_{ext}. The gain of the climate system can be estimated from simple energy balance considerations, as has been done, for example, by Schlesinger and Mitchell (1987) who find G_o to be about 0.3 K (W m^{-2})$^{-1}$. The feedback parameter, f, is still uncertain. Most models (including general circulation models) find that a doubling of the atmospheric concentration of carbon dioxide leads to an equilibrium surface temperature change (globally averaged) of somewhere between 2 and 5 K. Thus, the feedbacks in these models evidently range from about $f = 0.4$ to $f = 0.75$.

The above formalism for describing climate feedbacks cannot be applied to DMS-cloud albedo feedback because the cloud drop number density, N, as given by (15), cannot be expressed as a function of temperature. To account for the DMS-cloud albedo feedback, (18) can be modified to include all effects of changes in surface insolation, whether externally forced or induced as part of the climate response. The result is

$$\Delta T = G_o \frac{\gamma \Delta Q_{ext} + \Delta F_{ext}}{1 - \gamma f_{sw} - f_{lw}}, \tag{21}$$

where

$$\gamma = (1 - f_{DMS})^{-1}, \tag{22}$$

and f_{sw} and f_{lw} are feedbacks affecting shortwave and longwave radiative fluxes, respectively, and ΔQ_{ext} and ΔF_{ext} are imposed external changes in the shortwave and longwave forcing, respectively. For $\gamma = 1$ (i.e., no DMS-cloud albedo feedback), (21) reduces to (18).

In the previous section we found that f_{DMS} is no larger than about 0.1 so that γ is between about 0.9 and 1. According to (21), the DMS-cloud albedo feedback can affect climatic response in two ways. It can directly modify changes in shortwave forcing, as indicated by the coefficient multiplying the shortwave forcing term in the numerator of (21), which is consistent with the results of the previous section. The DMS-cloud albedo feedback can also affect the shortwave feedbacks, as indicated by the term, γf_{sw}, in the denominator of (21).

As an example of how the DMS-cloud albedo feedback moderates other shortwave feedbacks, we consider the CO_2-doubling experiment. The role of cloud feedbacks in the climate system is highly uncertain, but recent studies emphasize their importance in determining climate sensitivity (Cess *et al.*, 1989). We shall focus our attention on a single component of cloud feedback: cloud amount. Several general circulation models suggest that as the climate warms, cloud fraction decreases (Wetherald and Manabe, 1986; Wilson and Mitchell, 1987). The resulting decrease in planetary albedo provides a positive feedback on climate. The change in net incoming shortwave radiation resulting from a change in cloud fraction is given by

$$\Delta Q = -\frac{S_0}{4}(\alpha_{\text{overcast}} - \alpha_{\text{clear}})\Delta C, \tag{23}$$

where ΔC is the change in globally averaged cloud fraction and α_{overcast} and α_{clear} are albedos typical of overcast and clear regions, respectively.

Taking typical values for overcast albedo ($\alpha_{\text{overcast}} = 0.55$), clear-sky albedo ($\alpha_{\text{clear}} = 0.15$), and the change in cloud amount ($\Delta C = 0.035$), we find the change in net shortwave radiative flux at the top of the atmosphere due to cloud area feedback is 4.8 W m^{-2}. Recalling that the gain of the climate system is about 0.3 K (W $\text{m}^{-2})^{-1}$, and using a value for f that leads to a 4 K increase in temperature for a doubling of CO_2 (i.e. $f = 0.7$), we calculate that the contribution of the shortwave cloud area feedback to the total feedback is about 0.36.

We can now estimate the effect of including the DMS-cloud albedo feedback on the cloud area feedback. According to (21), the shortwave cloud area feedback strength will be reduced by a factor of $f_{\text{DMS}}/(1 - f_{\text{DMS}})$ when the DMS-cloud albedo feedback is included. For $f_{\text{DMS}} = 0.1$, the overall effect is to reduce the total feedback from 0.7 to 0.67. The climate sensitivity is thereby decreased from 1.0 to 0.9 K (W $\text{m}^{-2})^{-1}$. In short, if the cloud area feedback were the only feedback affecting shortwave surface insolation, then including DMS-cloud albedo feedback would reduce the temperature change resulting from a doubling of the concentration of carbon dioxide from 4 to 3.6 K.

This analysis has clearly not been comprehensive because all feedbacks affecting shortwave surface insolation have not been considered. Furthermore, we have most likely exaggerated the effect of the DMS-cloud albedo feedback by using a value of f_{DMS} that is likely to be its highest possible value. Even so, we find that the DMS-cloud albedo feedback modifies other feedbacks by 10% or less.

5. Conclusions

In this exploratory study we have for the first time estimated the strength of the DMS-cloud albedo feedback in a model that accounts for the empirical relationships found by Bates *et al.* (1987) between insolation, DMS flux, and condensation nuclei concentration. We have found that this feedback, proposed by Charlson *et al.* (1987) as a mechanism by which the biosphere could partially control climate, is much weaker than other feedbacks thought to be important in the climate system (Schlesinger, 1988; Hansen *et al.*, 1984).

Because so little is currently known about what controls the DMS production and destruction in the biosphere, we have relied on the correlation found by Bates *et al.* (1987), which relates DMS flux to surface insolation. We have also used their empirical relationship between DMS flux and condensation nuclei concentration. Until more is known about the processes that govern the DMS flux and cloud and aerosol microphysics, the estimates we have made here will remain uncertain. It is conceivable, for example, that the empirical relationships found by Bates *et al.* (1987) are valid only on seasonal time scales, not the decadal and longer time scales of interest in climate studies. This suggests that in order to increase confidence in the results reported here, further evidence (both theoretical and empirical) in support of the Bates *et al.* (1987) relationships should be sought. Only then would it be appropriate to repeat this study using a climate model of higher resolution (e.g., three-dimensional general circulation models).

It should be emphasized that even though this study indicates that the DMS-cloud albedo feedback is apparently of relatively minor importance, there can still be fundamentally important links between the marine biosphere and climate. The fact that the DMS flux to the atmosphere appears to be only weakly dependent on insolation does not diminish its apparent role as a source of CCN. To the extent that this source of CCN affects cloud albedo, the climate is dependent on the marine biosphere.

There is circumstantial evidence that supports the contention that changes in DMS flux can affect global temperature. Legrand *et al.* (1988) have analyzed Antarctic ice cores which show that non-seasalt sulfates (oxidation products of DMS) were 20–46% higher during the last ice age. In a similar study, Saigne and Legrand (1987) show that another oxidation product of DMS (methanesulfonic acid) was about a factor of 2 to 3 higher than today. It is certainly possible that increases in cloud albedo caused by relatively large increases in the flux of DMS to the atmosphere could have contributed to the cooler temperatures of the ice age. In our model a 10% increase in CCN over the oceans would lead to a decrease in the globally averaged net solar radiation of about 1 W m^{-2}. A typical estimate of climate sensitivity would translate this forcing into a global average equilibrium surface temperature change of about 1 K.

We conclude that although the empirical relationships of Bates *et al.* (1987) do not suggest the link between insolation and DMS flux is strong enough to create an important feedback mechanism, further investigations should be undertaken to determine the processes that might cause the observed variations in DMS products over time scales of several thousand years. Only then will it be possible to evaluate the role of phytoplankton in initiating, amplifying, or moderating global climate change.

Acknowledgements

The authors would like to thank S. Twomey for suggesting the cloud albedo parameterization used here. R. Cess, S. Warren, and G. Shaw provided thoughtful reviews and, in particular, R. Cess suggested that we consider the effects of the DMS-cloud albedo feedback on other feedbacks. We also would like to extend thanks to J. E. Kutzbach, A. DuVall, M. MacCracken, and C. Covey for helpful comments and suggestions. This work was performed under the auspices of the

U.S. Department of Energy Atmospheric and Climate Research Division by the Lawrence Livermore National Laboratory under Contract No. W-7405-Eng-48 while one of the authors (JAF) was visiting the Laboratory during the summer of 1988.

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(Received 13 March, 1989; in revised form 16 January, 1990)