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Global temperature stabilization via controlled albedo enhancement of low-level maritime clouds

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An assessment is made herein of the proposal that controlled global cooling sufficient to balance global warming resulting from increasing atmospheric CO_2 concentrations might be achieved by seeding low-level, extensive maritime clouds with seawater particles that act as cloud condensation nuclei, thereby activating new droplets and increasing cloud albedo (and possibly longevity). This paper focuses on scientific and meteorological aspects of the scheme. Associated technological issues are addressed in a companion paper.

Analytical calculations, cloud modelling and (particularly) GCM computations suggest that, if outstanding questions are satisfactorily resolved, the controllable, globally averaged negative forcing resulting from deployment of this scheme might be sufficient to balance the positive forcing associated with a doubling of CO_2 concentration. This statement is supported quantitatively by recent observational evidence from three disparate sources. We conclude that this technique could thus be adequate to hold the Earth's temperature constant for many decades.

More work—especially assessments of possible meteorological and climatological ramifications—is required on several components of the scheme, which possesses the advantages that (i) it is ecologically benign—the only raw materials being wind and seawater, (ii) the degree of cooling could be controlled, and (iii) if unforeseen adverse effects occur, the system could be immediately switched off, with the forcing returning to normal within a few days (although the response would take a much longer time).

Keywords: global cooling; seeding maritime clouds; cloud albedo; negative forcing

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1. Introduction

Atmospheric clouds exercise a significant influence on climate. They can inhibit the passage through the atmosphere of both incoming, short-wave solar radiation, some of which is reflected back into space from cloud tops, and they intercept long-wave radiation flowing outwards from the Earth's surface. The first of these effects produces a global cooling, the second a warming. On balance, clouds produce a cooling effect, corresponding to a globally averaged negative net forcing of approximately $-13\,\mathrm{W\,m^{-2}}$ (Ramanathan *et al.* 1989).

Since the estimated positive forcing resulting from a doubling of the atmospheric carbon dioxide (CO_2) concentration (from the value—approximately 275 ppm—existing at the beginning of the industrial period) is approximately $+3.7\,\mathrm{W\,m^{-2}}$ (Ramaswamy *et al.* 2001), it is clear that, in principle, deliberate modification of clouds to produce a cooling sufficient to balance global warming resulting from the burning of fossil fuels is feasible.

This paper presents and assesses a proposed scheme for stabilization of the Earth's global mean temperature (in the face of continually increasing atmospheric CO_2 concentrations) by seeding clouds in the marine boundary layer (MBL) with seawater aerosol, in order to increase the cloud droplet number concentration and thus the cloud albedo (and possibly longevity): thereby producing a cooling. This paper focuses attention on the physics and meteorology of the idea. Technological aspects are treated in a companion paper (Salter *et al.* 2008).

Section 2 outlines the global cooling scheme and some cloud model sensitivity studies designed to determine the sensitivity of cloud albedo enhancement to values of the meteorological and cloud microphysical parameters involved. Section 3 presents some simple calculations designed to illustrate the potential viability of the technique. Section 4 presents (GCM) climate computations that provide a more rigorous quantitative assessment. Technological implications from the results of these computations are discussed in §5. A brief discussion of questions and concerns that would need to be satisfactorily examined before any justification would exist for the operational deployment of the technique is presented in §6. Section 7 provides a provisional quantitative assessment of the extent to which global temperature stabilization might be possible with this technique.

2. Principle and first assessment of the idea

Low-level, non-overlapped marine stratiform clouds cover about a quarter of the oceanic surface (Charlson *et al.* 1987) and characteristically possess albedos, A, in the range 0.3–0.7 (Schwartz & Slingo 1996). They, therefore, make a significant (cooling) contribution to the radiative balance of the Earth. Latham (1990, 2002) proposed a possible technique for ameliorating global warming by controlled enhancement of the natural droplet number concentrations (N_0) in such clouds, with a corresponding increase ΔA in their albedo (the first indirect or Twomey effect (1977)), and also possibly in their longevity (the second indirect, or Albrecht effect (1989)), thus producing cooling. N_0 values in these clouds range typically from approximately 20 to $200\,\mathrm{cm^{-3}}$.

The technique involves dissemination—at or close to the ocean surface—of monodisperse seawater (NaCl) droplets approximately 1 μm in size, which possess sufficiently large salt masses always to be activated—as cloud condensation nuclei (CCN)—to form ΔN additional droplets when (shrinking by evaporation in the subsaturated air en route) they rise into the cloud bases. The total droplet concentration N after seeding thus lies between ΔN and $(N_0+\Delta N)$, because some of the natural CCN which would be activated in the absence of seeding may not be in its presence, owing to the lower supersaturations that prevail. The central physics behind this scheme, which have been authoritatively treated in a considerable number of studies (e.g. Twomey 1977, 1991; Charlson *et al.* 1987; Albrecht 1989; Wigley 1989; Slingo 1990; Ackerman *et al.* 1993; Pincus & Baker 1994; Rosenfeld 2000; Stevens *et al.* 2005), is that an increase in droplet concentration N causes the cloud albedo to increase because the overall droplet surface area is enhanced. It can also increase cloud longevity (tantamount to increasing cloudiness) because the growth of cloud droplets by coalescence to form drizzle or raindrops, which often initiates cloud dissipation, is impeded, since the droplets are smaller and the clouds correspondingly more stable. Possibly significant departures from this simple picture are outlined in §6.

Calculations by the above-mentioned workers indicate that a doubling of the natural droplet concentration (i.e. to $N=2N_0$) in all such marine stratiform clouds (which corresponds to an increase ΔA of approximately 0.06 in their cloud-top albedo) would produce cooling sufficient roughly to balance the warming associated with CO_2 doubling. Latham (1990, 2002) calculated that for a droplet diameter $d=0.8$ μm (associated salt mass $m_s=10^{-17}$ kg) the total (global) seawater volumetric dissemination rate dV/dt required to produce the required doubling of N in all suitable marine stratocumulus clouds is approximately $30\,m^3\,s^{-1}$, which appears well within the range of modern technology. It is considered (e.g. Charlson *et al.* 1987) that most natural CCN over the oceans consist of ammonium sulphate particles formed from dimethyl sulphide produced at the ocean surface by planktonic algae. In order to ensure at least a doubling of N, we would need to add at least $2N_0$ particles per unit volume. Ideally, their size distribution would be monodisperse, largely in order to avoid the production of ultra-giant nuclei, UGN (Woodcock 1953; Johnson 1982; De Leeuw 1986), which could act to promote drizzle formation and thus cloud dissipation. The monodispersity of the added particles may also make the clouds more colloidally stable, thus inhibiting coalescence and associated drizzle formation.

We point out that ship tracks are a consequence of inadvertent and uncontrolled albedo increase in such clouds, resulting from the addition of effective CCN in the exhausts from the ships and that our proposed deliberate generation of efficient sea-salt CCN at the ocean surface, thereby (usually) enhancing N, is of course basically a version of a process that happens naturally, via the catastrophic bursting of air bubbles produced by wave motion. However, except in conditions of high winds or in regions where other aerosol sources are weak, these sea-salt particles constitute only a small fraction of the CCN activated in marine stratocumulus (Latham & Smith 1990).

A simplified version of the model of marine stratocumulus clouds developed by Bower *et al.* (1999) was used (Bower *et al.* 2006) to examine the sensitivity of albedo enhancement ΔA to the environmental aerosol characteristics, as well as

those of the seawater aerosol of salt mass m_s and number concentration ΔN deliberately introduced into the clouds. Values of albedo change ΔA and total droplet number concentration N were calculated for a wide range of values of m_s, ΔN and various other cloud parameters. Computations were made for aerosol characteristics pertaining to clean, intermediate and polluted air masses (Spectra A, B and C, respectively). Values of ΔA were calculated from the droplet number concentrations using the method of Schwartz & Slingo (1996). It was found that, for Spectrum B, values of ΔA and N are insensitive to m_s over the range 10^{-17}–10^{-14} kg. For Spectrum A, the insensitivity range is 10^{-18}–10^{-15} kg, and the ΔA values are typically several times greater (for the same values of ΔN) than those for Spectrum B. For Spectrum C, the ΔA values are much lower than those for Spectra A and B. The above-mentioned threshold value of ΔA (0.06) was achieved for most parameter-value permutations for Spectrum A, a significant fraction for Spectrum B and scarcely any for Spectrum C. For all the three aerosol spectra, the calculated values of ΔA and total droplet concentration N were found to be highly sensitive to the imposed additional aerosol concentrations ΔN. The relationship between ΔN and ΔA was found always to be strongly nonlinear (e.g. Twomey 1991; Pincus & Baker 1994). These model computations provide provisional quantitative support for the physical viability of the mitigation scheme, as well as offering new insights (§5) into its technological requirements.

3. Relationships between spraying rate, albedo change and negative forcing

The simple calculations presented in this section are designed to illustrate the relationships between the deliberately imposed increase in cloud droplet number concentration, ΔN, the associated increase in cloud albedo, ΔA, the resultant globally averaged negative forcing ΔF, and the required continuous seawater aerosol volumetric spray production rate $\mathrm{d}V/\mathrm{d}t$. These calculations also provide some indication as to whether or not our global temperature stabilization scheme is quantitatively feasible. For the purposes of this discussion, we assume that the only clouds deliberately seeded with seawater CCN are non-overlapped marine stratiform clouds. As discussed in item 4 of §6, there are still many unknowns in our characterization of aerosol cloud interactions and aerosol *indirect effects*. However, theoretical calculations and global modelling both suggest that the first indirect effect generally dominates over the second, so the latter is disregarded in what follows. For similar reasons, we consider only short-wave radiative effects in this analysis.

The average solar irradiance F (W m^{-2}) received at the Earth's surface is

$$F = 0.25F_0(1 - A_P), \tag{3.1}$$

where F_0 ($=1370$ W m^{-2}) is the solar flux at the top of the atmosphere and A_P is the planetary albedo. Thus, an increase ΔA_P in planetary albedo produces a forcing ΔF of

$$\Delta F = -340\Delta A_P. \tag{3.2}$$

Table 1. Values of negative forcing ΔF (W m^{-2}) derived from equation (3.6) for selected values of N/N_0 (the ratio of the seeded to unseeded cloud droplet number concentration) and f_3, the fraction of suitable clouds seeded.

	f_3				
N/N_0	1.0	0.7	0.5	0.3	0.17
2	3.1	2.2	1.6	0.94	0.53
3	4.9	3.5	2.5	1.5	0.84
4	6.2	4.4	3.1	1.9	1.1
5	7.2	5.1	3.6	2.2	1.2
7	8.8	6.1	4.4	2.6	1.5
10	10.4	7.3	5.2	3.1	1.8

We define f_1 (=0.7) as the fraction of the Earth's surface covered by ocean, f_2 (=0.25) as the fraction of the oceanic surface covered by non-overlapped marine stratiform clouds, and f_3 as the fraction of oceanic stratiform cloud cover which is seeded. Thus, the average change ΔA in cloud albedo associated with a change ΔA_{P} in planetary albedo is

$$\Delta A = \frac{\Delta A_{\mathrm{P}}}{(f_1 f_2 f_3)} = -\frac{\Delta F}{60 f_3}, \tag{3.3}$$

from which it follows that, if $f_3=1$, to produce a globally averaged negative forcing of -3.7 W m^{-2}, the required increases in planetary and cloud albedo are 0.011 and 0.062, respectively: the associated percentage changes in albedo being roughly 3.7 and 12 per cent.

The cloud albedo increase resulting from seeding the clouds with seawater CCN to increase the droplet number concentration from its unseeded value N_0 to N is given (Schwartz & Slingo 1996) by

$$\Delta A = 0.075 \ln(N/N_0) \tag{3.4}$$

and it follows from equations (3.3) and (3.4) that

$$-\Delta F = 4.5 f_3 \ln(N/N_0), \tag{3.5}$$

which may be rewritten as

$$(N/N_0) = \exp(-\Delta F/4.5 f_3). \tag{3.6}$$

It follows from equation (3.6) that if $f_3=1$ (all suitable clouds seeded), the value of (N/N_0) required to produce a negative forcing of -3.7 W m^{-2} is 2.3, in reasonable agreement with the estimates of Charlson *et al.* (1987) and Slingo (1990).

Table 1 presents the values of globally averaged negative forcing, ΔF (W m^{-2}), derived from equation (3.5) for a range of values of f_3 and N/N_0. We see that if the fraction f_3 of suitable clouds that are seeded falls below approximately 0.3, it is not possible, for values of (N/N_0) realistically achievable on a large scale (a rough estimate is $(N/N_0)<10$), for our scheme to produce a negative forcing of -3.7 W m^{-2}. We also see the distinct nonlinearity in the relationship between (N/N_0) and ΔF.

The volumetric spraying rate

$$\frac{\mathrm{d}V}{\mathrm{d}t} = v_{\mathrm{d}}\frac{\mathrm{d}n}{\mathrm{d}t} = (\pi/6)d^3\frac{\mathrm{d}n}{\mathrm{d}t}, \tag{3.7}$$

where v_{d} is the volume (m^3) of a seawater droplet of diameter d (m) at creation and $\mathrm{d}n/\mathrm{d}t$ (s^{-1}) is the rate of spraying of seawater droplets.

We assume that in equilibrium the number of sprayed droplets residing in the atmosphere is constant, i.e. the deliberate creation rate of seawater droplets equals the loss rate. Thus,

$$\frac{\mathrm{d}n}{\mathrm{d}t} = (N - N_0)\frac{A_E H f_1 f_2 f_3}{f_4 \tau_R} = N_0(N/N_0 - 1)\frac{A_E H f_1 f_2 f_3}{f_4 \tau_R}, \tag{3.8}$$

where A_{E} (m^2) is the surface area of the Earth; H (m) is the height over which the seawater droplets are distributed; f_4 is the fraction of the sprayed droplets that are not lost at creation and do not move laterally away from regions of selected cloud cover; and τ_{R} (s) is the average residence time of the seawater aerosol in the atmosphere.

Thus (from equations 3.6–3.8), taking $A_{\mathrm{E}}=5.1\times10^{14}\,\mathrm{m}^2$, $f_1=0.7$, $f_2=0.25$, we obtain

$$\frac{\mathrm{d}V}{\mathrm{d}t} = 4.6\times10^{13} f_3 d^3 (HN_0/f_4\tau_{\mathrm{R}})[\{\exp(-\Delta F/4.5 f_3)\} - 1]. \tag{3.9}$$

Assuming that $f_3=1$: $f_4=0.5$: $d=0.8\,\mu\mathrm{m}=8\times10^{-7}$ m: $H=1000$ m: $N_0=100\,\mathrm{cm}^{-3}=10^8\,\mathrm{m}^{-3}$: $\tau_{\mathrm{R}}=3$ days$=2.6\times10^5$ s (M. C. Barth 2008, personal communication; D. H. Lenschow 2008, personal communication; M. H. Smith 2008, personal communication), it follows from equation (3.9) that, for a negative forcing $\Delta F=-3.7\,\mathrm{W\,m^{-2}}$, the required total volumetric seawater aerosol dissemination rate $\mathrm{d}V/\mathrm{d}t=23\,\mathrm{m^3\,s^{-1}}$. Keeping all the above parameter values the same but seeding only half of the suitable clouds yields a value of $\mathrm{d}V/\mathrm{d}t$ of approximately $37\,\mathrm{m^3\,s^{-1}}$.

4. Global climate modelling computations

Global aspects of the cloud albedo enhancement scheme were examined using two separate models.

The first of these was the HadGAM numerical model, which is the atmospheric component of the UK Hadley Centre Global Model, based on the Meteorological Office Unified Model (UM), v. 6.1. It is described in Johns *et al.* (2004) and contains the *New Dynamics Core* (Davies *et al.* 2005). It is run at N96L38 resolution, i.e. 1.25° latitude by 1.875° longitude with 38 vertical levels extending to over 39 km in height. N96 denotes a resolution of 96 two-grid-length waves, i.e. 192 grid points in longitude. It has a non-hydrostatic, fully compressible, deep atmosphere formulation and uses a terrain-following, height-based vertical coordinate. It also includes semi-Lagrangian advection of all prognostic variables except density, and employs the two-stream radiation scheme of Edwards & Slingo (1996). The aerosol species represented include sulphate, black carbon,

biomass smoke and sea salt. The convection scheme is based on the mass flux scheme of Gregory & Rowntree (1990) (but with major modifications), and the large-scale cloud scheme is that of Smith (1990).

The HadGAM model was used to calculate 3-year mean values of cloud-top droplet effective radius r_{eff} (μm), liquid water path, LWP (g m^{-2}) and outgoing short-wave radiation flux F_{sw} (W m^{-2}) at the top of the atmosphere (TOA). In the control run (no seeding) the globally averaged cloud droplet number concentration, N, was approximately 100 cm^{-3}, and the model was then run again with N increased—in all regions of low-level maritime cloud (below approximately 3000 m, 700 hPa)—to 375 cm^{-3}. Such a value of N should be readily achievable technologically, if our global temperature stabilization scheme were ever to be operationally deployed.

The computed 5-year mean distributions of layer cloud effective radius r_{eff} (μm) and LWP (g m^{-2}) for unseeded and seeded marine low-level clouds are displayed in figures 1 and 2, respectively. They show that increasing the cloud droplet number concentration N from natural values to the seeded figure of N=375 cm^{-3} leads to a general decrease in droplet size (figure 1, the first indirect effect) and increase in LWP, with consequent decrease in the efficiency of precipitation development (figure 2, the second indirect effect). The changes in effective radius are clearly evident in the regions of persistent marine stratocumulus off the west coasts of Africa and North and South America, and also over much more extensive regions of the southern oceans. Changes in LWP in these same regions can be perceived but are less pronounced.

Figure 3 reveals that the imposed increase in N has caused an overall significant negative change ΔF in radiative forcing, which would cause a cooling of the Earth's climate. The largest effects are apparent in the three regions of persistent marine stratocumulus off the west coasts of Africa and North and South America, mentioned earlier, which together cover approximately 3 per cent of the global surface. Lower but appreciable values of negative forcing can be seen throughout the much more extensive regions of the southern oceans. The 5-year mean globally averaged TOA negative forcing resulting from the marine low-level cloud seeding is calculated to be -8.0 ± 0.1 W m^{-2}, more than twice that required to compensate for the 3.7 W m^{-2} warming associated with a doubling of atmospheric CO_2 concentration.

A similar calculation was performed in a developmental version of the NCAR community atmosphere model (CAM). The simulations were performed at $1.9° \times 2.5°$ latitude/longitude resolution (26 layers with a top near 40 km) using a newly developed microphysics parametrization (Gettelman *et al.* 2008; Morrison & Gettelman 2008). That parametrization uses a two-moment scheme predicting cloud mass and particle number for four classes of condensed water (small particle liquid and ice, and precipitation-sized rain and snow). Three 5-year simulations were conducted. The first simulation (the control) calculated cloud drop number using the drop activation parametrization of Abdul-Razzak & Ghan (2005) with a functional dependence on aerosol type and concentration, and resolved and turbulent dynamical fields. The other two simulations overrode the cloud droplet number concentrations (N_{C}) below 850 hPa, prescribing them as 375 and 1000 cm^{-3}, wherever clouds were found. The influence of the warm cloud seeding geoengineering strategy is assessed by taking the difference between the seeding experiments and the control simulation.

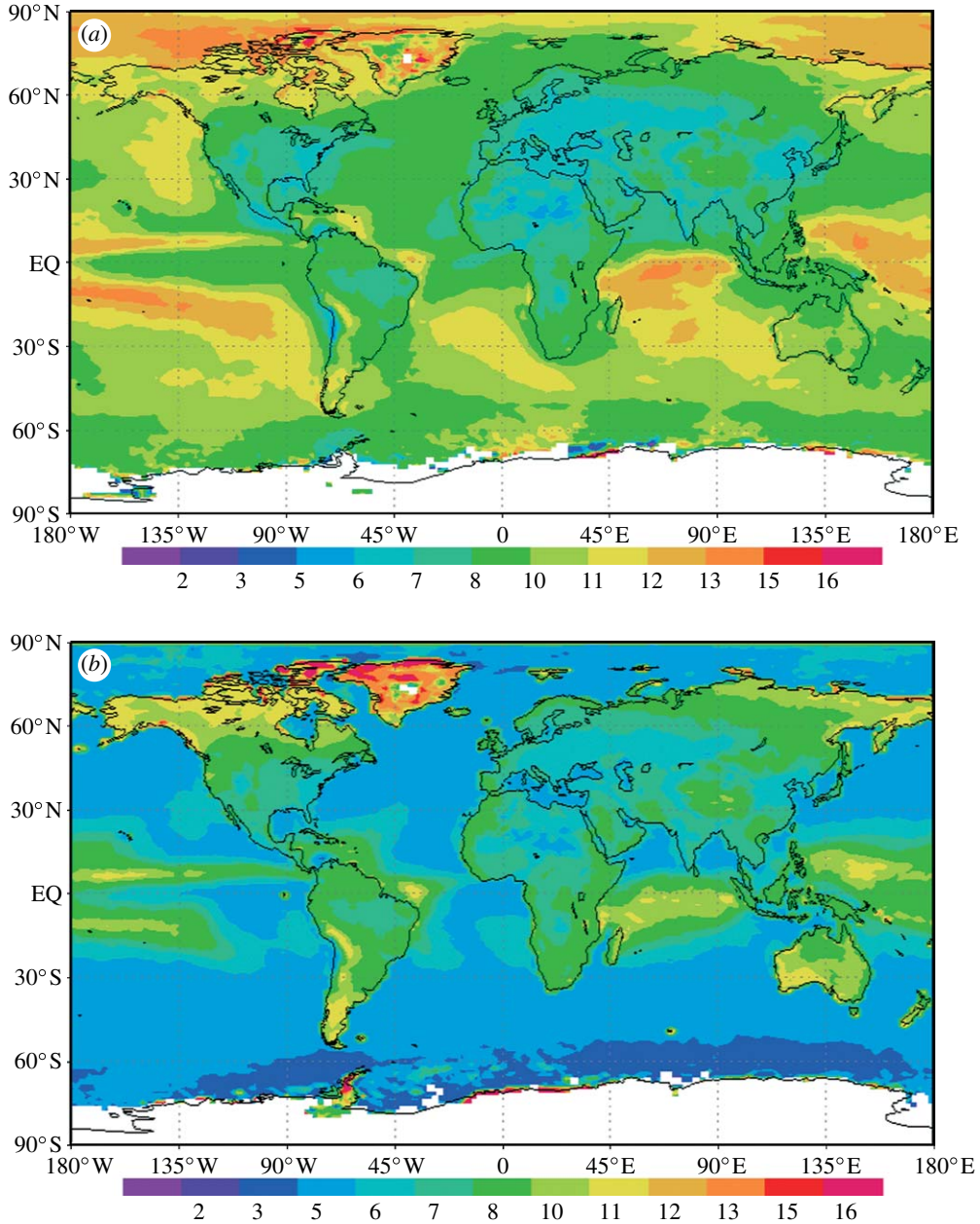

Figure 1. Three-year mean distributions of cloud-top effective radius r_{eff} (μm) in all regions of marine stratocumulus. (*a*) Control and (*b*) with N=375 cm^{-3} in regions of low-level maritime cloud.

Figure 4 shows the top of atmosphere SWCF (figure 4*a*) for the NCAR model for the three simulations and the difference between SWCF for the control and experiments where the drop number was prescribed to be 375 and 1000 cm^{-3} below 850 hPa (figure 4*b*,*c*, respectively, with 4*b* showing a similar quantity to that described in figure 3 for the HadGAM model). The NCAR model shows the effect of the cloud seeding in some of the same regions. The change in SWCF is

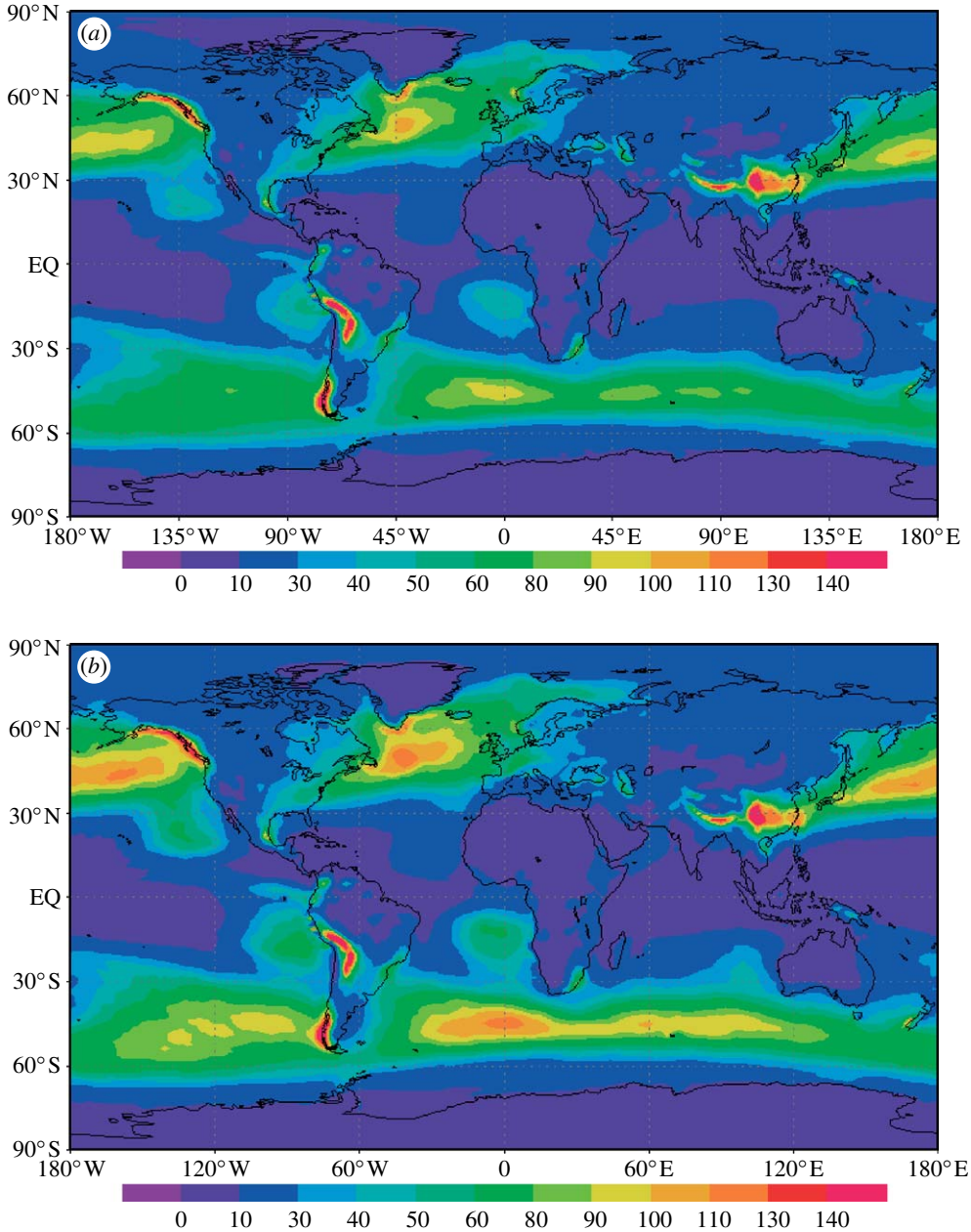

Figure 2. Three-year mean distributions of LWP (g m^{-2}). (*a*) Control and (*b*) with N=375 cm^{-3} in regions of low-level maritime cloud.

approximately half the amplitude of that seen in the HadGAM model in the marine stratus and trade cumulus regions. The HadGAM simulations prescribed the drop number to approximately 700 hPa, somewhat higher than the NCAR simulations, but the difference may also result from the many uncertainties in modelling cloud aerosol interactions in global climate models. Unlike the HadGAM simulations, there is also an intriguing response in the mid-latitude

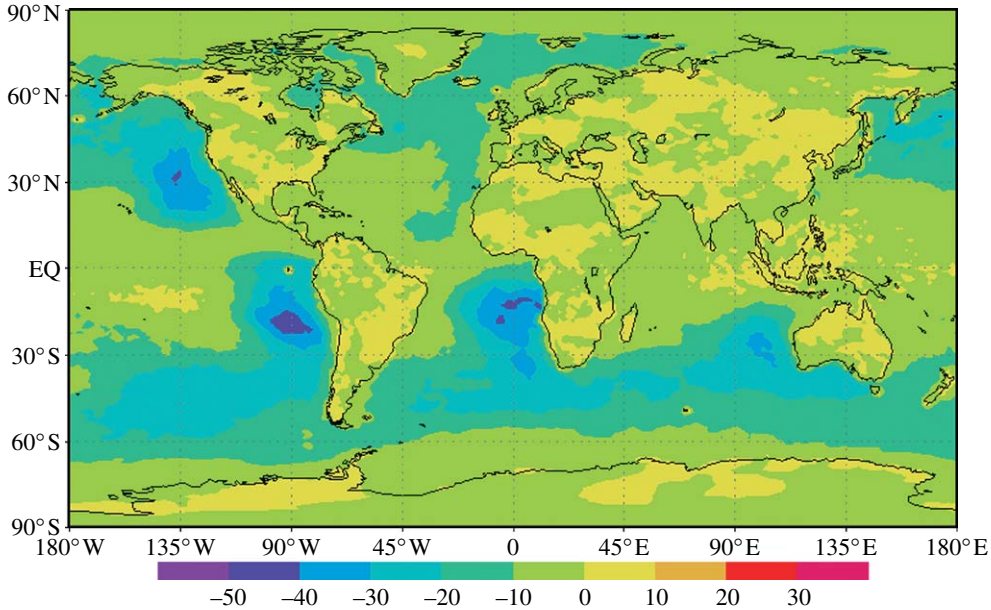

Figure 3. Three-year mean difference ΔF (W m^{-2}) in radiative forcing between the control simulation and that in which N=375 cm^{-3} in regions of low-level maritime cloud.

storm tracks, and there are some patches of positive ΔSWCF evident in the simulations as well. Some of the areas of positive ΔSWCF in the simulations with more moderate seeding (to 375 cm^{-3}; e.g. a weakening of the cloud forcing) occur downstream of regions strongly influenced by anthropogenic aerosols (e.g. downstream of China, and the eastern USA). In our model, producing clouds with 375 drops per cm^3 in these regions would actually constitute a reduction in N_C. Those regions are not seen in the simulation where the drop number is increased to 1000 cm^{-3}. Other regions where the SWCF increases (slightly, less than 6 W m^{-2}) in the central Pacific (Northern and Southern Hemispheres) are not common to the two simulations, and we believe that these regions are an artefact of the relative brevity of the simulations, and are indicative of interannual variability.

Because the ΔSWCF field exhibits significant spatial variation, it is clear that some geographical locations are more susceptible to cloud seeding than others. In other words, one may significantly reduce the cost of the warm cloud seeding geoengineering strategy by selecting the locations where cloud seeding should be applied to achieve the maximum amount of cooling. Therefore, it is important to identify these optimal locations for cloud seeding.

To achieve this goal, we first analysed the impact of warm cloud seeding by ranking the intensity of response (ΔSWCF) in all grid cells over the ocean surface. We considered the amplitude of the forcing change, and the area occupied by each grid cell (varying with the cosine of latitude) in performing the ranking. The accumulated forcing (ΔSWCF) based on the ranked orders of all grid cells over the ocean surface is presented in figure 5. Ranking was performed on the monthly mean forcings for each month of the simulation and then the results composited to produce figure 5.

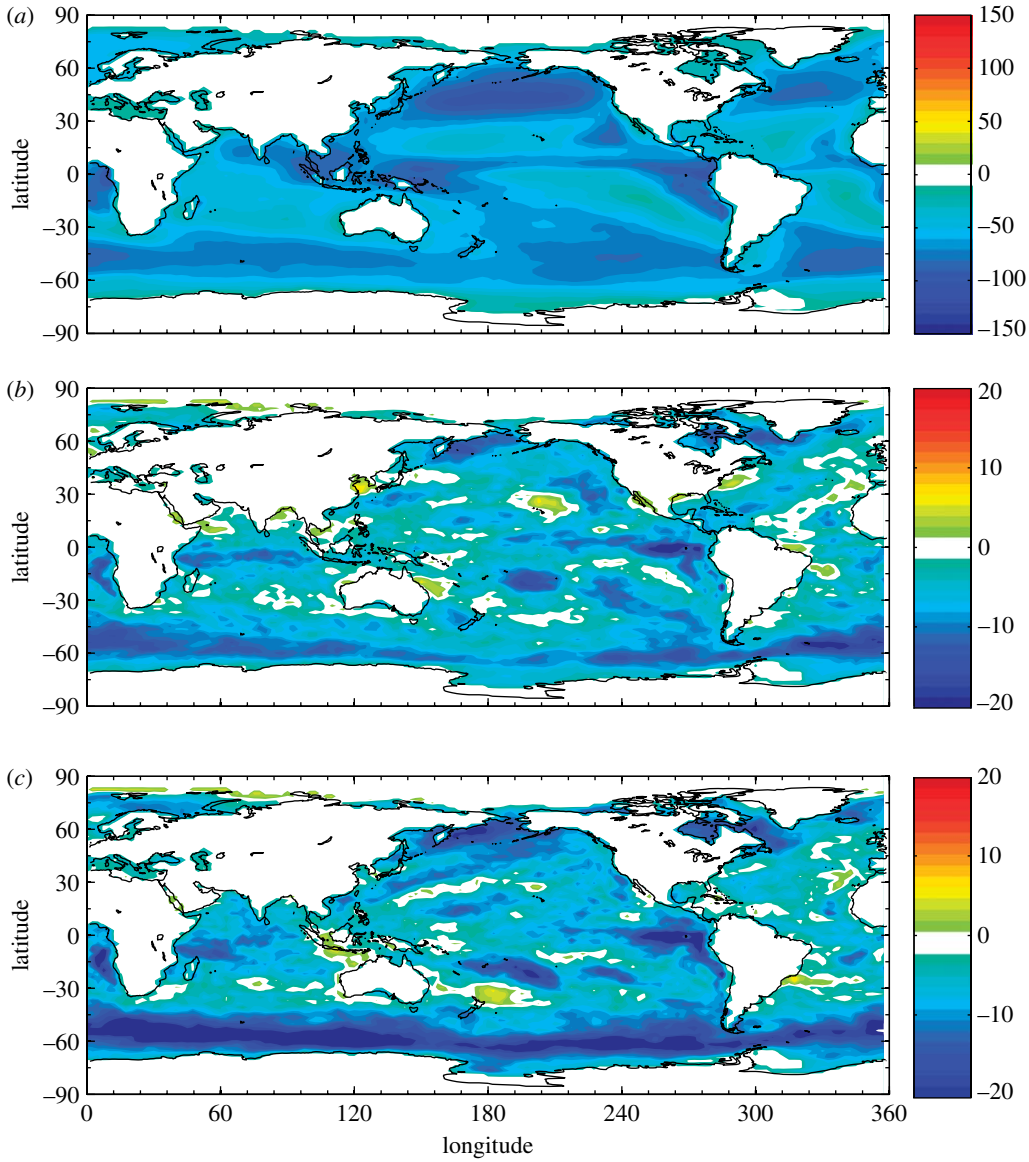

Figure 4. Annual average of (*a*) shortwave cloud forcing (SWCF) of the control simulation, (*b*) shortwave cloud forcing difference (ΔSWCF) between a geoengineering experiment by setting the cloud drop concentration to 375 cm^{-3} and the control simulation, and (*c*) ΔSWCF between a geoengineering experiment by setting the cloud drop concentration to 1000 cm^{-3} and the control simulation. For the purpose of this study, we only plot results over the ocean surface.

It is evident that the December–January–February (DJF) seasonal mean has the strongest response and the June–July–August (JJA) seasonal mean has the weakest response. The annual average (ANN) falls between the two seasonal means. Since the Sun is most intense in the Southern Hemisphere during DJF, we expect most of the important locations for seeding to reside in that hemisphere during that season, with the converse true during JJA. The stronger

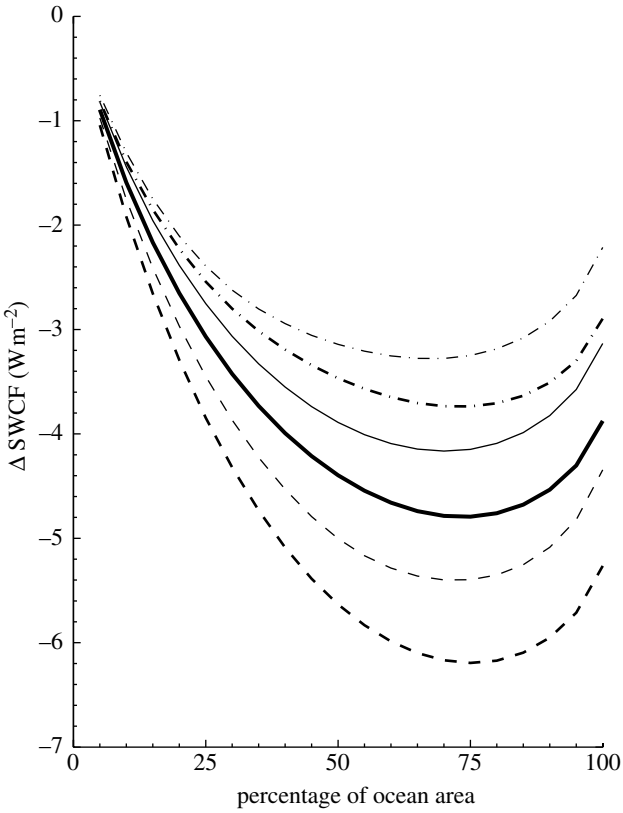

Figure 5. Cumulative ΔSWCF based on the ranked orders of all grid cells over the ocean for the two geoengineering experiments (N_C=375 cm^{-3} and 1000 cm^{-3}). The annual average (ANN) and the seasonal means (DJF: December–January–February; JJA: June–July–August) are plotted for both cases. Dashed curve, DJF 375; dot-dashed curve, JJA 375; solid curve, ANN 375; bold dashed curve, DJF 1000; bold dot-dashed curve, JJA 1000; bold curve, ANN 1000.

response to seeding during DJF for a given areal extent may be explained by the enhanced susceptibility of the more pristine clouds of the Southern Hemisphere. In the NCAR model, optimal cloud seeding over 25 per cent of the ocean surface might produce a net cooling close to 3.5 or 4 W m^{-2} in DJF if the cloud drop number concentration is 375 or 1000 cm^{-3}, respectively. Following this same strategy, weaker cooling is expected in JJA (approximately 2.5 W m^{-2}) in both geoengineering experiments. The reasons for the forcing reaching a maximum for cloud fractions below 100 per cent (figure 5) are still under investigation.

The corresponding optimal locations based on this ranking are displayed in figure 6 based on the N_C=375 cm^{-3} experiment for two choices of seeding area. The results indicate that the preferential locations for cloud seeding depend strongly on season. The optimal areas in the summer hemisphere occur first in marine stratus, and shallow trade cumulus regions, and secondarily in mid-latitude storm track regions. Both regions would need to be seeded to reach forcing amplitudes that could balance that associated with a doubling of CO_2.

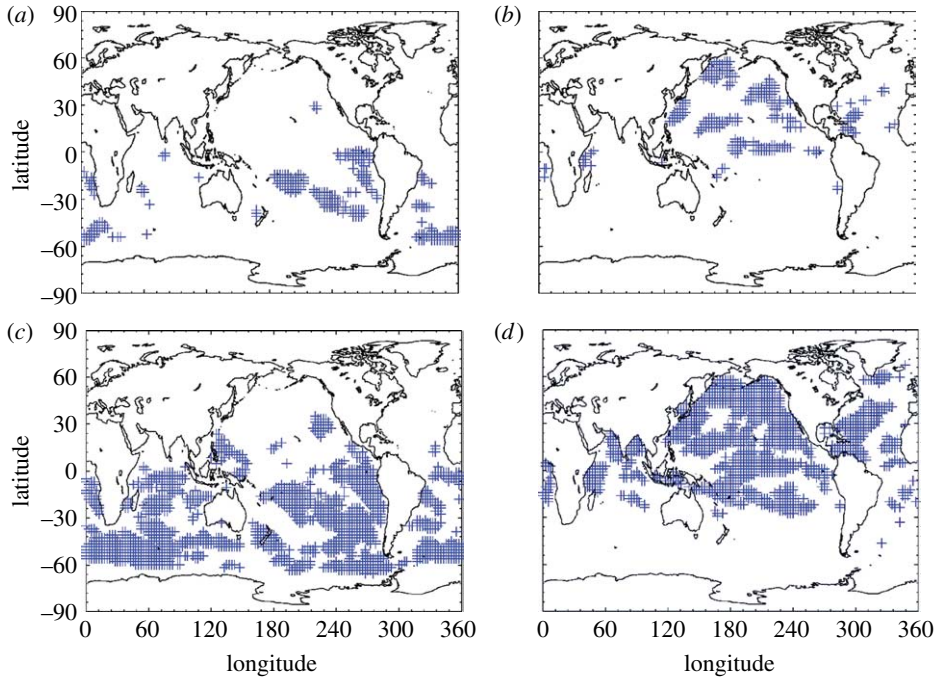

Figure 6. Optimal geographical locations for warm cloud seeding based on the $N_C = 375\ \mathrm{cm}^{-3}$ experiment. (*a*) 5% of the ocean area in DJF, (*b*) 5% of the ocean area in JJA, (*c*) 25% of the ocean area in DJF and (*d*) 25% of the ocean area in JJA.

5. Technological implications of the foregoing calculations

The calculations and computations presented in §§2–4 yield some significant implications—outlined below—with respect to technological aspects of the global temperature stabilization technique.

(i) The sensitivity studies (§2) show that the albedo changes ΔA are insensitive, over a wide range, to the values of salt mass m_s. It follows that the choice of disseminated droplet size can—to a considerable extent—be dictated by technological convenience. These studies also indicate that it would be optimal for albedo enhancement to confine our salt masses within the range 10^{-17}–10^{-15} kg, corresponding to seawater droplets in the approximate size range 0.8–4 μm. This is because smaller particles may not be nucleated and larger ones could act as UGN and thus perhaps promote drizzle onset and concomitant cloud dissipation. Thus, it seems sensible to disseminate seawater droplets of diameter approximately 0.8 μm, thus minimizing the required volumetric flow rate.

(ii) Monodispersity of the seawater aerosol, within the above-mentioned size range, has little impact on the values of ΔA. However, it remains desirable because it could enhance cloud stability and therefore longevity.

(iii) The calculations presented earlier indicate that optimal seeding of all suitable maritime clouds may produce values of globally averaged negative forcing ΔF of at least $-3.7\ \mathrm{W\ m^{-2}}$. If so (see discussion of

uncertainties in §6), the areal fraction of suitable cloud cover seeded, f_3, in order to maintain global temperature stabilization, could—for a period of some decades—be appreciably lower than unity, thus rendering less daunting the practical problem of achieving adequate geographical dispersal of disseminated CCN.

(iv) It follows from (iii) that there exists, in principle, latitude to: (a) avoid seeding in regions where deleterious effects (such as rainfall reduction over adjacent land) are predicted; (b) seed preferentially in unpolluted regions, where the albedo changes ΔA for a fixed value of ΔN are a maximum.

(v) The high degree of seasonal variability in the optimal geographical distributions of suitable cloud (§4) underlines the desirability of a high degree of mobility in the seawater aerosol dissemination system.

6. Issues requiring further study

In addition to requiring further work on technological issues concerning the cloud albedo enhancement scheme (Salter *et al.* 2008), we need to address some limitations in our understanding of important meteorological aspects, and also make a detailed assessment of possibly adverse ramifications of the deployment of the technique, for which there would be no justification unless these effects were found to be acceptable. Some of these issues were addressed by Latham (2002) and Bower *et al.* (2006), and so are not examined herein. Others are now outlined.

(i) It was assumed in the specimen calculations (§3) that approximately half of the seawater droplets disseminated near the ocean surface would be transported upwards by turbulent air motions to enter suitable clouds and form additional cloud droplets, i.e. $f_4 = 0.5$. In actuality, f_4 will probably vary considerably according to the meteorological situation. Thus, we need to obtain reliable estimates of f_4 for all situations of interest. Airborne measurements (Smith *et al.* 1993) and estimates based on bubble-bursting studies (Blanchard 1969) suggest that f_4 is greater than 0.1. D. H. Lenschow and M. H. Smith (2008, personal communications) suggest that the fraction will be close to 0.5. It appears that the calculated spraying rates (§3) are readily achievable technologically, and could easily be increased to accommodate any likely value of f_4.

(ii) It may prove useful to examine the possibility of charging the seawater droplets and harnessing the Earth's electric field to help transport them to cloud base.

(iii) If our technique were to be implemented, global changes in the distributions and magnitudes of ocean currents, temperature, rainfall and wind would result. Even if it were possible to seed clouds relatively evenly over the Earth's oceans, so that the effects of this type could be minimized, they would not be eliminated. Also, the technique would still alter the land–ocean temperature contrast, since the radiative forcing produced would be only over the oceans. In addition, we would be attempting to neutralize the warming effect of vertically distributed

greenhouse gases with a surface-based cooling effect, which could have consequences such as changes in static stability, which would need careful evaluation. Thus, it is vital to engage in a prior assessment of associated climatological and meteorological ramifications, which might involve currently unforeseen feedback processes.

It is important to establish the level of local cooling which would have significant effects on ocean currents, local meteorology and ecosystems. This will require a fully coupled ocean/atmosphere climate system model.

(iv) R. Wood (personal communication) states that an important recently identified aspect of the aerosol–cloud–climate problem for low clouds is that the macrophysical properties of the clouds respond to changes in aerosol concentration in ways not foreseen at the time of formulation of the Albrecht effect (i.e. that reduction in warm rain production—resulting from increasing cloud droplet concentration and reduced droplet size—leads to thicker clouds). For marine stratocumulus clouds, recent studies with a Large Eddy Simulation (LES) model (Ackerman *et al.* 2004) and a simple mixed layer model (Wood 2007) show that the response of the cloud liquid-water-path on relatively short timescales (less than 1 day) is a balance between moistening of the MBL due to precipitation suppression, which tends to thicken the cloud, and drying by the increased entrainment associated with the extra vigour that a reduction in precipitation content brings to the MBL. Under some conditions the clouds thicken, and under others the clouds thin. Thus, it is unjustifiably simplistic to assume that adding CCN to the clouds will always brighten them according to the Twomey equation. Also, even without precipitation, LES studies (e.g. Wang *et al.* 2003; Xue & Feingold 2006) show that the enhanced water vapour transfer rates associated with smaller, more numerous droplets can lead to feedbacks on the dynamics that tend to offset, to some extent, the enhanced reflectivity due to the Twomey effect. These effects are either not treated or are poorly treated by GCM parametrizations of clouds and boundary-layer processes. It is clearly critical to an authoritative assessment of our scheme to conduct a full quantitative examination of them. IPCC (2007) has stressed the importance and current poor understanding of aerosol–cloud interactions.

Other refinements or extensions that need to be made to current work on the albedo-enhancement idea include the following: inclusion of direct aerosol and long-wave radiative effects; examination of the lateral dispersion of aerosol from its dissemination sites; further estimation of the areal coverage of suitable maritime clouds; estimation of aerosol lifetimes—both within and outside of clouds—and the fraction of disseminated aerosol particles that enter suitable clouds. A rough comparison of the amounts of salt entering the MBL from natural processes and via seeding indicates that if the scheme were in full operation, i.e. spraying enough seawater to balance the warming associated with CO_2 doubling, seeding would contribute less than 10 per cent of the total salt. In the first few decades of operation, the amount of disseminated salt would be several orders of magnitude less than that produced naturally.

7. Discussion

It follows from the discussion in §6—particularly items 3 and 4—that although two separate sets of GCM computations (§4) agree in concluding that this cloud seeding scheme is in principle powerful enough to be important in global temperature stabilization, there are important clearly defined gaps in our knowledge which force us to conclude that we cannot state categorically at this stage whether the technique is in fact capable of producing significant negative forcing. There are also currently unresolved technological issues (Salter *et al.* 2008—companion paper).

If it is found that the unresolved issues defined in §6 (especially item 4) do not yield the conclusion that the cloud albedo seeding technique is much weaker than is estimated from the GCM computations, we may conclude that it could stabilize the Earth's average temperature T_{AV} beyond the point at which the atmospheric CO_2 concentration reaches 550 ppm but probably not up to the 1000 ppm value. The corresponding amount of time for which the Earth's average temperature could be stabilized depends, of course, on the rate at which the CO_2 concentration increases. Simple calculations show that if it continues to increase at the current level, and if the maximum amount of negative forcing that the scheme could produce is $-3.7\,\mathrm{W\,m^{-2}}$, T_{AV} could be held constant for approximately a century. At the beginning of this period, the required global seawater dissemination rate $\mathrm{d}V/\mathrm{d}t$ (if $f_3=1$) would be approximately $0.14\,\mathrm{m^3\,s^{-1}}$ initially, increasing each year to a final value of approximately $23\,\mathrm{m^3\,s^{-1}}$.

Recent experimental studies of both the indirect and direct aerosol effects involving data from the MODIS and CERES satellites (Quaas *et al.* 2006, 2008) have led to a study by Quaas & Feichter (2008) of the quantitative viability of the global temperature stabilization technique examined in this paper. They concluded that enhancement (via seeding) of the droplet number concentration in marine boundary-layer cloud to a uniform sustained value of $400\,\mathrm{cm^{-3}}$ over the world oceans (from 60° S to 60° N) would yield a short-wave negative forcing of $-2.9\,\mathrm{W\,m^{-2}}$. They also found that the sensitivity of cloud droplet number concentration to a change in aerosol concentration is virtually always positive, with larger sensitivities over the oceans. These experimental results are clearly supportive of our proposed geoengineering idea, as is the work of Platnick & Oreopoulos (2008) and Oreopoulos & Platnick (2008), which also involves MODIS satellite measurements.

Further encouraging support for the quantitative validity of our global temperature stabilization scheme is provided by the field research of Roberts *et al.* (2008) in which—for the first time—the enhancement of albedo was measured on a cloud-by-cloud basis, and linked to increasing aerosol concentrations by using multiple, autonomous, unmanned aerial vehicles to simultaneously observe the cloud microphysics, vertical aerosol distribution and associated solar radiative fluxes. In the presence of long-range transport of dust and anthropogenic pollution, the trade cumuli have higher droplet concentrations, and are on average brighter, the observations indicating a higher sensitivity of radiative forcing by trade cumuli to increases in cloud droplet concentrations than has been reported hitherto. The aerosol–cloud forcing efficiency was as much as $-60\,\mathrm{W\,m^{-2}}$ per 100 per cent cloud fraction for a

doubling of droplet concentrations and associated increase in liquid water content; the accompanying direct top of the atmosphere effect of this elevated aerosol layer was found to be $-4.3\ \mathrm{W\,m^{-2}}$.

Our view regarding priorities for work in the near future is that we should focus attention on outstanding meteorological issues outlined earlier in this paper, particularly in §6, as well as technological ones described in our companion paper. At the same time, we should develop plans for executing a limited-area field experiment in which selected clouds are inoculated with seawater aerosol, and airborne, ship-borne and satellite measurements are made to establish, quantitatively, the concomitant microphysical and radiative differences between seeded and unseeded adjacent clouds and thus, hopefully, to determine whether or not this temperature-stabilization scheme is viable. Such further field observational assessment of our technique is of major importance.

Advantages of this scheme, if deployed, are that (i) the amount of cooling could be controlled—by measuring cloud albedo from satellites and turning disseminators on or off (or up and down) remotely as required, (ii) if any unforeseen adverse effect occurred, the entire system could be switched off instantaneously, with cloud properties returning to normal within a few days, (iii) it is relatively benign ecologically, the only raw materials required being wind and seawater, and (iv) there exists flexibility to choose where local cooling occurs, since not all suitable clouds need to be seeded.

A further positive feature of the technique is revealed by comparing the power required to produce and disseminate the seawater CCN with that associated with the additional reflection of incoming sunlight. As determined in the companion paper, approximately 1500 spray vessels would be required to produce a negative forcing of $-3.7\ \mathrm{W\,m^{-2}}$. Each vessel would require approximately 150 kW of electrical energy to atomize and disseminate seawater at the necessary continuous rate (as well as to support navigation, controls, communications, etc.), so that the global power requirement is approximately 2.3×10^{8} Watts. Ideally, this energy would be derived from the wind. The additional rate of loss of planetary energy, resulting from cloud seeding, required to balance the warming caused by CO_2 doubling would be $\Delta F\cdot A_{\mathrm{E}}=-1.9\times10^{15}$ W. Thus, the ratio of reflected power to required dissemination power is approximately 8×10^{6}. This extremely high 'efficiency' is largely a consequence of the fact that the energy required to increase the seawater droplet surface area by four or five orders of magnitude—from that existing on entry to the clouds to the surface area achieved when reflecting sunlight from cloud top—is provided by nature.

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