Measurements of Cloud Susceptibility

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ABSTRACT

The sensitivity of warm stratocumulus cloud albedo to changes in droplet concentration, termed "cloud susceptibility," is calculated using data from the UKMO Meteorological Research Flight. Stratocumulus clouds in the eastern Pacific, South Atlantic, subtropical regions of the North Atlantic, and around the British Isles are studied. The range of susceptibility measured is large and maritime clouds are shown to have the largest susceptibility. Numerical simulations of the changes in cloud radiative and microphysical properties with increasing droplet concentration are carried out. These highlight the high sensitivity of maritime clouds to changes in droplet concentration and the rapid reduction in sensitivity as the cloud droplet concentration increases.

1. Introduction

Marine stratocumulus clouds have a high albedo relative to the underlying sea surface and so are likely to have a major influence on both local and, by virtue of their extensive coverage, global radiation balance. Of great interest to those studying greenhouse gas–induced climate change is how these large cloud systems will change in a warmer world. Will their microphysical characteristics and hence radiative properties change so as to enhance or reduce the greenhouse warming?

Clouds occur in the atmosphere through the condensation of water vapor onto populations of aerosols that can act as cloud condensation nuclei (CCN). For a given available water content, the mean droplet size is inversely proportional to the number of CCN that can be activated to form cloud drops. In a region of low CCN concentrations, droplets will tend to grow to a larger size because of the lack of competition for the liquid water.

The albedo of a cloud is dependent on the liquid water content of the cloud and the size of its droplets. Twomey (1974) pointed out that increasing pollution beneath a cloud layer generally leads to an increase in the CCN concentrations, hence increasing the number of cloud droplets.

The size of droplets is crucial in determining their radiative properties in terms of scattering and absorption. Smaller droplets will scatter more radiation in the visible and near-infrared regions of the spectrum. The albedo of a cloud that consists of small droplets will therefore be larger than the albedo of a cloud with the same liquid water content composed of larger droplets. Obviously any future modification of droplet size by either natural or anthropogenic aerosol concentration changes may have large effects on the cloud albedo, and hence the net effect of clouds on the earth radiation balance. Slingo (1990) carried out a sensitivity study using a three-dimensional general circulation model. He found that the top-of-atmosphere radiative forcing by doubled carbon dioxide concentrations can be balanced by increases of 15%–20% in amounts of low-level cloud or 20%–35% in liquid water path, or by decreases of 15%–20% in mean drop radius.

The study of cloud albedo changes induced by pollution is discussed in Twomey (1977); he shows that by increasing droplet concentration, and therefore the optical thickness of a cloud, pollution acts to increase the albedo and that the increased absorption coefficient resulting from the pollution, due to the aerosol itself, becomes dominant only for very optically thick clouds. It was suggested by Charlson et al. (1987) that dimethylsulphide (DMS) production may increase in a warmer climate. DMS is believed to be a major source of CCN in maritime air masses; hence, any increase in DMS and CCN, which may result in increased cloud albedo, could be an important feedback in climate change. The effects of anthropogenic aerosol on cloud microphysics are clearly shown by the ship track phenomenon, which has been studied with aircraft and satellite observations (Coakley et al. 1987; Radke et al. 1989).

Platnick and Twomey (1992, hereafter referred to as PT92) defined a quantity called the cloud susceptibility that relates the sensitivity of cloud albedo to changes in droplet concentration (liquid water content remain-

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ing constant). Platnick and Twomey showed that not all clouds are equally susceptible; the determining factors turned out to be droplet size and optical thickness.

The United Kingdom Meteorological Office (UKMO) Meteorological Research Flight (MRF) has for many years been actively studying the radiative and microphysical properties of marine stratocumulus. The MRF has built up an extensive library of cloud radiative and microphysical properties covering a wide range of conditions. In this paper the susceptibility of marine stratocumulus clouds is calculated using this library of data. Simple numerical studies of the impact of increasing droplet concentration on the cloud radiative and microphysical properties are then carried out using the observational data as initial conditions.

2. Cloud susceptibility

The property "cloud susceptibility" is discussed in detail in Twomey (1991), Platnick (1991), and PT92. If the cloud droplet concentration, N, and the broadband visible (0.3–0.7 μm) cloud albedo, A, are known, then given a change in droplet concentration ΔN, a measurement can be made to find the change in cloud albedo, ΔA. Under the condition of constant liquid water the quantity dA/dN is termed the cloud susceptibility. Since in general, $A = A(\omega_0, \tau, g)$ where ω_0 is the single scatter albedo, τ the optical thickness, and g the asymmetry factor, the derivative can be expressed in the form:

$$\frac{dA}{dN} = \frac{\delta A}{\delta \tau}\frac{\delta \tau}{\delta N} + \frac{\delta A}{\delta \omega_0}\frac{\delta \omega_0}{\delta N} + \frac{\delta A}{\delta g}\frac{\delta g}{\delta N}. \quad (1)$$

Following Twomey (1977), we assume that all components in the aerosol increase together and in the same proportion, so that the increase in cloud condensation nuclei and in aerosol absorption is proportional. Then, if the liquid water content remains constant irrespective of the increase in aerosol, the optical thickness, which is dominated by scattering, must then increase as $N^{1/3}$.

Furthermore, under the conditions of conservative scattering, which is applicable in the visible, $\omega_0 = 1$ and g is approximately constant with radius so Eq. (1) reduces to (under the condition of constant liquid water)

$$\frac{dA}{dN} = \frac{\delta A}{\delta \tau}\frac{\tau}{3N} = \frac{4\pi\rho_l}{9L}\tau\frac{\delta A}{\delta \tau}r_v^3. \quad (2)$$

Here ρ_l is the liquid water density, L is the liquid water content of the cloud, and r_v the mean volume radius.

Furthermore, it is shown in Twomey (1991) that the term $\tau\delta A/\delta\tau$ is given by the two-stream approximation as $A(1 - A)$, which has a maximum at $A = 0.5$.

A susceptibility of 0.01 cm^3 means that the increase in droplet density of one drop per cubic centimeter will increase albedo by 0.01.

The assumption that all components in the aerosol increase together and in the same proportion, leading to $\tau \propto N^{1/3}$, fundamental in the definition of cloud susceptibility, is discussed in Twomey (1977). Twomey points out there is an implicit assumption that the absorption is not greatly modified when a cloud forms; that will be the case if absorption is by particles that remain unchanged after cloud formation. The results of Prishivalko and Astafyeva (1974) suggest that even when droplets form on insoluble aerosol the overall absorption is not profoundly changed.

Martin et al. (1994) showed that the mean volume radius r_v and the effective radius r_e are related in the following way:

$$r_v^3 = kr_e^3, \quad (3)$$

where

$$r_e = \frac{\int_0^\infty n(r)r^3dr}{\int_0^\infty n(r)r^2dr}. \quad (4)$$

Here k is a constant, which has the value $k = 0.8 \pm 0.07$ in maritime air masses or $k = 0.67 \pm 0.07$ in continental air masses. In these calculations, the type of air mass and hence the value of k has been determined from the aerosol concentration using the criteria of Martin et al. (1994).

In this work the susceptibility is calculated under the conditions of constant liquid water using the expression:

$$\frac{dA}{dN} = \frac{4\pi\rho_l}{9L}A(1 - A)kr_e^3 \quad (5)$$

from measurements of L, A, and r_e. Terms L and r_e are measured from a straight and level run near cloud top and A is measured from a run above cloud vertically stacked above the in-cloud measurement run. In a later stage, the susceptibility is plotted as a function of aerosol concentration, the aerosol concentration being measured during a run just below cloud base.

The albedo of a cloud field will also change in time due to variations in solar zenith angle. In this work the observations are taken at a range of different solar zenith angles, and no allowance for differences due to these effects is made.

3. Aircraft instrumentation

The results described below have used data obtained by the MRF C130 Hercules during

- the First ISCCP (International Satellite Cloud Climatology Project) Regional Experiment (FIRE), off the west coast of California in July 1987,

- the First ATSR (Along Track Scanning Radiometer) Tropical Experiment (FATE), in the South Atlantic (8°S, 14°W) during November 1991,
- flights made around the British Isles between December 1990 and February 1992, and
- the Atlantic Stratocumulus Transition Experiment (ASTEX) in the vicinity of the Azores (37°N, 25°W) in June 1992.

In all, data from 33 flights have been analyzed.

The C130 Hercules of the MRF is extensively equipped for atmospheric research. Visible fluxes are calculated as the difference between two Eppley pyranometers, one having a clear dome with transmission between 0.3 and 3.0 μm and the other having a red dome that transmits between 0.7 and 3.0 μm. The pyranometers have a hemispherical field of view, and the cloud albedo is derived from the ratio of upward and downward views of pairs of pyranometers. The fluxes from these instruments are corrected for the aircraft pitch and roll, measured by an inertial navigation system, following the method of Saunders et al. (1992). For the purpose of cloud susceptibility calculations only, unbroken stratocumulus cloud sheets were used and all measurements are taken away from cloud edges. Due to uncertainty in the accuracy of the pyranometers during the FIRE campaign an albedo of $A = 0.5$ was assumed for all FIRE flights, which gives the maximum value of $A(1 - A) = 0.25$. The albedo of stratocumulus throughout all of the flights is observed to be in the range 0.2 to 0.8, which results in a minimum observed value of $A(1 - A) = 0.16$. The error associated with assuming $A = 0.5$ for the FIRE data is therefore small, the dominant term in the equation for susceptibility being r_e^3.

The effective radius of the cloud droplet distribution is measured using a Particle Measuring System (PMS) Forward Scattering Spectrometer Probe (FSSP). The FSSP measures droplets in 15 size bins and is normally operated to detect droplets with diameters in the size range $d = 2$–47.5 μm from which the effective radius r_e and droplet concentration N can be deduced. Liquid water content, L, is also measured with the FSSP using the relationship:

$$L = \frac{4}{3}\pi\rho_l \sum_{n=1}^{15} r_n^3 N(n), \qquad (6)$$

where n is the channel number. The r_e in stratocumulus clouds typically increases as with height to a maximum near the top. It is this cloud-top r_e that has been used in the calculations of cloud susceptibility as the albedo is dependent on this cloud-top value.

In the results presented, the cloud susceptibility is shown as a function of aerosol concentration (0.1–3.0 μm diameter) measured by a PMS Passive Cavity Aerosol Spectrometer Probe (PCASP). The aerosol concentration is taken as the mean concentration in a run just below cloud base. The PCASP was not fitted to the C130 for the FIRE campaign; here the aerosol concentration was estimated from droplet concentration using the expression for maritime air given by Martin et al. (1994).

4. The variation of cloud microphysical parameters

Figure 1 shows histograms of the mean cloud droplet effective radius at cloud top measured with the FSSP partitioned into geographical locations (a) the British Isles, (b) the Azores, (c) off the coast of California, and (d) the South Atlantic. Although the total number of observations made in the different locations is not the same there are some interesting variations in the distribution of the microphysical parameters.

The range of r_e observed around the British Isles is large, indicative of the varied air masses that can be observed in such midlatitude regions. There is a small peak in the r_e distribution between 6 and 8 μm, the

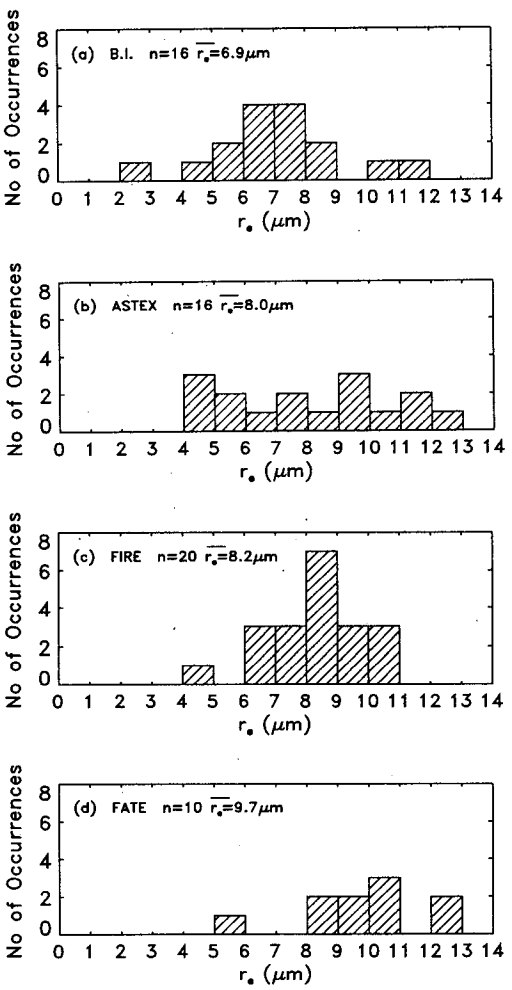

FIG. 1. Frequency histogram of r_e at cloud top for (a) around the British Isles, (b) ASTEX, (c) FIRE, and (d) FATE. The total number of observations n and the mean effective radius, $\bar{r}_e$, are given.

mean value of r_e being 6.9 μm. The measurements made around the Azores also show a large spread of r_e with values from 4 to 13 μm but with no pronounced peak, the mean r_e being 8.0 μm. The stratocumulus observed around the Azores was generally quite thin and appeared to be fed by large convective cumulus cells that penetrated the stratocumulus sheet. The range of r_e observed is likely therefore to be very dependent on the averaging carried out and, in particular, how many of these cumulus cells were penetrated during a straight and level run at cloud top. The large range of r_e observed is also due to the wide range of airmass types observed during this experiment.

Measurements off the coast of California show a large peak in the 8–9-μm region, and with the exception of the one observation at 4–5 μm, which was observed on a day with very thin stratocumulus, the range of r_e observed is from 6 to 11 μm. The mean r_e observed off California was 8.2 μm. The larger values of r_e seen off California are characteristic of the relatively clean unpolluted air mass in this region. The observations from the South Atlantic show a tendency towards larger r_e than the other regions with little evidence of a peak and quite a large spread of values, the mean r_e being 9.7 μm (although it is noted that there are fewer observations here). The largest r_e observed in all the datasets was in the South Atlantic, when during a profile an r_e of 16 μm was observed; this is not shown in the run average values presented. The mean r_e is largest for the relatively clean stratocumulus clouds observed in the southern Atlantic and is smallest for the clouds measured around the British Isles. The range of the mean r_e observed is 2.8 μm. It should be noted that the work of Slingo (1990) showed that a reduction in r_e from 10 μm to between 7.9 and 8.6 μm was large enough to balance the change in the top-of-atmosphere radiative forcing due to a doubling of CO_2. These results show that the geographical variability of r_e is larger than this modeled reduction, enforcing the point that r_e must be able to vary within a climate model if the radiative properties of stratocumulus clouds are to be accurately represented.

The magnitude of L, observed in the data, also shows some geographical variation (Fig. 2). The values of L around the British Isles are on the whole smaller than those observed elsewhere. The Azores data show a clear mix between values similar to those around the British Isles and larger values of the same magnitude as those observed off California and in the South Atlantic.

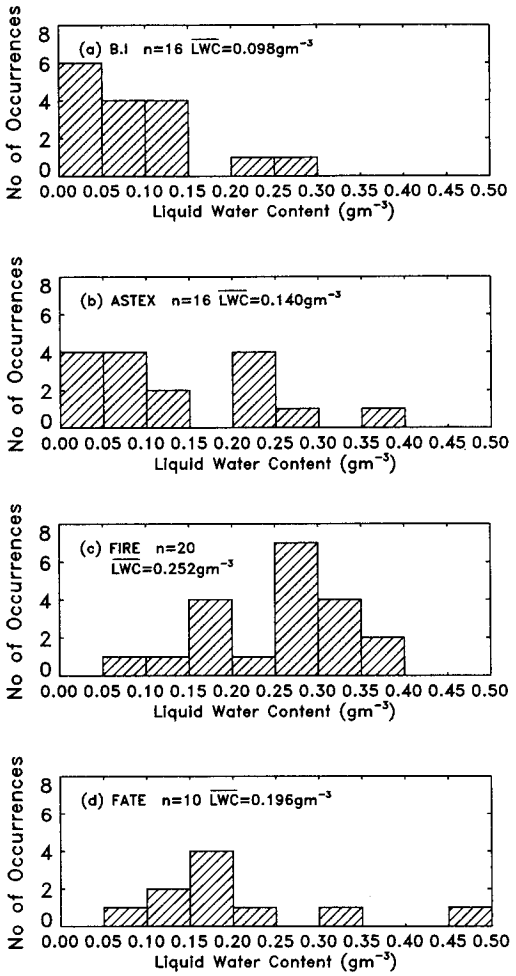

FIG. 2. Frequency histogram of L at cloud top for (a) around the British Isles, (b) ASTEX, (c) FIRE, and (d) FATE. The total number of observations n and the mean liquid water, $\overline{LWC}$, are given.

5. The variability of susceptibility

Cloud susceptibility has been calculated for several flights at various locations around the world. Figure 3 shows the distribution of cloud susceptibility at each geographical location. The susceptibility is seen to vary from 0.1 to >8.0 ($\times 10^{-3}$ cm^3) in broad agreement with the magnitudes estimated from satellite measurements by PT92.

The data from around the British Isles show very large variations in cloud susceptibility with a maximum value in excess of 8×10^{-3} cm^3. This is a sign of the wide range of air masses found around the British Isles with both very dirty continental air off mainland Europe and clean maritime air from the North Atlantic having been sampled.

The measurements made during ASTEX, around the Azores, show a wide variation of susceptibility with values from 0.5 to 5 ($\times 10^{-3}$ cm^3). This is somewhat surprising, considering the large distance from the continent of the Azores, and is the result of some major incursions of continental air from Europe into the area. This highlights the far reaching effects of continental pollution on maritime stratocumulus.

The measurements off California and in the South Atlantic are mostly characteristic of maritime air

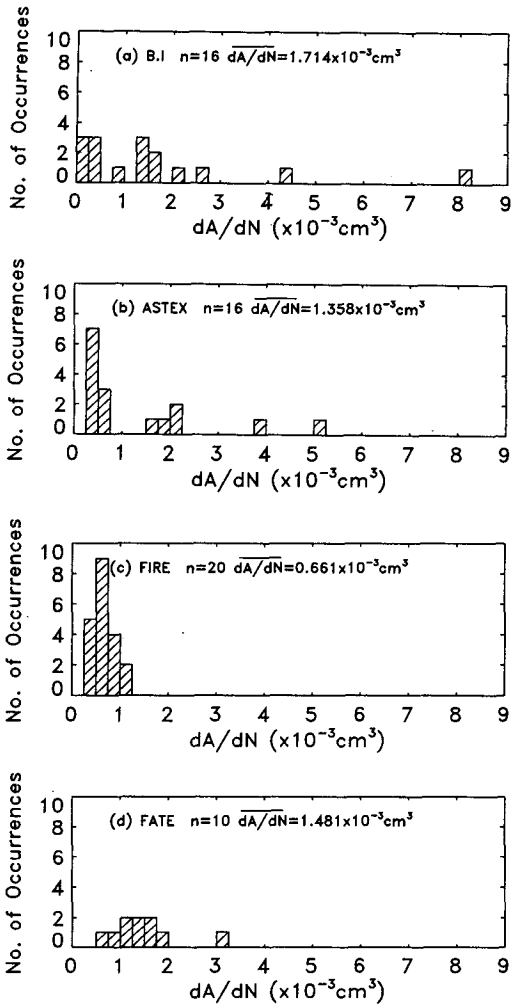

FIG. 3. Frequency histogram of dA/dN at cloud top for (a) around the British Isles, (b) ASTEX, (c) FIRE, and (d) FATE. The total number of observations n and the mean susceptibility, $\overline{dA/dN}$, are given.

masses with the cloud susceptibility varying between 0.5 and ~3 ($\times 10^{-3}$ cm^3), with the data from the South Atlantic showing on the whole larger susceptibilities than those found off the west coast of California.

Figure 4 shows the cloud susceptibility as a function of the aerosol concentration beneath the cloud sheet, the inset box being an enlargement of the region bounded by the dashed line.

Stratocumulus decks in a wide range of air masses have been sampled with aerosol concentrations varying from 1789 to 23 cm^{-3}. Of importance is the observation that the cloud susceptibility tends to a value of around 0.5 $\times$ 10^{-3} cm^3 for aerosol concentrations in excess of 500 cm^{-3}; this is to say that once the air mass has a high aerosol load, increasing the droplet concentration by injecting more aerosol into the air mass has little effect on the cloud susceptibility. However, for the clean maritime air masses there is a very strong gradient of susceptibility with aerosol concentration indicating that a relatively small increase in aerosol can have a large effect on the cloud susceptibility. The susceptibility has its largest values for aerosol concentrations below 300 cm^{-3}, which suggests that only clean maritime clouds will suffer a significant modification in their albedo given an increase in droplet concentration. Hence, further anthropogenic pollution of an already dirty air mass (i.e., one with a high aerosol loading) will have little effect.

6. Simulations of albedo changes

The calculations of cloud susceptibility have shown a large range of values indicating considerable variation in cloud sensitivity to changes in droplet concentration. In particular, we have seen that the gradient of susceptibility with aerosol concentration (Fig. 4) is very large at low aerosol concentrations (maritime air) and decreases tending to zero for high aerosol concentrations (continental air). These observations indicate that the initial effects on the cloud properties of introducing aerosol into a clean maritime air mass are likely to be quite dramatic, whereas introducing more aerosol into a continental air mass is likely to have little effect on the cloud properties.

In order to study these effects, a simple numerical simulation has been carried out using the observational data as initial conditions. In these simulations the liquid water content of the cloud has been kept constant at the initial value and the droplet concentration has been increased in steps of 1 cm^{-3}. Using a prescribed value of k and assuming a constant liquid water content, L, a new r_e was calculated given the change in N using the relation:

$$L = \frac{4}{3}\pi\rho_l k r_e^3 N. \qquad (7)$$

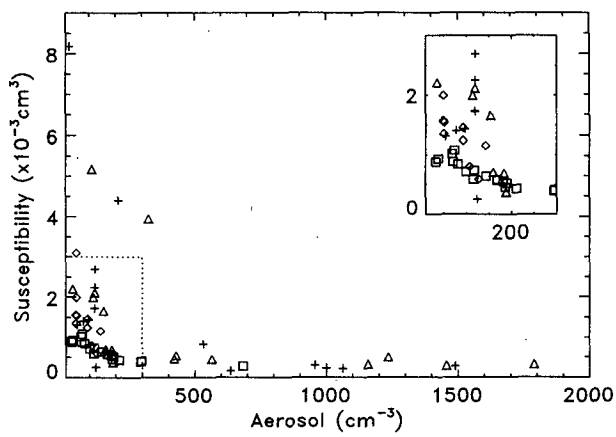

FIG. 4. The cloud susceptibility as a function of aerosol concentration below cloud base using aircraft measurements, where + is data from around the British Isles, ◇ from the FATE, □ from the FIRE, and △ from the ASTEX. The inset is an enlargement of the region bounded by the dashed line.

TABLE 1. Initial cloud conditions for simulations.

Flight details	Susceptibility ($\times 10^{-3}$ cm^3)	Aerosol concentration (cm^{-3})	Droplet concentration (cm^{-3})	r_e (μm)	L (g m^{-3})	Albedo	k
ASTEX A206	1.99	115	33.73	9.5	0.097	0.28	0.80
ASTEX A212	0.30	1456	281.29	5.2	0.111	0.52	0.67
ASTEX A211	5.17	110	15.47	10.1	0.053	0.60	0.80
ASTEX A213	0.32	1789	208.29	8.5	0.359	0.72	0.67

For each increment the cloud susceptibility was calculated using a new albedo, modified using the cloud susceptibility calculated at the previous concentration, and the new value of r_e.

In these simple numerical simulations the value of k has been fixed depending on the initial aerosol concentrations following the criteria of Martin et al. (1994). This condition of fixing k essentially applies the condition that the new aerosol added to the cloud, which increases the droplet concentration, is of the same type as the background aerosol. In reality it is possible for a new aerosol type, with some different chemical constituency, to be introduced to the system. The constant k is a strong function of both droplet concentration and spectral shape, the spectral shape being governed by the chemistry of the CCN. Therefore, the assumption that k is fixed in these simulations is a simplification; however, this simple experiment will still give insight into the changes in albedo to be expected if the droplet concentration is increased. In these simulations the cloud albedo input at the start is that measured by the aircraft; it therefore represents a snapshot of the cloud at that time, and any future modification of albedo due to changing solar zenith angle is not accounted for. The simple model uses a single layer cloud of uniform liquid water content, droplet concentration, and effective radius.

Table 1 shows the initial conditions for the four simulations carried out. The four cases studied here are all taken from the ASTEX dataset. These observations were made in the vicinity of the Azores in summer 1992. All the observations discussed were made between 35° and 37°N, 19° and 25°W, a distance of nearly 1000 km from the nearest major landmass.

Flight A212 (measurements at 1630 UTC 16 June 1992) had very high aerosol concentrations below the cloud (1456 cm^{-3}). In contrast, flight A206 (measurements at 1530 UTC 5 June 1992) was in a maritime air mass with an aerosol concentration of 115 cm^{-3}.

Figure 5 shows 950-mb back trajectories for a point in the boundary layer during flights A212 and A206; the arrowheads are plotted every 12 hours. The back

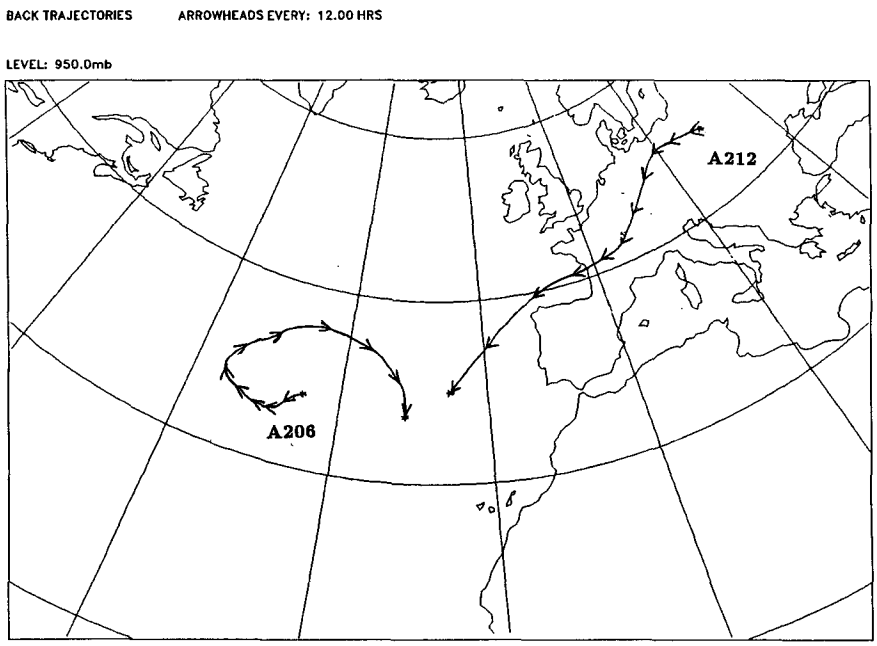

FIG. 5. Back trajectories (at 950 mb) for the boundary-layer air in flight A212 on 16 June 1992 and A206, 5 June 1992, with arrowheads every 12 hours.

trajectories were calculated from the operational analyses of the UKMO Unified Model. We can see that the air below the cloud in flight A212 has originated from northern Europe with a sea track of ~1600 km. Despite this long sea track this air still maintained a strong continental influence with aerosol concentrations in the boundary layer in excess of 1000 cm^{-3}. In contrast, the trajectory for flight A206 shows that the air mass has been confined to the ocean for the last 120 hours and can therefore be considered maritime.

The cloud-top height of the cloud sheets sampled in flights A206 and A212 was similar at around 900 m and the liquid water at cloud top, L, was also similar for the two cases (Table 1). These two flights (A206 and A212) therefore act as a good example of the effects of increasing the aerosol concentration in the boundary layer on clouds that are otherwise almost the same.

Flight A211 (15 June 1992) was in a maritime air mass with low aerosol concentrations and a large effective radius of 10.1 μm; flight A213 (19 June 1992) took place in a continental air mass but the effective radius was also quite large at 8.5 μm. The major difference between these two cloud sheets lay in their cloud-top liquid water contents, 0.053 g m^{-3} in A211 and 0.359 g m^{-3} in A213. These two cases will therefore act as a good test of the sensitivity of cloud susceptibility to different liquid water contents.

Figures 6a and 6b show the cloud susceptibility as a function of increasing droplet concentration for the four cases described in Table 1; (c) and (d) show the change in r_e with increasing droplet concentration and (e) and (f) the resulting changes in albedo.

Considering first the two flights A206 and A212 (Figs. 6a, 6c, and 6e). The solid line refers to flight A206; the dotted line to flight A212. As can be seen, the cloud susceptibility for flight A206 is initially higher than that for A212 but decreases rapidly with increasing droplet concentration, and the two flights have almost the same susceptibility once the droplet concentration has increased beyond 280 cm^{-3}. The r_e for the A206 case starts at 9.5 μm but quickly falls as the droplet concentration increases and results in values of r_e less than those observed in the continental cloud, A212, this being due to the slightly lower cloud-top liquid water content in A206 to that in A212. The albedo of the A206 cloud is initially 0.28, but this increases rapidly to around 0.4 for droplet concentrations of 200 cm^{-3} and would appear to be tending to some asymptotic limit > 0.5 as the droplet concentration increases past 600 cm^{-3}. The albedo of the A212 case is higher than that of the A206 cloud with an initial value of 0.52; the initial rate of increase of albedo, however, is considerably less with the albedo asymptoting to a value of around 0.59 for high droplet concentrations. Of importance is the very rapid change in albedo for relatively small increases in droplet concentration. This would suggest that the impact of polluting a maritime air mass even by a small amount is quite considerable. Also it suggests that phenomena like ship tracks will occur only where the air mass is initially very clean; injecting more aerosol (and hence increasing droplet concentration) will have a marked effect on cloud albedo only when the initial droplet concentration is low.

The comparison of cases A211 and A213 (Figs. 6b, 6d, and 6f) shows similar characteristics to that of A206 and A212. The cloud susceptibility for the maritime air mass A211 is initially very high and falls rapidly as the droplet concentration increases. The changes in effective radius are also very large, with a decrease in droplet size of 50% occurring for an increase in droplet concentration from ~12 to ~100 cm^{-3}. The susceptibility for the cloud in A213 shows only a gradual decrease with increasing droplet concentration with a correspondingly less dramatic change in effective radius. The change in cloud albedo for the A211 case is very dramatic with an initial rapid increase in albedo from 0.60 to 0.70 for an increase in droplet concentration from 12 to 60 cm^{-3}; the rate of change of cloud susceptibility and hence albedo then decreases, eventually tending to an asymptotic value of 0.83.

7. Discussion and conclusions

In view of the large range of cloud droplet effective radius observed, it is apparent that improved representation of stratocumulus clouds is required in both GCMs and climate models if changes in cloud properties induced by global warming are to be studied.

The observations of cloud susceptibility presented have shown a wide range of values indicating the inherent variability of low-level stratocumulus clouds. The flights made during ASTEX have clearly shown that continental air masses can have a long sea track (~ 1600 km) and still maintain a strong influence over the cloud microphysics.

The comparison of four cloud sheets measured during ASTEX has clearly demonstrated the nonlinear effects of increasing cloud droplet concentration on the cloud microphysical and radiative properties. In particular, it has been calculated that clean maritime clouds are very sensitive to increases in droplet concentration and that the rate of change of r_e and cloud albedo is, in these conditions, very large in the initial stages. If there is an increase in available cloud condensation nuclei in maritime air masses then, liquid water remaining constant, the increase in cloud albedo will be larger than for a similar increase in CCN in an already polluted air mass. Charlson et al. (1987) have suggested that dimethylsulphide (DMS) production may increase in a warmer climate. This could have significant impact on global climate as this will affect the highly susceptible maritime stratocumulus. The changes in r_e observed resulting from increasing droplet concentrations very quickly reach the magnitude required to offset the top-of-atmosphere radiative forcing induced by a doubling

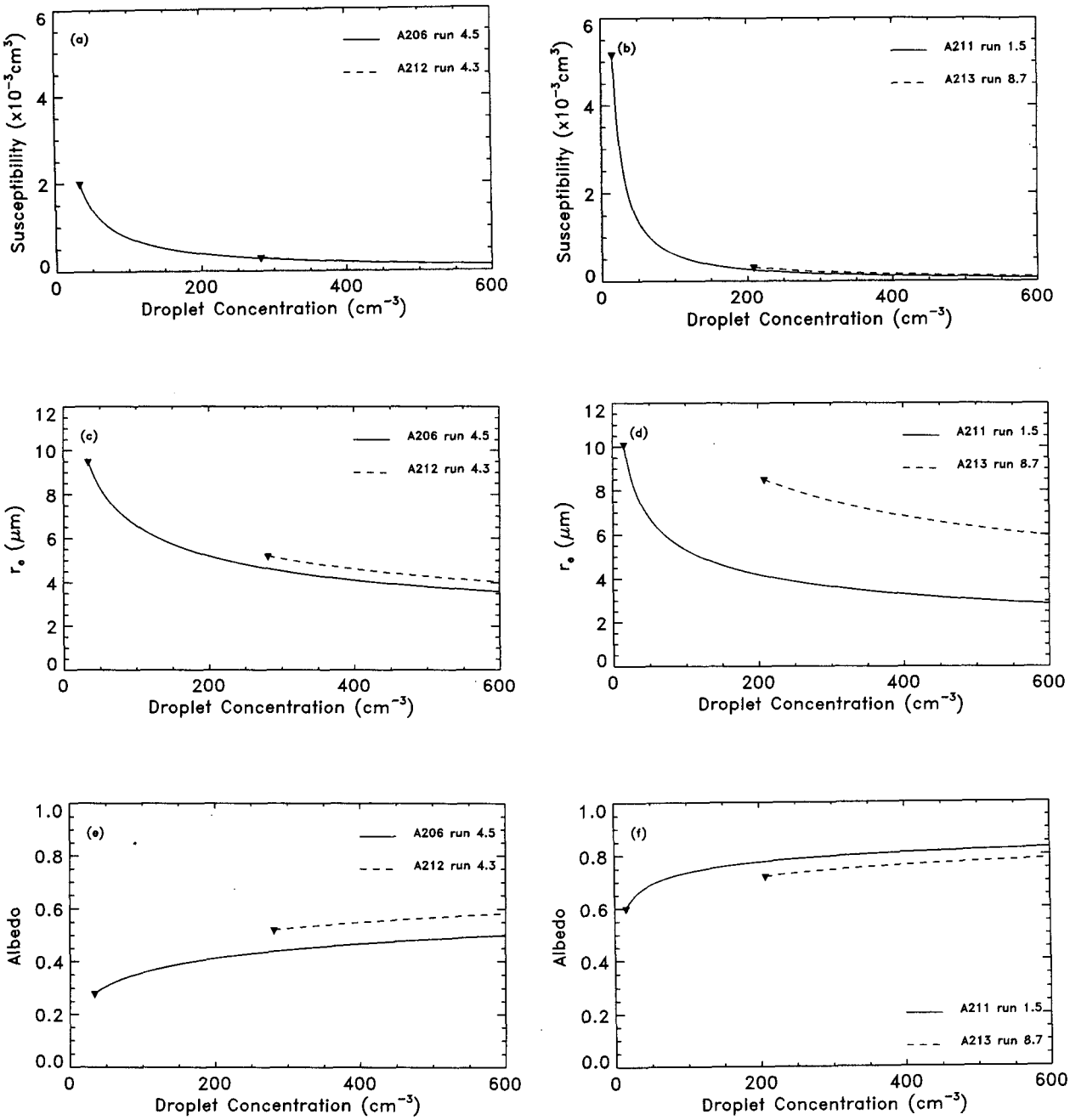

FIG. 6. Simulations of changes in cloud susceptibility (a) and (b), effective radius (c) and (d), and albedo (e) and (f), with increasing droplet concentration.

of carbon dioxide concentrations as calculated by Slingo (1990).

The rapid decrease in susceptibility as droplet concentration is increased suggests that ship track phenomena will be observed visually only where the background cloud is maritime; that is, the initial droplet concentrations are $\leqslant 100$ cm^{-3}. This would suggest that we would not expect to observe similar plumes of increased albedo from sources on the continent such as power stations injecting aerosol into a continental air mass.

It should be noted that all the simulations carried out here have assumed that the increased droplet concentration has arisen from the injection of new aerosol of the same type as the initial aerosol. In reality, this may not be the case, and aerosol of a different chemical

constituency may be injected into the air. This would result in a different droplet spectrum, and the relationship between the mean volume radius and the effective radius would be changed.

The measurements of cloud liquid water content, effective radius, and droplet concentration have been made with an FSSP. This was done in order to assure self-consistency. There is considerable debate as to the accuracy of different instruments in their measurement of cloud microphysical properties, in particular liquid water content. The measurements of susceptibility may therefore have been different if another instrument had been used to measure liquid water content; however, the dominant term in the equation for cloud susceptibility is r_e^3.

It is difficult to generalize as to the global distribution of cloud susceptibility given the limited dataset available here. However, the observations show that the susceptibility in the South Atlantic is generally larger than that observed during FIRE off the west coast of California, a result opposite to that given in PT92 where the reverse was observed. This difference may be due to the use of a fixed liquid water of 0.3 g m^{-3} in the work of PT92.

The data presented have covered only four regions of the earth and no stratocumulus over land has been studied. Further observations of the relatively clean (maritime) clouds in the Southern Hemisphere and detailed measurement of the climatically important arctic stratus would further enhance our understanding of the sensitivity of low-level water clouds to changes in boundary-layer aerosol.

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