Satellite Remote Sensing of the Liquid Water Sensitivity in Water Clouds

Qingyuan Han[*], Willi im B. Rossow[+], Jane Zeng, and Ronald Welch[+]

[*] University of Alabama in Huntsville

Huntsville, AL

[+]NASA/Goddard Institute for Space Studies

New York, NY

Abstract

In estimation of the aerosol indirect effect, cloud liquid water path is considered either constant (Twomey effect) or increasing with enhanced droplet number concentrations (drizzle-suppression effect, or Albrecht effect) if cloud microphysics is the prevailing mechanism during the aerosol-cloud interactions. On the other hand, if cloud thermodynamics and dynamics are considered, the cloud liquid water path may be decreased with increasing droplet number concentration, which is predicted by model calculations and observed in ship-track and urban influence studies. This study is to examine the different responses of cloud liquid water path to changes of cloud droplet number concentration. Satellite data (January, April, July and October 1987) are used to retrieve the cloud liquid water sensitivity, defined as the changes of liquid water path versus changes of column droplet number concentrations.

The results of a global survey reveal that 1) at least one third of the cases the cloud liquid water sensitivity is negative, the regional and seasonal variations of the negative liquid water sensitivity are consistent with other observations; 2) cloud droplet sizes are always inversely proportional to column droplet number concentrations. Our results suggest that an increase of cloud droplet number concentration leads to reduced cloud droplet size and enhanced evaporation, which weakens the coupling between water clouds and boundary layer in warm zones, decreases water supply from surface and desiccates cloud liquid water. Our results also suggest that the current evaluations of negative aerosol indirect forcingby GCMs, which are based on Twomey effect or Albrecht effect, may be overestimated.

1. Introduction

Aerosol radiative forcings, both direct and indirect, are among the main sources of uncertainties in climate change studies. Between them, the aerosol indirect forcing, which is related to the cloud radiative property change through cloud-aerosol interactions, shows the largest uncertainties among all known forcing mechanisms (IPCC, 1996). The importance of the aerosol indirect effect is further demonstrated by the suggestion that the indirect effect is the most possible candidate for the damped diurnal temperature cycle (Hansen et al., 1997).

Significant progresses have been made in recent years to evaluate the aerosol indirect effect using prognostic equations for predicting liquid water content, cloud droplet number concentration in global climate models (e.g., Del Genio et al., 1996; Lohmann et al., 1999; Rotstayn, 1999; Ghan et al., 2000; Menon et al., 2000). These physically based GCMs are more reliable in predicting changes in climate because they are not tuned to parameterizations that are only valid under specific conditions. However, the results of these models are quite different because cloud droplet number concentrations and cloud liquid water are calculated differently [e.g., Lohmann et al., 1999; Jones et al., 1999; Rotstayn 1997; 1999; Ghan et al., 1997]. To reduce the differences in model results, and thus the uncertainties in estimations of the aerosol indirect effect, satellite observations of cloud and aerosol properties and their relationships are crucially needed.

During the first phase of GACP (Global Aerosol Climatology Project), new variables and their relationships are retrieved from satellite observations that include near-global surveys of relationship between cloud albedo and effective radius (Han et al., 1998a), cloud column number concentration (Han et al., 1998b), and cloud column susceptibility (Han et al., 2000). Some of these results have been used for comparisons with model predictions. For example, in the study reported by Han et al. (1998a), results of a near-global survey reveal that cloud albedo and droplet radius are positively correlated for most thin

clouds (τ<15) and negatively correlated for most thick clouds (τ>15). Such a relationship was used for

comparison by several GCM groups and general agreements were found (e.g., Lohmann et al., 1999;

Ghan et al., 2000). Nevertheless, large uncertainties in estimations of the aerosol indirect effects still

exist. For example, it is found that the estimated aerosol indirect effect (-1.7 W/m^2) from MIRAGE

model (Ghan et al., 2000) is much larger than that (-0.4 W/m^2) estimated by Lohmann et al. (1999) using

ECHAM model although cloud liquid water change due to aerosol indirect effect is smaller in the

MIRAGE than in the ECHAM model. This indicates that detailed quantitative comparisons including

relationships between different parameters and their variations are needed.

Most GCMs include two important variables: cloud droplet number concentration and cloud liquid

water content (e.g., Del Genio et al., 1996; Lohmann et al., 1999; Ghan et al., 1997; Rotstayn, 1999;

Menon et al., 2000). Increases in cloud droplet number concentration are a direct indication of the

aerosol-cloud interaction and it is considered the driving force of the indirect effect. Observation during

the past several decades shows that the basic indication of aerosol-cloud interaction is the enhanced cloud

droplet number concentration, N (e.g., Warner and Twomey, 1967; Fitzgerald and Spyers-Duran, 1973;

Eagan et al., 1974; Alkezweeny et al., 1993; Hudson and Svensson, 1995).

The cloud liquid water content is the basis for calculating cloud droplet sizes, cloud radiative

properties (optical thickness) and precipitation. Therefore, model estimation of the aerosol indirect effect

includes two branches: one is to model the relation between cloud droplet number concentration and

aerosol concentrations (e.g., Hudson et al., 2000 and references therein); the other is to predict the cloud

liquid water content with changing cloud droplet number concentrations (e.g., Durkee et al., 2000 and

references therein). Great efforts have been made in the first branch that includes empirical relations

between aerosol concentrations and cloud droplet number concentrations (e.g., Jones et al., 1994, 1999;

Boucher and Lohmann, 1995, Jones and Slingo, 1996; Rotstayn, 1999), or physically based aerosol

activation process (e.g., Ghan et al., 1997; Lohmann et al., 1999). The intention of this study is to investigate the second branch, i.e., examine the responses of cloud liquid water to changes of cloud droplet number concentrations.

From consideration of cloud microphysics, it is hypothesized that increased droplet number concentration leads to smaller droplet sizes that makes precipitation difficult (Albrecht et al., 1989), which is supported by observations showing that liquid water path increases due to suppression of drizzles in ship track studies (Radke et al., 1989; Ferek et al., 2000) and in smoke plumes (Rosenfeld et al., 1999). However, from considerations of cloud dynamics and thermodynamics, model studies show that cloud base cooling can lead to reduced boundary-layer mixing, which restricts the supply of water vapor and results in reduction of cloud liquid water (e.g., Lilly, 1968; Bougeault, 1985; Turton and Nicholls, 1987). Cloud base cooling can be caused by aerosol-cloud interactions because when more CCNs are activated into cloud droplets, the total droplet surface area, and thus evaporation, increases due to a greater droplet concentration and smaller average size of the droplets. The increased evaporation at cloud base leads to a greater decoupling between the cloud and the subcloud layers that causes a thinning of cloud layer (Ackerman et al., 1995). The decreased cloud liquid water content with increased droplet number concentration is observed in ship track studies (e.g., Platnick et al., 2000; Ackerman et al., 2000a) and in urban influences on cloud properties (Fitzgerald and Spyers-Duran, 1973).

In current GCMs, the response of cloud liquid water to changes in droplet number concentration is through the influence of droplet number on the autoconversion of cloud water to rain, i.e., larger droplet concentration will either decrease the autoconversion rate of cloud droplets (e.g., Beheng, 1994; Lohmann and Feichter, 1997) or increase the critical threshold for autoconversion to start (e.g., Rotstayn, 1999). These mechanisms lead to a general increase in cloud liquid water content with increasing droplet number (e.g., Ghan et al., 2000). Although evaporation and its influence on droplet sizes are considered in a few

GCMs (e.g., Lohmann et al., 1999), its influence on thermodynamics and the feedback on cloud liquid water is difficult to parameterize partially due to the coarse vertical resolution in GCMs (Del Genio, personal communication).

The questions are: what is the general behavior of cloud liquid water in response to increased droplet number concentrations? what are its temporal and spatial variations? If the majority of cloud liquid water increases with increased droplet number, then the consideration of cloud microphysics is good enough and we are confident about the responses of cloud liquid water (and thus cloud optical properties) to aerosol-cloud interactions. If this is not the case, then more efforts in the models have to be made to include the difficult but important effect of cloud dynamics and thermodynamics for an accurate estimation of the aerosol indirect effects.

This study is to answer these questions through satellite observations. In section two, we define the cloud liquid water sensitivity that is appropriate for comparisons of results from model calculations and satellite retrievals. Section three presents the satellite data used in this study. Section four shows results and section five is discussion and conclusions.

2. Cloud liquid water sensitivity

We start with a definition that makes the comparison between results of model prediction and satellite observation appropriate. Since observations show that changes in cloud geometrical thickness during aerosol-cloud interactions cannot be ignored (e.g., Hobbs et al., 1970, Ackerman et al., 2000a) consistent with model predictions (Pincus and Baker, 1994; Ackerman et al., 1993), column-integrated values of cloud droplet number concentration, N, and liquid water content, LWC, is more appropriate in describing this relationship to avoid assumptions of constant geometrical thickness of clouds. Satellite

remote sensing has provided these column-integrated parameters, i.e., column droplet number

concentration (Han et al., 1998b),

$$N_c = N \cdot h \tag{1}$$

and liquid water path (e.g., Greenwood et al., 1995, Han et al., 1994)

$$LWP = lwc \cdot h \tag{2}$$

where h is the cloud geometrical thickness.

We define the cloud water sensitivity as

$$\delta = \frac{\Delta LWP}{\Delta N_c} \tag{3}$$

Note that this definition is similar to the definition of "cloud column susceptibility" (Han et al., 2000), in

which Da (changes in cloud spherical albedo) is replaced by DLWP (changes in cloud liquid water path).

The reason that we do not use the term "susceptibility" here is that it means "apt to" or " the potential to

be affected by" and therefore is determined by properties of individual cloud as first proposed by Twomey

(1991). However, aerosol-cloud interactions are not only determined by properties of clouds and

aerosols, they are also determined by the conditions of environment such as thickness of boundary layer

(e.g., Durkee et al., 2000). This is the reason that ship tracks are not found in many clouds with high

susceptibilities (e.g., Platnick and Twomey, 1994; Coakley et al., 2000). In our approach, the cloud water

sensitivity, δ, is derived using the least-square linear regression to determine the slope of ΔLWP and ΔN_c

for all water clouds within a 2.5°x2.5° grid box during one month period. Therefore, the derived value

describes "what actually happened", which is determined not only by clouds, but also by the condition of

environments. In this sense, the terminology "cloud column susceptibility" used in Han et al. (2000) is

not accurate. It should be modified to "cloud albedo sensitivity" when it was derived based on monthly

data from a grid box.

Liquid water sensitivity represents the absolute change of liquid water path for changes in column droplet number concentration, which is affected by the total water availability: clouds in a moist environment (e.g., maritime) tends to have larger liquid water sensitivity than those in a dry environment (e.g., continent). To this end, the relative liquid water sensitivity may describe the effect of aerosol-cloud interaction for different environments, which is defined as

$$b = \frac{\Delta LWP / LWP}{\Delta N_c / N_c} = \frac{\Delta \ln(LWP)}{\Delta \ln(N_c)} \tag{4}$$

If the relation between effective radius and volume average radius is used, i.e.,

$$r_e^3(1-b)(1-2b) = \bar{r}^3 \tag{5}$$

where b is the effective variance for gamma distribution (Han et al., 1998b), then we have

$$\Delta \ln(LWP) = 3\Delta \ln(r_e) + \Delta \ln(N_c) \tag{6}$$

It is readily seen that the relative liquid water sensitivity, β, and the power γ in the relation

$$r_e \propto N_c^\gamma \tag{7}$$

are closely related by

$$\beta = 3g + 1 \tag{8}$$

For the case of constant liquid water path, $\beta=0$ and $\gamma= -(1/3)$.

3. Method and Data

The data used are near global datasets of cloud properties including cloud optical thickness, effective radius, liquid water path and column number concentrations for January, April, July and October, 1987 developed using ISCCP data (Han et al., 1994, Han et al., 1998b). The original ISCCP analysis separates cloudy and clear image pixels (area about 4 x 1 km^2 sampled to a spacing of about 30 km) and retrieves cloud optical thickness and top temperature (T_c) from radiances measured by AVHRR at wavelengths of 0.54 - 0.80 μm (Channel 1) and 10.0 - 11.6 μm (Channel 4), assuming r_e = 10 μm. The

analysis uses the NOAA TIROS Operational Vertical Sounder (TOVS) products to specify atmospheric temperature, humidity and ozone abundance and also retrieves the surface temperature (T_s). The ISCCP analysis is extended by retrieving r_e from AVHRR radiances at wavelengths of 3.44 - 4.04 μm (Channel 3) and revising the values of τ to be consistent for clouds with $T_c \geq 273$ K (Han *et al*. 1994, 1995). Only liquid water clouds are considered in this study because 90% of the tropospheric aerosol are distributed below 3 km altitude (Griggs 1983). Moreover, aerosol effects on ice clouds may be different than on liquid water clouds. The radiances are modeled as functions of illumination/viewing geometry by including the effects of Lambertian reflection/emission from the surface (the ocean reflectance is anisotropic, see Rossow *et al*. 1989), absorption/emission by H_2O, CO_2, O_3, O_2, N_2O, CH_4, and N_2 with the correlated k-distribution method (Lacis and Oinas 1991), Rayleigh scattering by the atmosphere and Mie scattering/absorption by horizontally homogeneous cloud layers using a 12-Gauss point doubling/adding method. The droplet size distribution is assumed to the gamma-distribution. Error sources are discussed and validation studies are reported in Han *et al*. (1994, 1995). Note that the satellite-measured radiation is only sensitive to the droplet sizes in the topmost part of the clouds; therefore, the values of LWP obtained by this analysis may be biased if r_e at cloud top is systematically different from the vertically averaged value (Nakajima et al. 1991). For non-precipitating clouds (LWP ≤ 150 g/m^2), the results of this method agree well with ground based microwave radiometer measurements (Han *et al*. 1995). Lin and Rossow (1994, 1996) show excellent agreement of microwave (from SSM/I) determinations of LWP over the global ocean with those obtained from the ISCCP results, assuming 10 μm droplets, and Greenwald *et al*. (1997) compare microwave retrievals of LWP from SSM/I and from GOES-8 over the Pacific Ocean.

All of the individual pixel values are collected for each 2.5° x 2.5° map grid cell for each month, representing both spatial variations at scales ~ 10 - 100 km and daily variations over each month. Only clouds with cloud top temperature warmer than 273 K were used in this study. To reduce the possible effects

of cloud fractional cloud cover on cloud droplet radius (Han et al., 1995), only pixels with cloud optical thickness larger than unity were included. Since thinner clouds are more apt to be influenced by the aerosol indirect effect, only results of clouds with $1<\tau\leq15$ are shown. Typically, about 100 samples per map grid cell per month are available; results are not reported if there are fewer than 10 samples.

The liquid water sensitivity, δ, is derived by least square linear regression between LWP and N_c values. The power γ in the power law relation of r_e and N_c, which is related to the relative liquid water sensitivity, β, by Eq. (8), is derived by least square linear regression between $\ln(r_e)$ and $\ln(N_c)$.

4. Results

4.1 Liquid Water Sensitivity

Figure 1 is a near-global survey of the liquid water sensitivity in water clouds for January, April, July and October 1987. Considering the whole range and appropriate details in spatial variations, the unit used is [g m^{-2}/3x10^6 cm^{-2}]. For a typical 300 m thickness of cloud, 1 [g m^{-2}/3x10^6 cm^{-2}] corresponds to an increase of cloud liquid water path by 1g m^{-2} for a change of cloud droplet number concentration by 100 cm^{-3}. Green and blue colors represent negative liquid water sensitivities and yellow and red colors stand for positive liquid water sensitivities.

The most obvious feature is that negative liquid water sensitivities are by no means rare cases -- they are everywhere. For continental clouds, most clouds show neutral or slightly negative liquid water sensitivities. For maritime clouds, there are areas with both large negative and positive liquid water sensitivities with a strong seasonal dependence, i.e., negative liquid water sensitivity is common in the summer hemisphere. If the negative liquid water sensitivity is caused by decoupling of boundary, then one has to explain why this decoupling happens more often in warm areas than cold areas. This is found by observations of four years of surface remote sensing data from the ARM (Atmospheric Radiation

Measurement) Cloud and Radiation Testbed site (Del Genio and Wolf, 2000). In an effort of explaining

the negative dependency of cloud optical thickness on surface temperature, they found that the types of

boundary layers are different for cold and warm surface temperatures: with stratified and convective

boundary layer associated with cold temperature and mixed or decoupled boundary layer with warm

temperature. Detailed analyses of boundary layer condition show that while decoupling of boundary layer

is responsible for decreasing of cloud liquid water and thinning of cloud layer, it is not related to surface

temperature (Del Genio and Wolf, 2000). In other words, warmer surface temperature alone is not the

cause of the decoupling of boundary layer and the decreasing of cloud liquid water path; other factor must

play a role in this process. The coincidence of negative liquid water sensitivity in warmer seasons shown

in the Figure 1 suggests a possible role of cloud microphysics, which is predicted by model studies (Lilly,

1968; Bougeault, 1985; Turton and Nicholls, 1987). That is, increased droplet number concentration leads

to decreases of droplet size (which is a global phenomena as will be shown later), hence to enhanced

cloud base cooling due to evaporation and to reduced water supply from surface due to a weakened

coupling between clouds and boundary layer. Due to the differences in boundary layer types, this

decoupling effect is more significant for warm zones.

Figure 2 is the histogram of the percentage of liquid water sensitivity for each category with its

values listed in the Table 1. It shows that, as an annual average, about one third of the times cloud liquid

water sensitivities are negative, while about a quarter of the times they are positive although the

percentage varies with season.

4.2 Relative liquid water sensitivity

Figure 3 is a near-global survey of the relative liquid water sensitivity, β, which is related to the γ

value in the relation of $r_e \propto N_c^\gamma$, through Eq. (8). It is apparent that, unlike the near neutral absolute liquid

water sensitivities, the relative liquid water sensitivities are mostly negative over land, which means that

the relative change in liquid water path is notably affected by the relative change in column droplet

number concentration over land although the absolute change is insignificant.

Figure 4 is the histogram of the percentage of relative liquid water sensitivity for each category

with its values listed in the Table 2. It shows that, as an annual average, about 40% of the times the

relative liquid water sensitivities are negative, while about 28% of the times they are positive although the

percentage varies with season.

Figures 3 and 4 reveal that the effective droplet radius and column droplet number concentration

are always negatively correlated, suggesting that enhanced droplet number concentration always leads to

decreased droplet size, although to different degrees. Many field observations find that the $-(1/3)$ power

law is valid for relations between droplet radius and *volume* number concentrations and it was also

noticed that variations in cloud thickness cannot be neglected (e.g., Durkee et al., 2000; Ackerman et al.,

2000a). Our results show that about one third of the cases the minus one third power law ($-0.37 < \gamma < -0.33$)

is valid even for droplet radius and *column* number concentrations (which means cloud thickness

variations included).

5. Discussions and Conclusions

The response of cloud liquid water path to column droplet number concentration is the basis for

estimating the aerosol indirect effect. This response has been parameterized in GCMs as either constant

(Twomey Effect) or increased with increasing droplet number concentrations due to suppression of

drizzling (Albrecht Effect). Although model studies and field observations suggest that there may be

another behavior of response, i.e., cloud liquid water content may be decreased with increasing droplet

number concentrations, the relative importance has kept unknown. This study examines this response by

retrieving the liquid water sensitivity on a near-global scale using satellite data and finds that more than

one third of the cases, the liquid water sensitivities are negative, i.e., cloud liquid water path decreases with increasing column number concentrations. Another finding of this study is that cloud droplet sizes always decrease with enhanced column droplet number concentrations.

Seasonal variations of the liquid water sensitivity show that most negative values are in the "warm zone" or summer hemisphere. This can be explained by the findings using four years of ground observations that the types of boundary layer are different in warm zone from that in cold zone: well-mixed or decoupled convective boundary for warm zone and well-stratified boundary layer for cold zone (Del Genio and Wolf, 2000). They also found that the decoupled boundary layer is strongly associated with decreased liquid water path but the decoupling is not dependent on surface temperature, which suggests that while the boundary layer is apt to be decoupled in warm zone, surface temperature does not play decisive role in the decoupling. Combined with their findings, our results suggest that the increased droplet number concentration leads to decreased droplet size and enhanced evaporation at cloud base, which cause the boundary layer decoupled in warm zones, which is consistent with simulations of model studies (Lilly, 1968; Bougeault, 1985; Turton and Nicholls, 1987).

We note that the pattern of retrieved liquid water sensitivity may include contributions from clouds formed in different air masses, which is especially true for areas close to coastlines. For example, maritime clouds with small droplet number concentration and continental clouds with large droplet number concentration are often found in certain coast regions (e.g., Minnis et al., 1992; Twohy et al., 1995). Using observational data during the Indian Ocean Experiment (INDOX), model simulations show that cloud liquid water path decreases with increasing droplet number concentration if different air masses are considered (Ackerman et al., 2000b). Nevertheless, the negative liquid water sensitivity found in vast areas including remote ocean area and relatively clean southern hemisphere suggests that enhanced

droplet number concentration plays important role in inducing the decoupling of boundary layer, reducing water vapor supply from surface and desiccating cloud liquid water.

We also note that the results of this study should not be regarded as "before and after" aerosol-cloud interactions for individual clouds, instead, the results are statistical in nature. This should not be a problem when used for comparison with GCM results because cloud properties predicted by GCMs are also statistical in nature – they are not specific predictions for individual clouds in a weather system. Therefore, GCMs should be able to simulate the effects of advections, including aerosol and water vapor, when the liquid water sensitivity is contributed by different air masses in certain regions.

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Figure Captions

Figure 1. Liquid water sensitivity of water clouds for January, April, July and October 1987. The unit is in [g m^{-2}/3x10^6 cm^{-2}]. For a typical 300 m thickness of cloud, 1 [g m^{-2}/3x10^6 cm^{-2}] corresponds to an increase of cloud liquid water path by 1g m^{-2} for a change of cloud droplet number concentration by 100 cm^{-3}.

Figure 2. Histogram of the liquid water sensitivity for January, April, July and October 1987.

Figure 3. Relative liquid water sensitivity (β) and power γ in the relation $r_e \sim N_c^\gamma$ of water clouds for January, April, July and October 1987.

Figure 4. Histogram of the power γ in the relation $r_e \sim N_c^\gamma$ of water clouds for January, April, July and October 1987.

Table I. Percentage of cloud liquid water sensitivity for different ranges

	$\delta=\Delta LWP/\Delta N_c<0$				$\delta\approx0$	$\delta=\Delta LWP/\Delta N_c>0$					
	$\delta<-35$	$-35<\delta$ <-25	$-25<\delta$ <-15	$-15<\delta$ <-5	$-5<\delta$ <5	$5<\delta$ <15	$15<\delta$ <25	$25<\delta$ <35	$35<\delta$ <45	$45<\delta$ <95	$95<\delta$
Jan	1.07	1.98	8.93	27.3	40.9	9.69	3.76	2.12	1.49	1.24	1.49
	39.3				**40.9**	**19.8**					
Apr	1.89	1.84	6.15	20.8	38.9	11.4	5.97	3.77	2.99	2.52	3.77
	30.7				**38.9**	**30.4**					
Jul	1.31	2.85	9.55	22.6	37.1	8.72	5.04	3.31	2.52	2.60	4.46
	36.3				**37.1**	**26.6**					
Oct	0.73	1.28	5.11	19.6	44.9	11.8	7.77	4.26	2.23	1.28	1.03
	26.7				**44.9**	**28.4**					

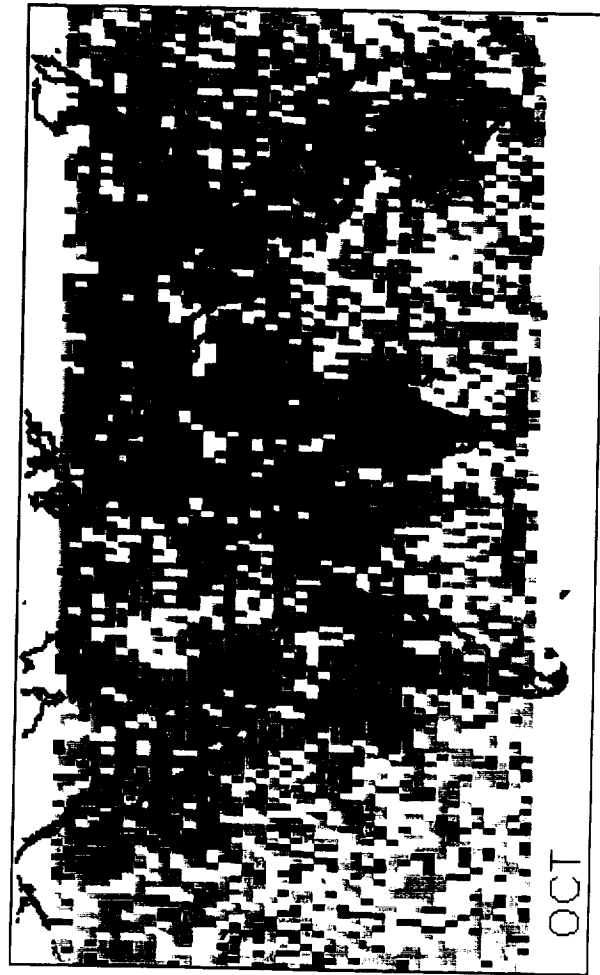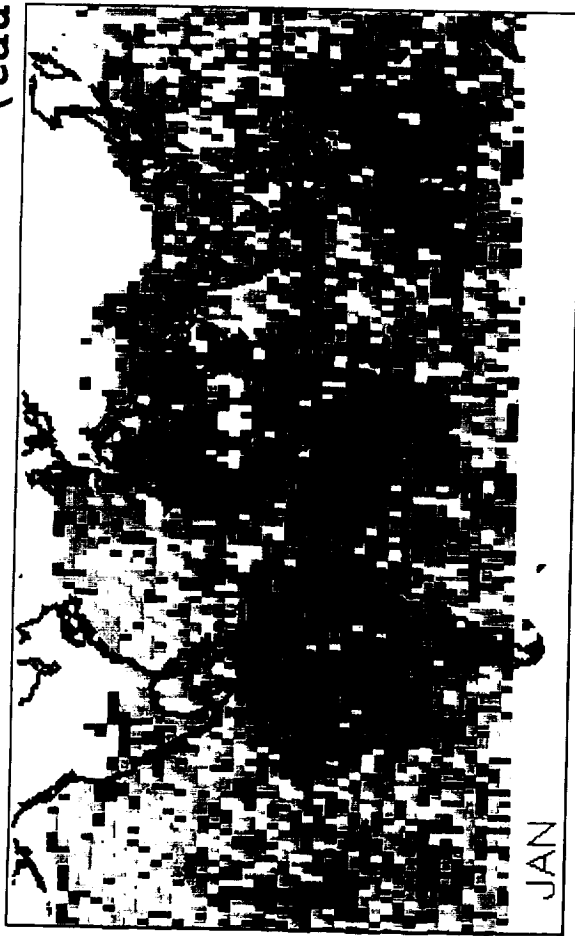Table II, Relative liquid water sensitivity, β, and γ values in relation $r_e \sim N_c^\gamma$

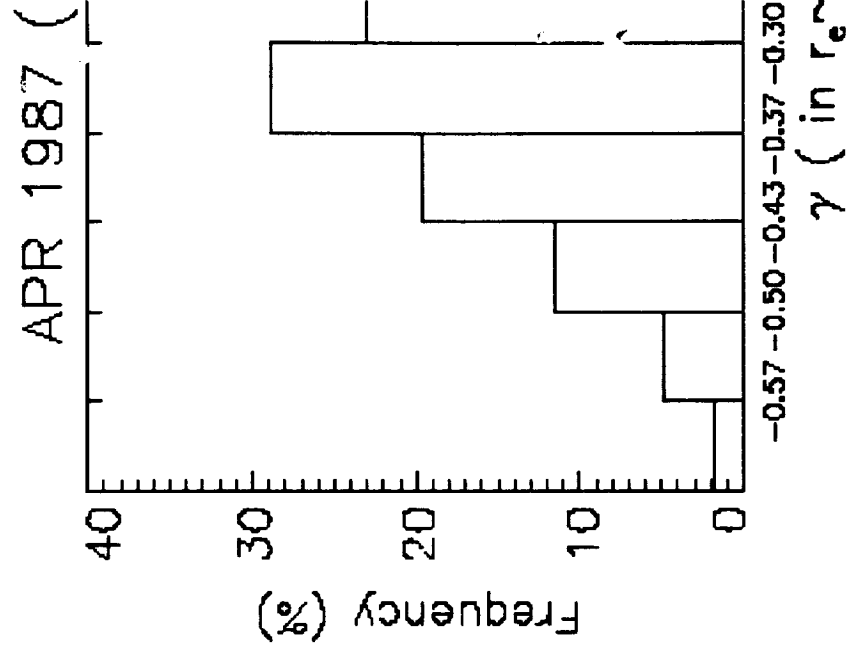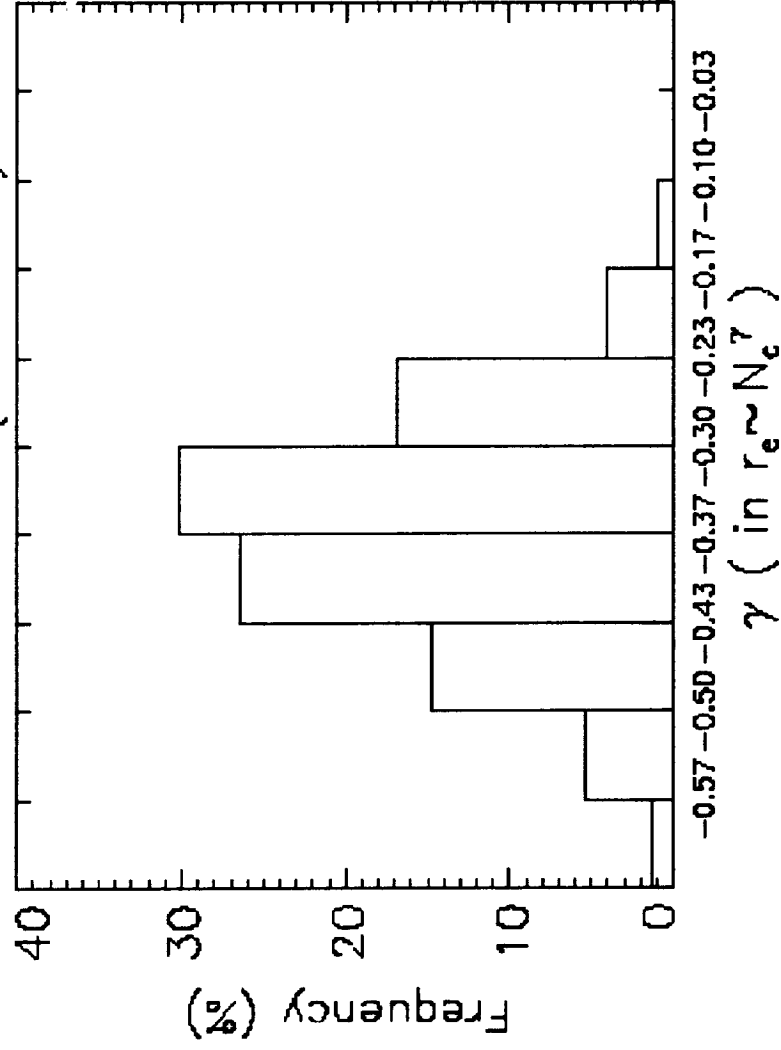|	b=Dln(LWP)/Dln(N$_c$) <0				b≈0	b=Dln(LWP)/Dln(N$_c$) >0			
	β<-70%	-70%<β <-50%	-50%<β <-30%	-30%<β <-10%	-10%<β<10%	10%<β <30%	30%<β <50%	50%<β <70%	β>70%
	γ<-0.57	-0.57<γ< -0.50	-0.50<γ< -0.43	-0.43<γ< -0.37	-0.37<γ< -0.30	-0.30<γ< -0.23	-0.23<γ< -0.17	-0.17<γ< -0.10	-0.10<γ
Jan	0.95	4.24	14.3	26.7	31.8	17.2	3.76	0.97	0.08
	46.2				**31.8**	**22.0**			
Apr	1.36	4.15	11.2	20.0	30.0	23.8	7.88	1.36	0.21
	36.7				**30.0**	**33.3**			
Jul	1.84	5.77	12.4	23.2	27.8	20.4	7.23	1.39	0.08
	43.2				**27.8**	**29.1**			
Oct	0.37	1.49	7.45	24.4	38.3	23.1	4.40	0.46	0.09
	33.7				**38.3**	**28.0**			

Linear Power Parameter (LWP - Nc) (TAU: 1-15)

JAN

APR

JUL

OCT

-35 -25 -15 -5   5   15 25 35 45 95

Linear Power Parameter (ln(Re) - ln (Nc))
(tau: 1-15)

JAN

APR

JUL

OCT

γ : -0.57 -0.50 -0.43 -0.37 -0.30 -0.23 -0.17 -0.10 -0.03
β : -0.7 -0.5 -0.3 -0.1 0.1 0.3 0.5 0.7 0.9

APR 1987 (

Frequency (%)

40 30 20 10 0

-0.57 -0.50 -0.43 -0.37 -0.30

γ (in r_e

OCT 1987 (

Frequency (%)

40 30 20

JAN 1987 ($1 \leq \tau < 15$)

Frequency (%)

40 30 20 10 0

-0.57 -0.50 -0.43 -0.37 -0.30 -0.23 -0.17 -0.10 -0.03

γ (in $r_e \sim N_c^\gamma$)

JUL 1987 ($1 \leq \tau < 15$)

Frequency (%)

40 30 20

Abstract

In estimation of the aerosol indirect effect, cloud liquid water path is considered either constant (Twomey effect) or increasing with enhanced droplet number concentrations (drizzle-suppression effect, or Albrecht effect) if cloud microphysics is the prevailing mechanism during the aerosol-cloud interactions. On the other hand, if cloud thermodynamics and dynamics are considered, the cloud liquid water path may be decreased with increasing droplet number concentration, which is predicted by model calculations and observed in ship-track and urban influence studies. This study is to examine the different responses of cloud liquid water path to changes of cloud droplet number concentration. Satellite data (January, April, July and October 1987) are used to retrieve the cloud liquid water sensitivity, defined as the changes of liquid water path versus changes of column droplet number concentrations.

The results of a global survey reveal that 1) at least one third of the cases the cloud liquid water sensitivity is negative, the regional and seasonal variations of the negative liquid water sensitivity are consistent with other observations; 2) cloud droplet sizes are always inversely proportional to column droplet number concentrations. Our results suggest that an increase of cloud droplet number concentration leads to reduced cloud droplet size and enhanced evaporation, which weakens the coupling between water clouds and boundary layer in warm zones, decreases water supply from surface and desiccates cloud liquid water. Our results also suggest that the current evaluations of negative aerosol indirect forcing by GCMs, which are based on Twomey effect or Albrecht effect, may be overestimated.