NOTES

Aerosol Size Spectra in a Convective Marine Layer with Stratus: Results of Airborne Measurements near San Nicolas Island, California

V. RAY NOONKESTER

Naval Ocean Systems Center, San Diego, CA 92152

9 February 1981 and 15 May 1981

ABSTRACT

Airborne measurements of the aerosol size spectra $n(r)$ (r is radius) were made in a vertical plane extending northeastward 18 km from San Nicolas Island, California. Thin, patchy, stratus clouds were present in a deepening convective marine layer capped by a strong temperature inversion based at ~147 m. These data show that the aerosol spectra shape changed systematically as a function of the liquid water content w, calculated from $n(r)$, as w increased with elevation into the stratus layer. Clouds formed when w approached 0.05 g m^{-3} and a mode formed in $n(r)$ near the region $3 < r < 4$ μm. The horizontal variability of w was large, particularly in the middle of the convective layer.

---

1. Introduction

Over-ocean measurements of the aerosol size spectra $n(r)$ (r is radius) are sparse and are needed for research and electro-optical systems development concerning atmospheric optical properties. During an experiment entitled Cooperative Experiment in West Coast Oceanography and Meteorology—1978 (CEWCOM-78) the Naval Ocean Systems Center measured $n(r)$, air temperature, dew point and elevation z over the ocean near San Nicolas Island (SNI), California using a twin engine Piper Navajo. On 9 May 1978, horizontal runs were made at five elevations in a vertical plane extending 18 km northeastward from the northwest tip of SNI when moderate convective activity had created thin, patchy, stratus clouds in marine air. This paper discusses the systematic change in the spectral shape of $n(r)$ as a function of the liquid water content w and the change of w with elevation from the surface through the cloud layer and up to the top of the mixed layer at elevation z_i. The systematic changes in $n(r)$ agreed remarkably well with several models for moderate convection in marine air.

2. Sensors and data

The particle size distributions $n(r)$ were measured by a PMS ASSP-100 aerosol drop size spectrometer. The temperature T was measured by an HP (Model 2801A) quartz thermometer probe accurate to ±0.1°C and the dew point temperature was measured by an EG&G (Model 137-C3) instrument accurate to ±0.5°C. A pressure sensor provided elevation measurements accurate to ±9 m.

Dytch and Carrera (1976) describe the spectrometer probe (PMS ASSP-100) which counts particles with radii from 0.23 to 14.7 μm using 15 subrange sizes in four major overlapping size ranges. Two of these probes, operated by NOSC, have been simultaneously operated side-by-side (in both low and high visibilities) and have produced highly similar drop-size spectra. Intermittent size range calibrations with latex spheres and periodic examination of the probes by the manufacturer have provided confidence that the ASSP-100s continuously operate properly. The liquid water content w was determined by integrating over $n(r)$ assuming the aerosols are spherical water drops.

A comparison of $n(r)$ given by several colocated PMS spectrometers operating simultaneously (Jensen, 1980) revealed differences in $n(r)$ by factors of ~10 for all regions of r. Although similar errors are probably present in the $n(r)$ considered here, the general relationships presented between spectral shape, w and z are thought to approximate closely the actual conditions.

A particle-size spectrum was measured every 4 s, and the temperature, dew point and pressure altitude were measured every 5 s. The aircraft flew at ~139 kt (71.6 m s^{-1}) along horizontal runs so that each spectrum $n(r)$ represented a 286 m horizontal distance. A horizontal distance of 358 m was traversed each 5 s between sampling times of the other parameters. The ratio of the average vertical separation (resolution) to the horizontal separation (resolution) in the aerosol measurements was 1:5.

The aircraft flew at near-constant elevations in a vertical plane extending 18 km along a bearing of 50° from true north from the NW tip of SNI. The

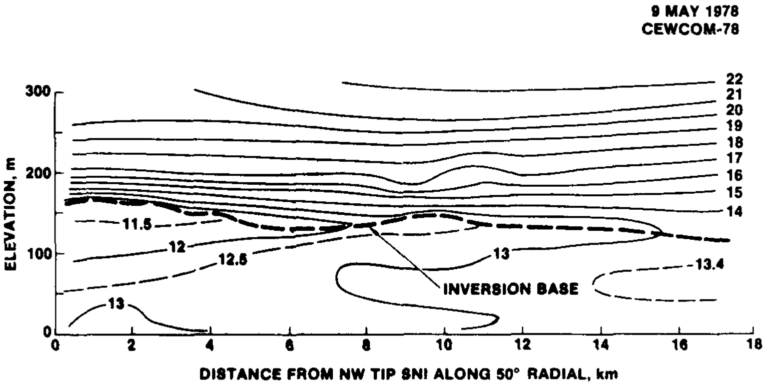

FIG. 1. Isopleths of temperature (°C) in vertical plane extending northeast of San Nicolas Island.

northwest region of SNI was in clouds on 9 May according to a DMSP visual depiction at 0955 PST, about two hours before the aircraft measurements. Because the surface winds at the northwest tip of SNI were 312° at 4 m s^{-1}, the measurement plane was essentially oriented in a cross-wind direction.

A DMSP IR depiction of the Los Angeles Bight at 1147 PST on 11 May revealed that the sea surface temperature decreased northwest of SNI. The air moving toward SNI in the northwest wind was probably being heated by the water on 9 May. The presence of a surface-based superadiabatic layer was revealed by a radiosonde released from SNI just before the airborne measurements. Near-surface atmospheric profile measurements on the northwest tip of SNI (Blanc, 1978) indicated that the surface heat flux varied from 32 to 105 W m^{-2} within two hours after the aircraft measurements. Surface heat fluxes of these magnitudes are expected to create moderate convective activity (Friehe and Schmitt, 1976).

Using determinations of the Monin-Obukhov length L (Blanc, 1978) and the height of the inversion base z_i at SNI a few hours after the aircraft measurements, $-z_iL^{-1} \approx 10$. This value is about an average of the data found by Fitzjarrald (1978) for 17 over-ocean free-convective periods. Conditions on 9 May apparently were representative of ocean convection.

According to measurements of radon (^{222}Rn) (Larsen and Bressan, 1980) and Aitken particle count (Niziol *et al.*, 1980) and air trajectory analysis (Rosenthal *et al.*, 1980), the air at SNI had a marine origin on 9 May. Thus the data presented here appears to represent moderate convection in marine air.

The vertical cross section of T along the plane of measurement is shown in Fig. 1. Cool air was present below the inversion base at elevation z_i. A similar analysis of the relative humidity RH revealed a moist layer below z_i with a maximum RH near 100% immediately below z_i out to $\sim$11 km from SNI and a rapid decrease in RH above z_i (measurement error in RH was several percent). The convective layer depth z_i decreased from 160 m near SNI to 120 m 16 km from SNI. The RH and $n(r)$ data and visual observations aboard the aircraft indicated the presence of clouds out to $\sim$11 km from SNI. Consequently, the subsequent analysis is restricted to the vertical plane out to 11 km from SNI, where z_i was apparently above the convective condensation level z_c (cloud base near 120 m).

Construction of representative vertical profiles of w required z to be normalized by z_i because z_i decreased 30 m over the 11 km path from SNI. This scaling is commensurate with scaling techniques generally used in mixing-layer theory.

3. Vertical variations of w and $n(r)$

The relative humidity generally increases with elevation up to a level near the convective condensation level (RH $\sim$ 100%) z_c in a well mixed convective layer. Water vapor will condense on aerosols in vertically moving air. This increases the number of large aerosols and hence the liquid water content w will increase with elevation. If the total water vapor uptake (evaporation) is by aerosol growth (decay) in the radius range measured by the spectrometer and the aerosol size changes do not appreciably lag changes in RH, then w is a rough indicator of RH, at least, below the cloud base and about 100 m above the cloud base (Pruppacher and Klett, 1978). Above the convective condensation level z_c (cloud base) w will increase steadily (Warner, 1955) and a mode in $n(r)$ is expected to form in a region near an r of 3–6 μm in the lowest regions of the cloud. Because w is a more direct measure of the water vapor uptake by aerosols than relative humidity, w is used to relate characteristics of $n(r)$ as a function of elevation in the convective layer.

Isopleths of w were constructed for the vertical plane out to 11 km from SNI. Although some subjec-

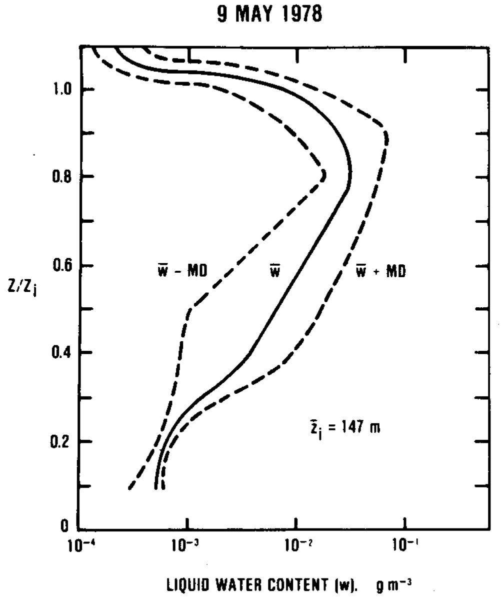

FIG. 2. Vertical profile of the average liquid water content $\bar{w}$ and the mean deviation (MD) above and below $\bar{w}$.

tivity entered the graphical analysis, a reasonable pattern was clearly present. Fig. 2 shows the vertical profile $\bar{w}(z/z_i)$, constructed from the isopleth pattern, where $\bar{w}$ is the average w and where elevation has been normalized to the height of the temperature inversion base z_i. The average inversion base height $\bar{z}_i$ was 147 m. Because $w(z/z_i)$ was not normally distributed at each z/z_i, the mean deviation (MD) of w above and below $\bar{w}$ was used to estimate dispersion. Values of $\bar{w}$ were estimated at $z/z_i > 1$.

If w is assumed to be an indicator of RH, then convective mixing would be expected to produce the profiles presented in Fig. 2 (Deardorff, 1976; Warner, 1955). Strong unorganized mixing would produce minor dispersion at low levels. Adiabatic cooling in vertically moving convective cells would cause RH and w to increase with elevation, and mixing along the edges of the cells would cause the dispersion to increase with elevation as exhibited in the region $0.3 < z/z_i < 0.9$. Above a z/z_i of about 0.9 mixing of the convective moist air with the overlying dry air would produce a rapid decrease of w. The rate of decrease of w above $z/z_i = 0.9$ may have been greater than shown.

Data in Fig 3 show $n(r)$ observed on 9 May and $n(r)$ modeled for stratus cloud conditions. Observed and modeled spectra for cloud conditions by Neiburger and Chien (1960) closely approximate the in-cloud spectra observed near SNI. Deirmendjian's (1964) general cloud model does not except for the maximum n near an r of 4 μm. The spectrum for 1141:48 PST was observed below the cloud (1150:38 spectrum) at an elevation of 30 m and represents "pre-cloud" conditions. This "pre-cloud" spectrum is approximated by as "pre-cloud" spectrum modeled by Neiburger and Chien (1960). Neiburger and Chien assumed salt nuclei in their model.

Table 1 gives significant spectral parameters for some "near-," "in-" and "out-" of cloud spectra. The values of w, the total particle count N (cm^{-3}), the modal radius r_{mo}($r > 1$ μm), and the aerosol density at r_{mo} closely approximate values found by Neiburger and Chien (1960) also given in Table 1. Mordy (1959) computed the spectral evolution of droplets having a salt nuclei size distribution given by Woodcock (1953) for a surface wind speed about 6–8 m s^{-1} in

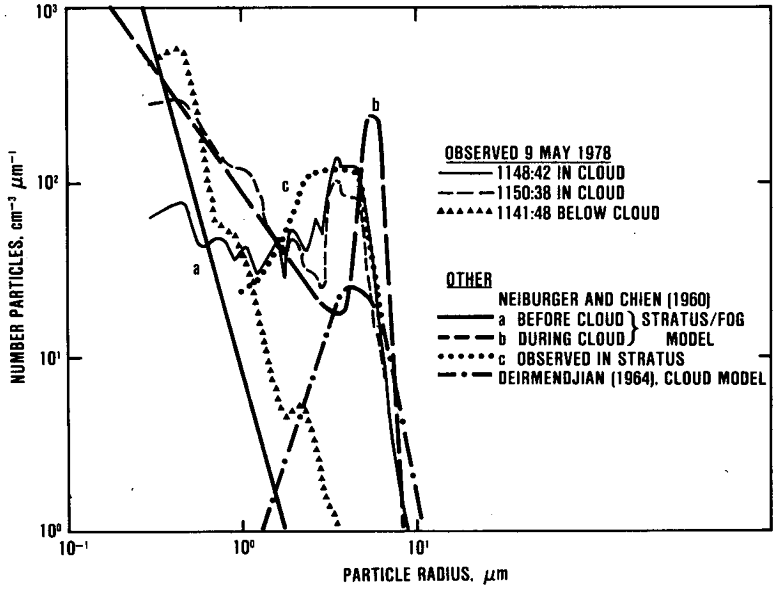

FIG. 3. Observed and modeled aerosol spectra for stratus clouds.

the large radius range with an arbitrary extension of aerosol spectrum into the Junge-Aitken radius range. A mode in $n(r)$ was found to form in the 3–5 μm region ~ 60 m above the cloud base for an updraft velocity of 1 m s^{-1}. In a similar computational study for salt nuclei Saad *et al.*, (1976) found a mode to form in the 3–5 μm region at a distance of ~40 m above the cloud base for an updraft velocity of 1 m s^{-1}.

The formation of a maximum near an r of 4 μm, signifying the formation of cloud, occurred when w exceeded ~0.05 g m^{-3} on 9 May. The minimum liquid water content has been reported to range from 0.01 to 0.05 g m^{-3} for stratiform clouds (Houghton, 1951) and fogs (McCartney, 1976). The profile of $(\bar{w} + MD)(z/z_i)$ in Fig. 2 exceeds 0.05 g m^{-3} between $0.76z_i$ and $0.93z_i$ indicating a cloud depth of 25 m, in close agreement with estimates of the cloud depth (~30 m) made aboard the aircraft.

Average aerosol size spectra $\bar{n}(r)$ were constructed for various values of $\bar{w}$ (narrow range of w in each $\bar{w}$) in the band $10^{-3} < w < 10^{-1}$ g m^{-3} and are displayed in Fig. 4. These average spectra show a systematic change in $\bar{n}(r)$ as $\bar{w}$ increased. In particular, a maximum in $\bar{n}(r)$ formed when $\bar{w}$ exceeded about 0.05 g m^{-3}. The slope $\Delta(\log n)/\Delta(\log r)$ for $r > 5$ μm approached a constant when $\bar{w} > 0.05$ g m^{-3} as exhibited in Fig. 4 and in spectra not presented here. If this slope remained constant for $r > 14.7$ μm, unobserved aerosols for $r > 14.7$ μm would contribute an insignificant amount of liquid water to the total when $w > 0.05$ g m^{-3}. The number of aerosols with $r > 5$ μm did not increase, and the number of aerosols with $r < 1.5$ μm decreased after the mode formed.

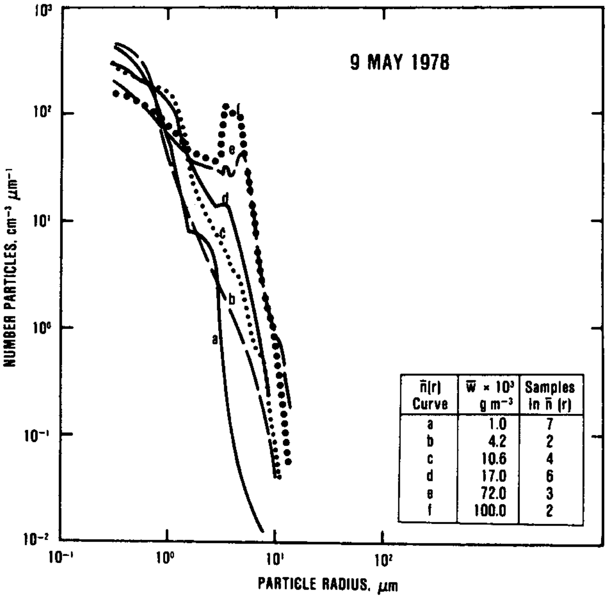

FIG. 4. Average aerosol spectra for six values of the average liquid water content $\bar{w}$.

TABLE 1. Spectral parameters for convective conditions on 9 May 1978 arranged according to the liquid water content. The exponents given by numbers are the powers of 10 by which the numbers must be multiplied.

Central spectral time (PST)	Liquid water content w (g m^{-3})	Total particle count (cm^{-3})	Modal radius r_{mo} ($r > 1$ μm) (μm)	Aerosol density at r_{mo} (cm^{-3} μm^{-1})	Represents*
1150:30	1.55^{-4}	208	none	—	O
1141:50	5.31^{-4}	222	none	—	O
1150:58	1.90^{-3}	239	none	—	O
1148:06	3.60^{-2}	304	none	—	N
1150:52	5.53^{-2}	204	2.6, 4.7	77, 59	C
1150:34	7.29^{-2}	304	4.7	50	C
1150:38	9.70^{-2}	424	3.5	108	C
1148:42	1.02^{-1}	361	3.5	115	C
Neiburger and Chien (1960); observed		464	3.5	115	C

* O: out of cloud; N: near cloud; C: in cloud.

Although the aerosol spectra in Fig. 4 represent different volumes of air, the spectra will be assumed to be taken from the same volume of air and compared with the model spectra given by Neiburger and Chien (1960). They generated a temporal sequence of spectrum by cooling air with an initial aerosol spectrum of salt nuclei only (no mode above $r = 10^{-2}$ μm) at a rate of 6°C h^{-1} at constant pressure and an RH of 75% until supersaturation was attained (cooling for 6000 s). Their resulting sequence of spectrum at times up to 3000 s are highly similar to those in Fig. 4 as partially displayed in Fig. 3. Based on their discussion and aerosol growth theory, the mode near 3–4 μm in Fig. 4 represents an accumulation of activated aerosols growing from smaller r. The decrease in the number of small aerosols must have been caused by the evaporation of unactivated aerosols adjusting their radius to the decrease in water vapor pressure being reduced by the rapid uptake of vapor by the activated aerosols. The average spectrum f in Fig. 4 represents elevations just below z_i where adiabatic cooling attains a maximum. Continued vertical motion without mixing with the surrounding air would be expected to shift slowly the single mode to larger radius and correspondingly slowly increase the number of aerosols at r above the mode as demonstrated by Neiburger and Chien. However, observations have shown that the spectrum broadens more rapidly to include larger droplets and usually has multiple peaks in deeper convective clouds. As recently discussed by Telford and Chai (1980), sequential mixing of rising and sinking cloud elements can produce the observed deep-cloud spectra, and details of supersaturation and aerosol growth are important only in the initial formation of a cloud element. Thus, the average spectra shown in Fig. 4 appear to represent a rising air parcel where the aerosol size distribution is controlled by the details of aerosol growth such as given by Neiburger and Chien (1960).

4. Summary and discussion

The temporal evolution of the aerosol size spectrum $n(r)$ as the relative humidity increases in a convective marine layer has been modeled by several authors (e.g., Neiburger and Chien, 1960; Mordy, 1959; Saad *et al.*, 1976), but appropriate measurements for their verification have been sparse. Several types of independent data showed that convective activity was present near San Nicolas Island on 9 May and that the air was marine. The liquid water content w, calculated from $n(r)$, was used as an indirect indicator of the relative humidity or water vapor uptake by aerosols. The appearance of a maximum in $n(r)$ near an r of 4 μm was used as an indicator of the formation of clouds.

The data suggest that characteristics in the aerosol spectrum $n(r)$ change systematically in a rising cell of air, while the liquid water content increases steadily. As the liquid water content increased to $\sim$0.05 g m^{-3}, a peak in $n(r)$ formed near an r of 4 μm and the number of small aerosols ($r < 1$ μm) decreased revealing the formation of a cloud. The maximum w observed was 10^{-1} g m^{-3} in the thin, patchy, stratus clouds observed to be $\sim$30 m thick just below the temperature inversion base near 147 m. These systematic changes in the spectrum characteristics support several marine, stratus cloud models and are considered to characterize conditions in a rising moist cell on 9 May. Large horizontal variations in w and $n(r)$, indicated by the large dispersion of w around $\bar{w}$, were probably created by horizontal/temporal changes created by rising and sinking air.

The relations between $\bar{w}(z/z_i)$ and $\bar{n}(r)$ found here may apply to similar moderate convective conditions over the ocean if the inversion depth and cloud base elevation do not depart appreciably from 150 to 120 m, respectively.

REFERENCES

Blanc, T. V., 1978: Micrometeorological data for the 1978 Cooperative Experiment for West Coast Oceanography and Meteorology (CEWCOM-78) at San Nicolas Island, California. Naval Research Lab. Memo Rep. 3871 [Available from Naval Research Lab., Washington, DC].

Deardorff, J. W., 1976: On the entrainment rate of a stratocumulus-topped mixed layer. *Quart. J. Roy. Meteor. Soc.*, **102,** 563–582.

Deirmendjian, D., 1964: Scattering and polarization properties of water clouds and hazes in the visible and infrared. *Appl. Opt.*, **3,** 187–196.

Dytch, H. E., and N. J. Carrera, 1976: Cloud droplet spectrometry by means of light-scattering techniques. *Atmos. Tech.*, No. 8, 10–16 (unpublished manuscript).

Fitzjarrald, D. E., 1978: Horizontal scales of motion in atmospheric free convection observed during the GATE experiment. *J. Appl. Meteor.*, **17,** 213–221.

Friehe, C. A., and K. F. Schmitt, 1976: Parameterization of air-sea interface fluxes of sensible heat and moisture by the bulk aerodynamic formulas. *J. Phys. Oceanogr.*, **6,** 801–809.

Houghton, H. G., 1951: On the physics of clouds and precipitation. *Compendium of Meteor.*, Thomas F. Malone, Ed., Amer. Meteor. Soc., p. 165.

Jensen, D. R., 1980: Intercomparison of PMS Particle Size Spectrometers. Naval Ocean Systems Center Tech. Note 555, San Diego, CA 92152.

Larsen, R. E., and B. J. Bressan, 1980: Air mass characteristics over coastal areas as determined by radon measurements. *Preprints Second Conf. Coastal Meteorology*, Los Angeles, Amer. Meteor. Soc., 94–100.

McCartney, E. J., 1976: *Optics of the Atmosphere*. Wiley, 408 pp.

Mordy, W., 1959: Computations of the growth by condensation of a population of cloud droplets. *Tellus*, **11,** 16–44.

Neiburger, M., and C. W. Chien, 1960: Computations of the growth of cloud drops by condensation using an electronic digital computer. *Physics of Precipitation*, Monograph No. 5, Amer. Geophys. Union, 191–209.

Niziol, T. A., C. W. Rogers, E. J. Mack and C. K. Akers, 1980: Characteristics of the marine boundary layer in the vicinity of San Nicolas Island during CEWCOM-78. *Preprints Second Conf. Coastal Meteorology*, Los Angeles, Amer. Meteor. Soc., 101–104.

Pruppacher, H. S., and J. D. Klett, 1978: *Microphysics of Clouds and Precipitation*. D. Reidel, 714 pp.

Rosenthal, J., T. E. Battalin, H. Hendon and V. R. Noonkester, 1980: Marine/continental history of aerosols at San Nicolas Island during CEWCOM-78 and OSP III. *Preprints Second Conf. Coastal Meteorology*, Los Angeles, Amer. Meteor. Soc., 84–93.

Saad, A. N., J. Podzimek and J. C. Carstens, 1976: Some remarks on modeling in the early stage of cloud formation in a simulation chamber. *J. Appl. Meteor.*, **15,** 145–156.

Telford, J. W., and S. K. Chai, 1980: A new aspect of condensation theory. *Pure Appl. Geophys.*, **118,** 720–742.

Warner, J., 1955: The water content of cumuliform cloud. *Tellus*, **7,** 449–457.

Woodcock, A. H., 1953: Salt nuclei in marine air as a function of altitude and wind force. *J. Meteor.*, **5,** 362–371.