The vertical distribution of tropospheric aerosols at Barrow, Alaska

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ABSTRACT

The vertical distribution of tropospheric aerosols near Barrow, Alaska (latitude 71°21′ N, longitude 156°30′ W) was determined by measuring the optical atmospheric transmission in the mid-visible at different altitudes with an airborne photometer. Measurements of the vertical aerosol distribution for periods in April and July 1972, are discussed. The aerosol concentration for both periods decreased exponentially with increasing height with a scale height equal to 1.4 ± 0.3 km. A seasonal variation in turbidity (and the corresponding columnar aerosol loading) was found. In general, the spring turbidity values were larger than the mid-summer values, thereby suggesting an aerosol mechanism which operates at low temperatures. It is hypothesized that the aerosols may be ice crystals seeded by open leads and mixed into the troposphere. A minimum value of aerosol optical depth ($\tau_D = 0.05 \pm 0.01$) occurred after the passage of a cold front. Possible effects of ice crystal aerosols on the heat budget of the arctic basin are discussed.

Introduction

Particulates, or atmospheric aerosols, exist in quasiequilibrium on a global scale and are found even in air samples obtained from remote regions of the earth such as over the Greenland ice pack or above Antarctica (Kuhn, 1972; Fischer, 1967). Although normally comprising only a small proportion of atmospheric mass loading, the aerosols, by virtue of their size, strongly interact with the optical and infrared radiation fluxes in the atmosphere. Thus, the radiation balance with its ramifications in climatology (Mitchell, 1971; Schneider, 1971) can be affected by the presence of particulate matter contained in the air. Aerosols may also have importance in influencing cloud growth.

Since the radiation balance is a particularly sensitive parameter for climate in the polar regions (Fletcher, 1965), it is imperative to learn more about the generation, transportation and removal processes of aerosols in arctic and subarctic air masses. We report here the initial results of continuing measurements made at Barrow, Alaska (Fig. 1), of the vertical distribution of the volume optical extinction coefficient. This parameter was determined by measuring, at different altitudes, the depletion of direct solar radiation due to scattering and absorption by the atmospheric aerosols. Assuming that the relative size distribution of aerosols does not change with height, the extinction coefficient is, in turn, related linearly to the concentration of aerosols. Thus, under the restriction of this assumption, the present report discusses the features of the vertical distribution of mass loading of atmospheric aerosols. The measurements should be representative of the vertical distribution of aerosols over the arctic basin in general.

Remarks on the theory

The flux of solar radiation $F(\lambda, h)$ at wavelength λ and height h decreases in magnitude as it passes through the atmosphere because of a combination of absorption and scattering by atmospheric gases and aerosols. Scattering processes remove energy from the incident beam at all wavelength intervals; absorption, on the other hand, occurs only in discrete wavelength intervals.

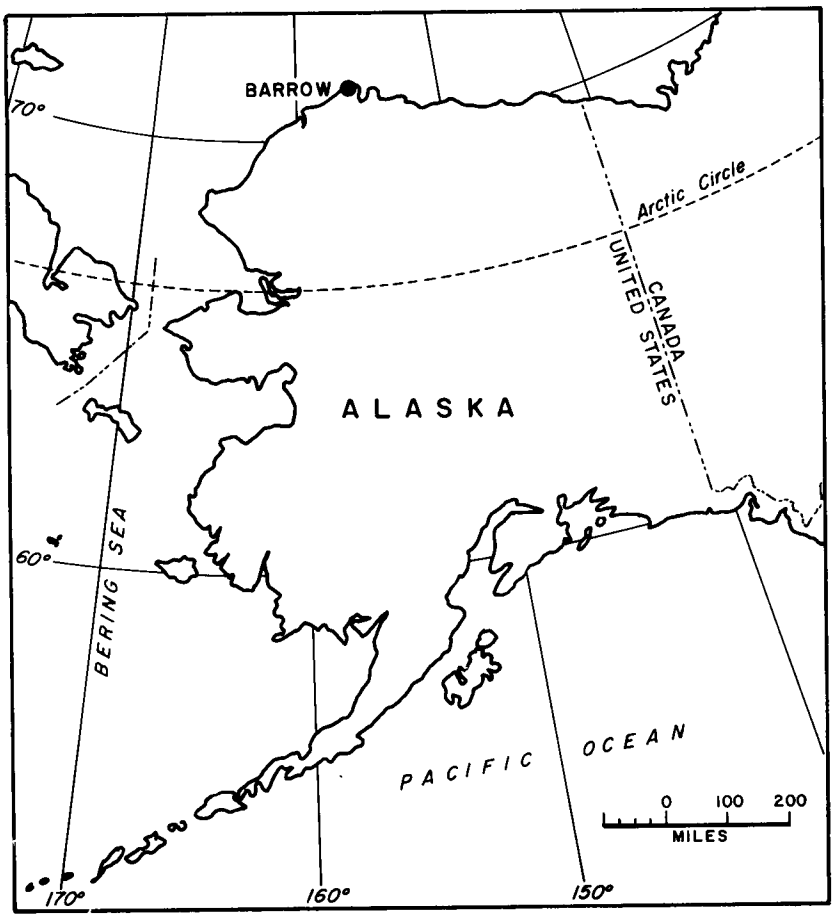

Fig. 1. Map showing measurement location at Barrow, Alaska.

For an optically thin atmosphere ($\tau < 1$), one can express the height and wavelength dependence of solar flux passing through the atmosphere as

$$F(\lambda, h) = F_0(\lambda) \exp -m(z) \left(\int_h^\infty \beta'(h, \lambda)\, dh' \right) + S(\tau, \lambda, z)\, \Delta\omega_i \qquad (1)$$

where

$F_0(\lambda)$ = incident solar flux at wavelength λ (watts, cm^{-2}, nm^{-1}),

$F(\lambda, h)$ = direct solar flux as viewed from a height h above the surface and at wavelength λ (watts cm^{-2} nm^{-1}),

$m(z)$ = air mass at solar zenith angle z,

$\beta(h, \lambda)$ = optical extinction coefficient at h, λ (km^{-1}),

$S(\tau, \lambda, z)$ = diffuse sky intensity (watts cm^{-2} nm^{-1} sr^{-1}),

$\Delta\omega_i$ = instrumental field of view (steradians).

The solar zenith angle, z, is determined by solving the celestial spherical triangle,

$$\cos(z) = \sin(l) \sin(\delta) + \cos(l) \cos(\delta) \cos(T_0) \qquad (2)$$

where

l = local latitude,

δ = solar declination at time of observation,

T_0 = hour angle of sun referenced to solar noon.

The optical extinction coefficient, $\beta(h, \lambda)$, results from a linear combination of terms due to molecular (Rayleigh) scattering, β_R, absorption by gases, β_G, and scattering and absorption by atmospheric aerosols, β_D.

$$\beta(h, \lambda) = \frac{p(h)}{p(0)} \beta_R(0, \lambda) + \beta_G(h, \lambda) + \beta_D(h, \lambda) \tag{3}$$

where $p(h)$ and $p(o)$ are the atmospheric pressures at heights h and at ground level, respectively. The optical depth (referred to the vertical direction) is defined in the usual manner as,

$$\tau(h, \lambda) = \int_h^\infty \beta(h', \lambda)\, dh' \tag{4}$$

Whenever the solar elevation angle $(\pi/2 - z_0)$ is larger than about 25°, the air mass, $m(z_0)$, is simply equal to the secant of the solar zenith angle $[m(z) = \sec z_0]$. For smaller values of elevation angle, however, account must be taken of effects imposed by the earth's curvature. Since measurements reported here were taken at high latitude (71° N), it was sometimes necessary to acquire data during times when the sun's elevation angle was low, of the order of 20°. For these occasions, we used values of air mass calculated by Kasten (1964) for a model arctic atmosphere.

The term in eq. 1, $S(\tau, \lambda, z)\Delta\omega_i$, can be expressed to a reasonable degree of accuray by the single-scatter solution to the equation of transfer (Bullrich, 1964) which, written in terms of appropriate variables is

$$S(\tau, h, z) = \frac{m_0}{m - m_0} \cdot P_0(\lambda, h) \times \left\{e^{-\tau(h,\lambda)m(z)} - e^{-\tau(h,\lambda)m_0(z)}\right\} \cdot F_0 \tag{5}$$

where

$m(z)$ = air mass of direction of view,
z = zenith angle of direction of view,
$P(\theta)$ = scattering phase function normalized to $\int P(\theta)dw = 4\pi$,
θ = scattering angle.

Diffuse sky flux, evaluated in a field of view centered on the sun ($m \to mo$, $\phi \to \phi_0$, where ϕ = azimuth), is obtained from eq. 5 by employing La Hospitals rule to derive,

$$S(\lambda, h, z_0)\Delta\omega i = \frac{\overline{P(\theta)}}{4\pi} \frac{F_0(\lambda)\tau(h, \lambda)e^{-\tau(h,\lambda)m_0}}{m_0} \tag{6}$$

where $\overline{P(\theta)}$ is the mean phase function over the detector's field of view.

Numerical evaluations of eq. 6 were made for an atmosphere with the following parameters: $\tau_D = 0.1$, $\lambda = 0.5$ μ, $h = 0$, $\Delta\omega = 5.95 \times 10^{-3}$ steradians (corresponding to a circular field of view of 5°), $m = 2$. The function $\overline{P(\theta)}$ was calculated by Mie theory for a Junge size distribution of particulates $dn/d \ln (r) = r^{-\gamma}$ with $\gamma = 3.0$. We remark that the above conditions represent a rather turbid atmosphere. (Linke turbidity coefficient $T = 2.02$; Ångström turbidity coefficient $\beta = 0.041$.) For the above set of conditions, the contribution of scattered sunlight in the field-of-view represents only approximately 1 part in 10^3 of the direct solar flux which enters the detector, and therefore the term $S(\tau, h, z_0)$ was ignored for later calculations.

For this study, the optical extinction coefficient, $\beta(h, \lambda)$, was experimentally derived by measuring the direct solar flux, $F(h, \lambda)$ as a function of altitude, h, and employing the relationship,

$$\beta(h, \lambda) = d \ln (F(h, \lambda))/(m \cdot dh) \tag{7}$$

The known extinction coefficients arising from molecular scattering, $\beta_R(h, \lambda)$ and gaseous absorption, $\beta_G(h, \lambda)$, were then subtracted from the experimentally determined value of β to finally yield the aerosol extinction coefficient $\beta_D(h, \lambda)$ which constitutes the basic data for this report. In the end, a check on the accuracy was made by comparing the observed aerosol ground-level optical depth with the optical depth calculated by integrating the extinction coefficient through the atmosphere (eq. 4). In the event of slight disagreement, the values of β were adjusted slightly until the two optical depths agreed. In practice the calculated and observed ground-level optical depths generally checked to within a few percent.

Instrumentation

A portable photometer was designed to measure the direct quasi monochromatic solar flux

in the visible region. The photometer was carried aloft in a Cessina 180 aircraft; the instrument was hand held and pointed toward the sun through an aperture in the aircraft window. Narrow-band interference filters were employed to define quasi monochromatic samples of solar flux. The instrument employed a solid-state silicon photo detector. Incoming light was modulated with a mechanical chopper, and the DC temperature-dependent diode leakage current was effectively suppressed by amplifying only the sinusoidal component that resulted from the modulated light beam. Laboratory calibration indicated that the instrument was drift-free, temperature independent and linear to better than 1 % over its useable range.

Experimental measurements

General remarks

All airborne measurements were made within the vicinity of the Naval Arctic Research Laboratory (NARL) at Barrow (Latitude 71°21° W, longitude 156°30′ W). A total of eight separate flights were made during the period 4 April 1972 to 23 April 1972. This period was chosen because of the favorable chances for clear sky conditions, and also because concurrent surface studies were being carried out further north on the pack ice at the AIDJEX camp site. Additional summer measurements were taken from 11 to 13 July, 1972. The data were typically acquired to a maximum altitude of about 3.5 km. Height was derived from the aircraft altimeter; air temperature was recorded with a thermistor probe mounted on the wing tip.

Climb rate and speed limitations of the aircraft required that approximately 100 km of horizontal distance be spanned to reach maximum altitude and it is not necessarily valid to assume horizontal homogeneity over distances of this scale. As a check we would periodically maintain a constant altitude for one or two minutes to insure that the solar intensity remained constant. Upon occasion, we found, by using this method, that the solar intensity would inexorably decrease (or increase) thereby indicating a change in turbidity with horizontal scale. At certain times when the sky appeared to be clear the solar intensity was noted to decrease slightly, all the time while flying at a constant altitude, until eventually one would become aware of very tenuous cirrus clouds near the sun. Whenever this occurred the data were excluded from analysis.

Report on measurements made on April 8, 1972

During this flight the solar elevation was in the range of 25.7° to 23.6°. The sky was clear; visibility approximately 20 km; ground temperature −22.5°C; winds calm. Climb-rate was adjusted to 2.5 m/sec; after leveling out at 3.3 km (10 000 ft) the aircraft then descended to ground level. Periodic back tracks were made to insure horizontal homogeneity; the maximum range extended 50 km from Barrow.

Fig. 2 illustrates the aerosol extinction coefficient at $\lambda = 500$ nm as determined by the analysis method described in the previous sections. Plotted adjacent to the optical extinction coefficient is the temperature profile. The radiative inversion extended to a height of 500 m (1 500 ft) with an average lapse rate of +34.0°C/km; thereafter, a roughly isothermal region extends to about 1 810 m (6 000 ft) elevation after which the lapse rate approached the moist adiabatic value of −6°C/km for higher altitudes. There is some slight structure in the temperature curve with a relative maximum (warm region) near 920 m (3 000 ft). The optical extinction coefficient, $\beta_D(h, 0.5\ \mu)$, shows a general decrease with altitude. The individual points have rather considerable scatter, mainly due to error in reading the slope of the $I(\lambda, h)$ vs. h curve. The tendency for the extinction coefficient to peak near 460 m (1 500 ft) altitude may be real; other short periodicities are probably noise and can be averaged out. A line drawn through the data points gives a relation

$$\beta_D(h, 0.5\ \mu) = 0.25\ e^{-h/H}(\mathrm{km}^{-1}) \qquad (8)$$

where $H = 1.5$ km, $h < 3.5$ km.

Ground measurements carried out before this flight indicate that the total optical depth was equal to 0.550 at $\lambda = 500$ mm. Subtraction of extinction terms due to molecular scattering and ozone absorption leaves a residual aerosol optical depth $\tau_D\ (0.5\ \mu) = 0.39$. This can be compared with Elterman's (1968) standard mid-latitude value of $\tau_D(0.5\ \mu) = 0.26$. The large optical depth is surprising considering the remoteness of Barrow. Measurements of atmosphere transparency at McCall (69°18′ N,

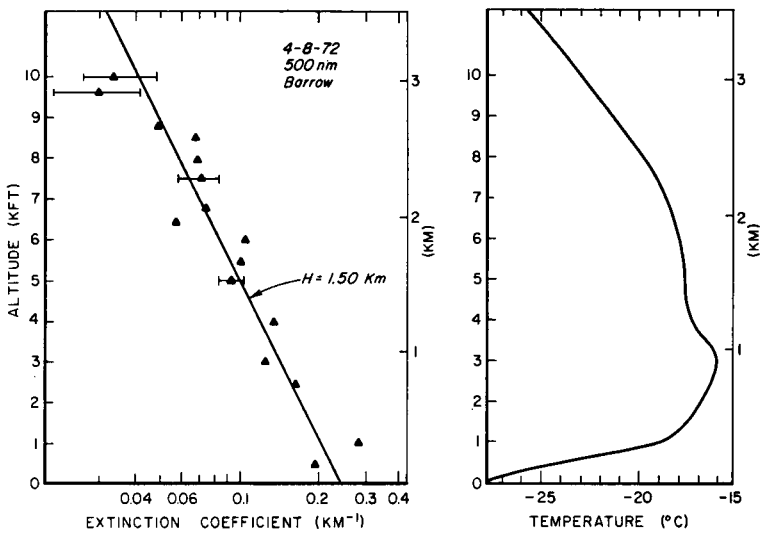

Fig. 2. Optical extinction coefficient arising from scattering and absorption by aerosols at wavelength 500 nm and air temperature as a function of altitude for the region of Barrow, Alaska on April 8, 1972.

143°48′ W) in northeast Alaska in late spring and summer months yielded an Ångström turbidity coefficient of about 0.030 (Shaw & Wendler, 1972) corresponding to an aerosol optical depth (at $\lambda = 0.5\ \mu$) of 0.074 (assuming a $\lambda^{-1.3}$ wavelength dependency). At 1 700 m (the altitude of McCall Glacier) our measured aerosol optical depth, τ_D, was numerically equal to 0.082, slightly larger than the average clear-sky McCall Glacier value for spring and early summer.

We note that the ground-level aerosol optical extinction coefficient $\beta_D(0,\ 0.5\ \mu) = 0.25 \pm 0.05$ is remarkably large. Porch et al. (1970) report measurements made near Barrow in March 1970 using an integrating scattering nephelometer. They report the value of $b_{\text{scat}}/b_{\text{rayleigh}} = 1.95$, where b_{scat} and b_{rayleigh} represent the total scattering (aerosol plus molecular) and molecular (rayleigh scattering) coefficients integrated over the scattering geometry of the nephelometer. Making the assumption that the "mean" scattering coefficient b_{scat} and b_{ray}, are strictly proportional to the corresponding extinction coefficients, $\beta_{\text{ray}} + \beta_{\text{mie}}$ and β_{ray}, one obtains from Porch's et al. measurements, $\beta_{\text{aerosol}}\ (0,\ 0.55\ \mu) = 0.011\ \text{km}^{-1}$, much smaller than the ground-level extinction coefficient observed on April 8, 1972. There are several possible causes for the rather substantial disagreement. Firstly, aerosols scatter light predominantly into the forward or near-forward, direction. The integrating nephelometer used by Pilot et al. may not have "captured" some of the forward-scattered radiation, which would result in underestimating the optical extinction coefficient. Also, it is possible that some or all of the aerosols may be ice crystals which could have sublimed before entering into the scattering chamber. Finally, it may simply be that the air in the vicinity of Barrow was "cleaner" during the time Pilot et al. accumulated their readings. It would be useful if, in the future simultaneous nephelometer measurements and monochromatic solar extinction measurements could be performed.

Some clue to the cause of the high turbidity can be found by inspecting the synoptic maps. Fig. 3 shows the synoptic situation at the surface and at the 850 mb level during the time of the flight just described. By inspecting previous maps and temperature profiles, it was determined that a cold polar air mass had recently been intruded by a relatively warm continental southerly flow. Surface winds were from the south; at higher levels the winds veered clockwise to 290° at 830 mb, to about 10° at 500 mb. It is conceivable that the high turbidity had its origin, at least in part, from the southerly trajectory over interior Alaska which

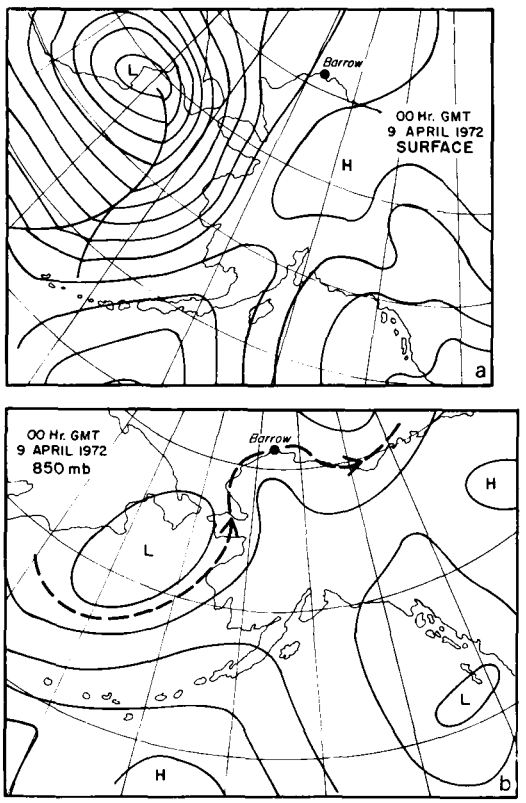

Fig. 3. Synoptic maps for 0 hr UT, 9 April 1972. (*a*) Surface; (*b*) 850 mb level.

may have advected aerosols into the area. The high values of turbidity could also result, at least in part, from ice crystals or blowing snow from the surface.

Measurements of April 10, 1972

This flight extended from 14: 22 to 15: 53 local mean time (LMT = UT-10 hr). The sky was clear; visibility 20 km; wind ENE at 15 knots; ground temperature −17°C. The solar elevation angle during the flight varied from 23.5° to 19°. Very tenuous cirrus clouds were seen close to the horizon after gaining altitude and carefully scanning the sky, but the sky near the sun appeared clear.

Optical data were accumulated at 0.5 and 0.7 μ. There was evidence that horizontal inhomogeneities degraded the data at 0.7 μ. However, the data acquired at 0.5 μ during both ascent and descent seem to be consistent and thus these measurements will be discussed:

The aerosols extinction coefficient (at $\lambda = 0.5\ \mu$) evaluated at the ground and at peak altitude are:

$\tau_D = 0.280$ at ground level

$\tau_D = 0.062$ at $h = 3.60$ km (12 000 ft)

Fig. 4 shows the extinction coefficient. The scatter in the data points is caused mostly from errors in reading dF/dh (see eq. 7) from the analog record. The process is statistically unbiased, however, and thus the average trends are meaningful. The extinction coefficient is again seen to fall off exponentially. The scale height is 1.2 km.

It is possible that the stratosphere may contain a slight aerosol loading which could cause a detectable optical depth. In order to estimate this possible residual stratospheric optical depth component, we proceeded as follows: First, the optical depth remaining above h_{max}, the maximum altitude reached by the aircraft, was estimated by comparing the solar intensity readings at h_{max} to the known extraterrestrial value of solar intensity. (The photometer had previously been accurately calibrated in absolute irradiance units.) The total aerosol optical depths above h were then calculated by subtracting the molecular scattering and ozone absorption components. Assuming a continuation with increasing height of the exponential fall-off of aerosol extinction coefficient, one may deduce a value for the remaining tropospheric optical depth by integrating from h_{max} to the mean tropopause height (about 11 km). One can then obtain the stratospheric aerosol optical depth as the remainder between the total optical depth above h_{max}, and the calculated tropospheric optical depth. Doing this it was found that τ_D (stratosphere) = 0.050. It is difficult, however, to accurately assess optical depths of these small magnitudes and an error analysis indicates that the value may be in error by approximately ± 0.03.

Three day sequence of measurements during July 1972

Optical measurements were made for three consecutive clear days in July 1972. Table 1 summarizes the flight time periods and surface conditions: Fig. 5 illustrates the aerosol optical depth that remains above height h for the time

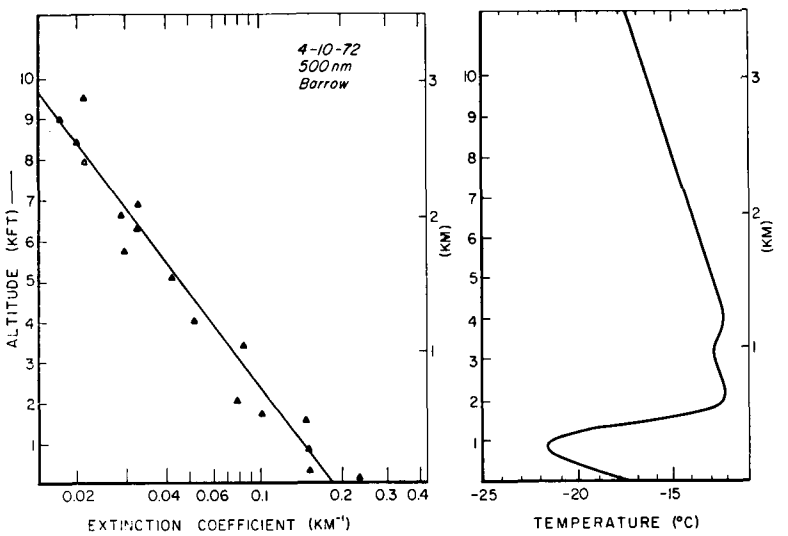

Fig. 4. Optical extinction coefficient arising from scattering and absorption by aerosols at a wavelength of 500 nm and air temperature near Barrow, Alaska for April 10, 1972.

of each of the three flights. The aerosol optical depth during flight 1B is seen to be essentially independent of altitude, implying that the entire aerosol content resides above 4 km and thus the air in the lower troposphere during this flight contained an extremely small concentration of aerosols. The aerosol optical depth increased significantly, however, during the 24-hour period between flights 1B and 2B. Inspection of Fig. 5 indicates that the increased optical depth is due almost entirely to increased extinction (by aerosols) in the height region 0.9 km $< h <$ 3.7 km. Twenty-four hours later, during flight 3B, conditions changed little, although a slight decrease in extinction coefficient is apparent at all levels. The sequence of data could not be continued for a 4th consecutive day because of the onset of cloudiness.

Table 1. *Surface parameters during 3-day sequence (Barrow, Alaska)*

Flight 1B	
Takeoff time	23: 37 UT July 11, 1973
Duration	110 minutes
Air temp.	+8°C
Wind	110°, 10 knots
Humidity	74 %
Sky conditions	Clear
Flight 2B	
Takeoff time	00: 21 UT July 13, 1973
Duration	70 minutes
Air temp.	+13°C
Wind	90°, 10 knots
Humidity	69 %
Sky cover	0.2
Flight 3B	
Takeoff time	00: 05 UT July 14, 1973
Duration	103 minutes
Air temp.	+11°C
Wind	80°, 10 knots
Humidity	69 %
Sky cover	0.1

The initially low turbidity during flight 1B, and the dramatic increase the following day may provide a clue to the generation and transportation mechanisms for the arctic aerosols. Reference to synoptic charts and climatological data for these periods indicates that a frontal system had passed through Barrow about 20 hours before the occurrence of flight 1B (see Fig. 6*a*, *b*). During the passage of this system, an exchange of air masses took place bringing polar air into the region from a northerly direction. A high pressure cell centered 300 km to the NW dominated the circulation during flight 1B; during this time the wind was from the east. Figs. 6*c* and 6*d* shows the synoptic situation at the surface and at 850 mb during the exceptional low turbidity period of flight 1B.

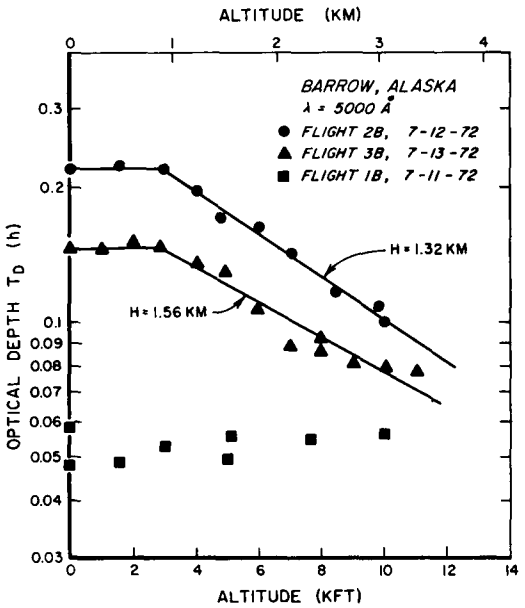

Fig. 5. Optical depth at a wavelength of 500 nm remaining above height h. Barrow, Alaska, July, 1972.

The following day during flight 2B the anticyclonic system to the NE was still dominant, however a warming of approximately 7°C over the level extending from the surface to the 850 mb level, took place in the previous 24-hour period. It is not immediately obvious why the aerosol loading should have increased, although it seems likely that aerosols might have been mixed into the air mass from the low pressure region located to the SW. Previously, during the first flight period, the cyclonic system to the south had not exerted much influence on the circulation near Barrow, but during the intervening period, between flights 1B and 2B, the high cell to the nothwest had weakened somewhat, and southerly cyclonic flow had become more dominant. Synoptic maps for the corresponding later time of flight 3B show that the low pressure system located to the SSE continued to intrude into the region of Barrow.

Discussion of data

As is well known, the vertical distribution of atmospheric aerosols depends on the interaction between numerous atmospheric variables including removal mechanisms, such as precipitation, scavenging and fallout and generation mechanisms, such as blowing dust, smoke from combustion processes and coagulation of ion clusters.

Vertical distribution

Penndorf (1957) found from airborne sampling, that the aerosol concentration in the troposphere decreases with height in an approximately exponential manner with a scale height of about 1.2 km. Elterman's (1968) bistatic searchlight experiment in New Mexico and ground-based lidar experiments indicate that generally the aerosols in the first several kilometers decrease with height exponentially with scale heights of one to two kilometers, although complex structure certainly can occur. The majority of columnar aerosol mass in Elterman's standard model (an average from many day's data) resides in the lowest few kilometers. The measurements at Barrow show that this is also the case for the polar aerosols; apparently those aerosols are also distributed with a scale height of $1.0 < H < 1.7$ kilometer which is in numerical agreement with the values found by Penndorf (1957) and Elterman (1968) for lower latitudes. We did not observe any difference in aerosol scale height in summer as compared to spring, probably because of the minor role that convection plays in the atmosphere at these latitudes.

Origin of the polar aerosols

Since no direct sampling measurements were conducted during these experiments one can only speculate about the probable chemical composition of the tropospheric aerosols at Barrow. However, Fenn et al. (1963) found by conducting direct sampling experiments on the Greenland icecap, that there was a natural polar background aerosol mass loading of approximately 1 $\mu g\ m^{-3}$. The majority of the aerosol mass was concentrated in the size range $0.4 < d < 1.5$ μ diameter; chemical analysis showed that approximately 40 % of the particulates were sulphates.

It is known that the vapor pressure over sulphate compounds is generally quite low, which implies that they can have long lifetimes in a zero vapor pressure environment. To take an example, the sublimation time for ammonium sulphate in an atmosphere with zero constituent

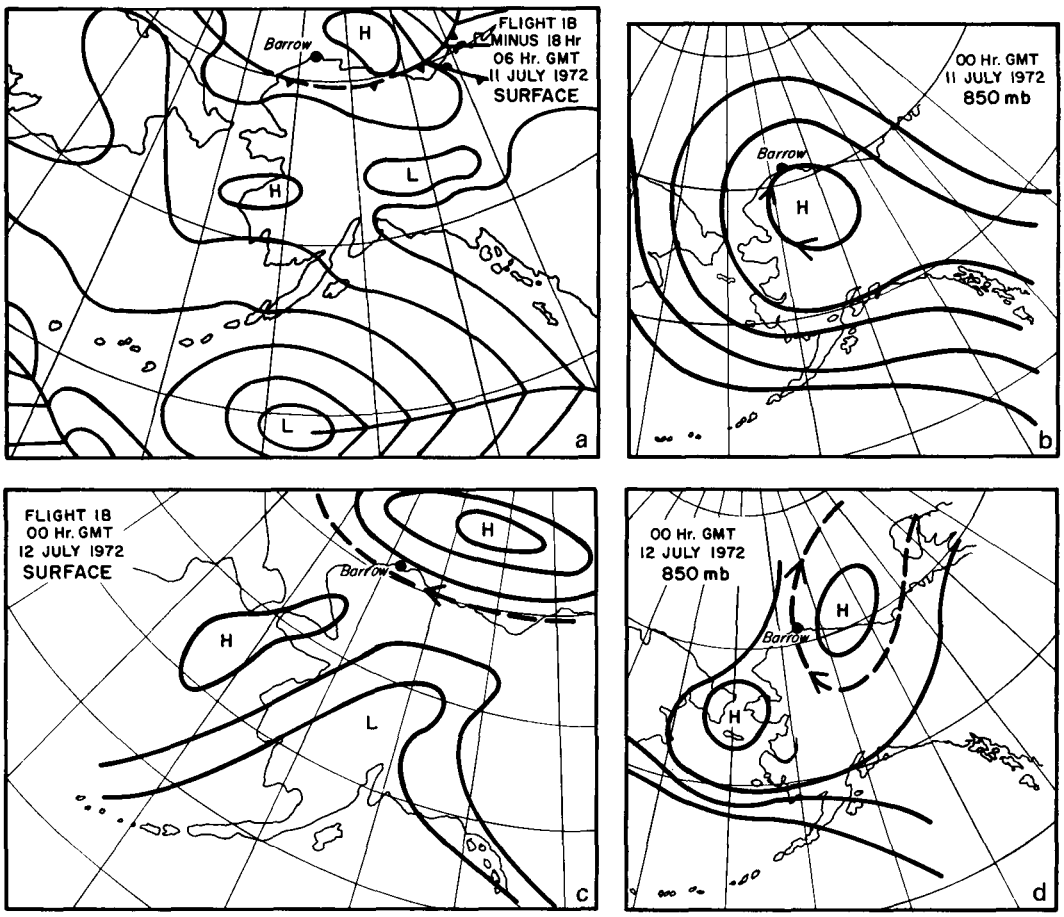

Fig. 6. Synoptic maps. (*a*) Surface conditions 18 hours prior to flight 1B, (*b*) isoheights at the 850 mb level 24 hours prior to flight 1B, (*c*) surface conditions during flight 1B, (*d*) isoheights at the 850 mb level during flight 1B.

vapor pressure is on the order of 4 days. This residence time is consistent with the hypothesis that sulphates may be advected into northern Alaska from great distances.

Another possible source for the aerosols near Barrow might involve the discovery by Scott & Duffy (1969) that indicates that stable solid products form from the reaction between ammonia and sulphur dioxide when temperatures are below approximately -10°C. Lamb (1971) presents curves of the equilibrium vapor pressure of these products as a function of temperature and mole fraction of NH_3 in the vapor phase. Whenever the partial pressure of NH_3 and SO_2 exceeds the equilibrium vapor pressure of the compounds, it is possible for these solid products to form. Taking probable estimates of the maximum concentrations of gaseous ammonia and sulphur dioxide at Barrow of 20 $\mu g\ m^{-3}$ and 5 $\mu g\ m^{-3}$ for SO_2 and NH_3 respectively (Junge, 1963), and by employing the equilibrium vapor pressure curves of Lamb (1971), it was determined that the equilibrium temperature at which the solid products can be formed is equal to $T = -60$°C. Since temperature minima in the troposphere over northern Alaska in the spring months are rarely below -50°C, it appears somewhat unlikely that the solid products resulting from the reactions between SO^2 and NH_3 could form and exist in equilibrium. However, this is marginal and cannot be ruled out as a possible source for the polar region aerosol in spring time and winter.

The mass loading of particulates was not ex-

plicitly measured during these experimental flights, but a minimum limit for mass loading may be obtained from the observed aerosol extinction coefficient by calculating the maximum possible extinction that can arise from a given mass loading of particulates. If the aerosols are assumed to be monodisperse spheres with radius r_0, then the following relation holds between the aerosol mass loading and extinction coefficient:

$$\beta(\alpha) = \frac{3\gamma}{4\varrho} \times 10^{-12} \frac{Q}{r_0} \tag{9}$$

where

γ = aerosol mass loading (μg m^{-3}),
Q = Mie scattering efficiency factor (ratio of extinction cross-section to geometric cross-section),
r_0 = particle radius (cm),
ϱ = compositional density of particulate material (g cm^{-3}),
β = optical extinction coefficient (cm^{-1}),
$\alpha = 2\pi r/\lambda$.

The maximum value of Q/r_0 for an assumed index of refraction equal to 1.52 (corresponding to ammonium sulphate) and a wavelength of 0.5 μ, is $Q/r_0 = 1.84$ μ^{-1} at $r_0 = 0.296$ μ. Thus, the maximum extinction possible (at $\lambda = 0.5$ μ) for a constant mass loading would occur if the particles had radii equal to 0.296 μ. If this were the case, the ground level mass loading necessary to cause the extinction coefficient observed during the flight of April 10, 1972 (0.19 km^{-1}) would be $\gamma = 140$ μg m^{-3}. Considering that the above procedure yields a minimum value, and realizing that the actual mass loading is more probably a factor of two or so larger than this, we conclude that the surface mass loading at Barrow on April 8, 1972, must have been of the order 300 μg m^{-3}, or appoximately two orders of magnitude larger than the background level found over the Greenland ice-cap by Fenn et al. (1963).

Seasonal variations

An analysis of the air mass trajectory at the 850 mb level for the times of the two flights in April with high turbidity indicates that for the past several days the air mass had been over interior Alaska. The greatest portion of the land area in interior Alaska is snow covered in April, and a strong surface inversion generally is present at these times, and hence it is unlikely that surface dust mixed into the air could have been a major contributor to the high turbidity. However, it does seem likely that the high turbidity could in part be due to ice crystals that are generated in open leads and then mixed into the ice-saturated troposphere.

The extremely low turbidities observed in July 1972, during flight 1B, shows that there are periods at Barrow when the atmosphere is exceedingly clean. However, in a time period of 24 hours the optical depth had once again increased to a moderate value ($\tau_D = 0.150$ at 0.5 μ). Since the major increase in optical extinction coefficient occurred for $h > 1$ km, it is most probable that the aerosols were advected into the region, rather than mixed in from ground sources. Trajectory analysis supports this assertion. The trajectory taken backwards in time indicates that the regions which were invaded by aerosols ($h > 1$ km) was connected with cyclonic flow, whereas the lower, and less turbid, levels had their air mass origin from the anticyclone flow to the NE. Thus there is evidence that the arctic aerosol, at least for the summer periods, may be associated with cyclonic flow. This suggests that perhaps hydroscopic nuclei may be nucleated in the higher humidities of cyclonic air flow. It is certain that ice crystals were not responsible for the summer turbidity increase as temperatures were above freezing to the 850 mb level.

Concluding remarks

The vertical distribution of aerosols in the lower troposphere near Barrow decreases in an approximately exponential fashion with a scale height of 1.0 to 1.7 km.

Ice crystals are probably largely responsible for the occurrence of large turbidites (Ångström turbidity coefficient $\beta = 0.160$) found at Barrow in spring time when air temperatures are below freezing at all levels. Ice has absorption features in the 8–12 μ window (Irvine) & Pollack, 1968), and therfore it is likely that these ice crystals mixed throughout the troposphere air would have an effect on the heat budget of the polar regions. In winter when solar radiation is negligible their presence would lead to

a net heating at the surface; however, the net heat loss from the earth-atmosphere system could either increase or decrease depending upon the vertical distribution of aerosols, and the temperature structure in the aerosol layer. If, for example, the aerosols were concentrated at the top of the surface inversion, then a net heat loss would occur from the aerosols.

During summer when the temperature of the lowest several kilometers are above freezing, the turbidity at Barrow became extremely low [$\tau_D(0.5\ \mu) \doteq 0.050$; Ångström turbidity coefficient, $\beta \doteq 0.020$.] following the passage of a cold front. Comparison of observed data with the synoptic situation indicates that summertime southerly cyclonic flow at Barrow enhances turbidity.

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ВЕРТИКАЛЬНОЕ РАСПРЕДЕЛЕНИЕ ТРОПОСФЕРНЫХ АЭРОЗОЛЕЙ В БЭРРОУ, АЛЯСКА

Вертикальное распределение тропосферных аэрозолей вблизи Бэрроу, Аляска (71°21′ с. ш., 156°30′ з. д.), определялось путем измерений оптического пропускания атмосферы посредине видимого диапазона спектра на различных высотах с помощью самолетного фотометра. Обсуждаются измерения вертикального распределения аэрозолей в периоды апреля и июля 1972 года. Концентрация аэрозоля для обоих периодов экспоненциально уменьшается с высотой со шкалой высоты, равной $1{,}4 \pm 0{,}3$ *км*. Найдена сезонная вариация в мутности (и в соответствующей величине оседания аэрозоля из столба атмосферы). В общем величины весенней мутности были больше летних, что предполагает механизм генерации аэрозоля, действующий при низких температурах. Предлагается гипотеза, что этим аэрозолем могут быть ледяные кристаллы, возникающие на открытых разводьях и вводимые над ними в тропосферу. Минимальные величины оптической толщи аэрозоля ($\tau_D = 0{,}05 \pm 0{,}01$) наблюдались после прохождения холодного фронта. Обсуждается возможный эффект аэрозоля кристаллического льда на тепловой баланс арктического бассейна.