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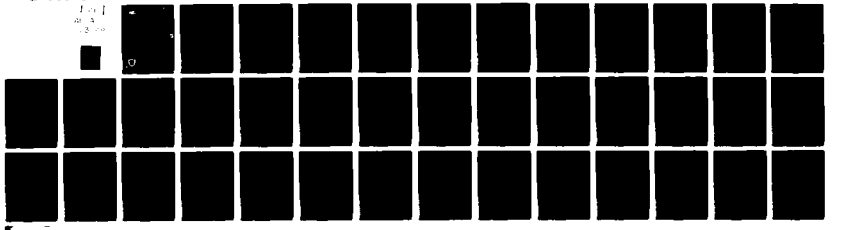

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SOME OPTICAL PROPERTIES OF BLOWING SNOW

JUNE 1981

By

Mary Ann Seagraves

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US Army Electronics Research and Development Command

Atmospheric Sciences Laboratory

White Sands Missile Range, NM 88002

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Windblown snow causes severe reduction in visibility and is a principal cause of obscuration in some regions of the world. Mechanisms of snow transport are presented along with particle size distributions, shape, and fall velocity. Relationships between visibility and windspeed and also between visibility and mass density in blowing snow are discussed. Variation of visible and infrared extinction coefficients with height are derived. In the near-millimeter wave region, the Rayleigh approximation is adequate for wavelengths greater than 4

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20. ABSTRACT (cont)

mm. Near-millimeter wave extinction coefficients are given as functions of visibility and visible extinction coefficients.

SUMMARY

Visibility may be severely reduced in blowing snow. It has been shown that visibility is inversely proportional to mass density at eye level and decreases with increasing windspeed.

Extinction in windblown snow is approximately independent of wavelength in the visible and infrared regions. The extinction coefficient, β_{vis}, in this wavelength region varies with height, Z, according to

$$\beta_{vis} = \beta_1 \left(\frac{Z}{Z_1} \right)^{-\frac{0.596}{u_*}},$$

where u_* is friction velocity and the subscripts indicate a reference height.

For near-millimeter wavelengths greater than 4 mm, the Rayleigh scattering approximation may be used to determine β. Extinction increases with increasing temperature over the range of -20°C to 0°C. The relationship between β and visibility, V, for wavelength, λ, is

$$\beta = \frac{2.09}{\lambda V} \text{ Im } \left(- \frac{m^2 - 1}{m^2 + 2} \right),$$

where m is the complex index of refraction for ice, λ is in units of millimeters, and V is in units of kilometers.

Note that the relationships derived in this report have not been verified by field measurements and may be subject to change should such measurements become available.

CONTENTS

INTRODUCTION

Windblown snow is one of the principal causes of obscuration in the polar regions and can occur in many other geographical areas. When the snow on the ground is dry and composed of fine-grained, loose particles, it is easily swept into the air by gentle or moderate winds. When the wind is in the range of 4 to 6 m/s, it raises snow crystals a few feet off the ground and obscures many surface objects and features. With speeds exceeding 7 m/s, wind can carry the snow to much greater heights. These critical windspeeds are often exceeded in the polar regions where blowing snow is reported on an average of 5 to 10 percent of the observations in the winter season.[1] At Barter Island Air Force Base, Alaska (latitude 70°N), blowing snow is reported on the average in about 25 percent of all observations in November. The snow blown along in a major blizzard in the polar regions is not necessarily confined to a very shallow surface layer but may form a cloud as deep as 300 m.[2]

Blowing snow occurs when the windspeed and the surface roughness are sufficient to develop a shear stress great enough to break particles free from the surface. The magnitude of this critical shear stress will vary with the size of the snow grains and with the degree of intergranular bonding in the surface layer. With cold, cohesionless, fine-grained snow, windspeeds of a few meters per second may suffice to dislodge particles but not to diffuse them into the airstream by turbulent exchange. Under these circumstances the particles will roll or bounce along the surface in a thin layer, commonly no more than 10 cm thick. Not until turbulence is well developed will particles be carried up to and above eye level.[2]

Snow particles may also be supplied directly by precipitation falling during the windstorm and would occur at much lower windspeeds than the particles lofted from the surface. In practice, it is difficult to distinguish between particles from the two sources, and in fact it may be difficult to determine whether precipitation is even occurring during the windstorm.

MECHANISMS OF SNOW TRANSPORT

Three transport mechanisms have been identified in the movement of windblown sand and snow and other fluid-borne sediments.[3] As illustrated in figure 1,[2] these mechanisms are:

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[1]J. M. Mitchell, Jr., 1958, "Visual Range in the Polar Regions with Particular Reference to the Alaskan Arctic," Polar Atmosphere Symposium Part I, Meteorology Section, Pergamon Press, London

[2]M. Mellor, 1965, "Blowing Snow," Monograph III - A3c, US Army Cold Regions Research and Engineering Laboratory, Hanover, NH

[3]R. A. Bagnold, 1941, The Physics of Blown Sand and Desert Dunes, Chapman and Hall, London

1. Surface creep in which particles roll or creep along the surface and generally remain in a layer less than 1 cm thick.

2. Saltation in which particles bounce along the surface and rebound and/or eject other particles into the air when they impact. Saltation particles are primarily contained in a layer above the surface 0.1 to 1.0 m thick.

3. Turbulent diffusion in which particles are held in suspension by vertical mixing.

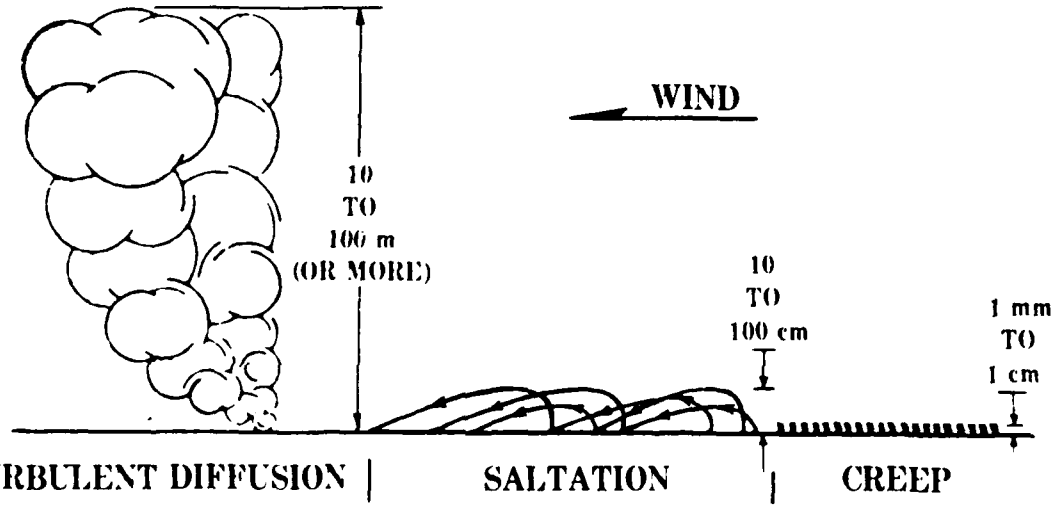

Figure 1. Methods of transport of windblown snow (Mellor[2]).

Turbulent diffusion is the dominant mechanism for snow transport when appreciable concentrations are carried along at and above eye level. The particles carried by turbulent diffusion will be the only ones considered here since they have the major effect on visibility and the transfer of electromagnetic energy through the atmosphere. However, the saltation and surface creep particles may affect the snow background and target contrast characteristics and in that sense would affect the operation of electro-optical devices.

When stationary conditions exist in which turbulence transports upward as much snow as settles downward under the influence of gravity, Loewe[4] found that

$$w\, s = K \left(\frac{ds}{dz}\right) , \qquad (1)$$

where

w = fall velocity of snow particles

s = particle concentration

K = eddy diffusivity

When the wind profile is logarithmic, as Budd, Dingle, and Radok[5] found to be the case in blowing snow, then the eddy viscosity, A, is proportional to height, that is,

$$A = k\, u_{*}\, (Z + Z_0) , \qquad (2)$$

where

k = 0.4 is von Karman's constant

u_{*} = friction velocity (usually 0.5 to 1.0 m/s over snow)

Z_0 = roughness length

If eddy diffusivity is equal to eddy viscosity, then (1) and (2) may be combined and integrated so that

$$s = s_1\, (Z/Z_1)^{w/ku_{*}} , \qquad (3)$$

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[4]F. Loewe, 1956, "Etudes de Glaciologie en Terre Adelie," Expeditions Polaires Francaises, Paris

[5]W. F. Budd, W. R. J. Dingle, and U. Radok, 1966, "The Byrd Snow Dirft Project: Outline and Basic Results," Studies in Antarctic Meteorology, American Geophysical Union, Washington, DC

where the index refers to a reference level.[6] Equation (3) is not valid near the surface where saltation and surface creep are predominant transport methods; nor is it valid when turbulence is strong enough to carry many more particles upward than are settling downward through gravitation.[7] It is assumed here that particles of different sizes settle at different speeds but do not interfere with each other and that equation (3) may be used to determine the variation of number density with height for particles of various sizes.

SHAPE AND FALL VELOCITY OF WINDBLOWN SNOW PARTICLES

Snow crystals blown about by strong winds are broken and abraded into roughly equidimensional grains with rounded or nearly angular corners. Particles occur in greatest numbers with effective radii in the range 10μm to 200μm, where the effective radius is 1/2 ✓ length x breadth (Mellor[2]).

According to Budd,[8] blowing snow particles are usually larger than those for which Stokes's law is valid for determining fall velocities but smaller than those for which a linear dependence upon size is applicable. However, in this study Stokes's law for spherical particles was used to determine w:

$$w = -\frac{2r^2\rho}{9\eta}, \qquad (4)$$

where

r = particle radius

g = gravitational acceleration

ρ = particle density

η = coefficient of viscosity

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[6]R. Sommerfeld and J. A. Businger, 1965, "The Density Profile of Blown Snow," Journal of Geophysical Research, 70:3303-3306

[7]J. A. Businger, 1965, "Eddy Diffusion and Settling Speed in Blown Snow," Journal of Geophysical Research, 70:3307-3313

[2]M. Mellor, 1965, "Blowing Snow," Monograph III - A3c, US Army Cold Regions Research and Engineering Laboratory, Hanover, NH

[8]W. F. Budd, 1966, "The Drifting of Nonuniform Snow Particles," Studies in Antarctic Meteorology, American Geophysical Union, Washington, DC

This determination results in computed fall velocities which are usually somewhat low, perhaps as much as 10 percent. The low velocities cause larger particle number densities at higher altitudes than would otherwise occur.

PARTICLE SIZE DISTRIBUTIONS

Budd, Dingle, and Radok;[5] Dyunin;[9] and Lister[10] measured particle size distribution of blowing snow. The size distribution given by Budd, Dingle, and Radok[5] was used in this study since the data were taken at specific heights and published with finer resolution than the others. Figure 2 shows the size distribution that was measured at 2 m above the surface. This distribution was used as the initial size distribution. Then size distributions for various altitudes up to 200 m were determined by applying equations (3) and (4) to each size interval. Resulting mass densities for $u_* = 1.0$ m/s are shown in figure 3.

VISIBILITY IN BLOWING SNOW

Liljequist[11] studied the relationship between visibility and mass density and found that visibility theoretically should be inversely proportional to mass density at eye level. This finding was confirmed by the measurements made by Budd, Dingle, and Radok[5] who found that

$$V = 0.1/\kappa \ , \qquad (5)$$

where

V = visibility (kilometers), and

κ = mass density (grams per cubic meter).

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[5]W. F. Budd, W. R. J. Dingle, and U. Radok, 1966, "The Byrd Snow Drift Project: Outline and Basic Results," Studies in Antarctic Meteorology, American Geophysical Union, Washington, DC

[9]A. K. Dyunin, "The Structure of Storm Snow and the Laws of Snow Transport," Voprosy ispol'zovaniya snega, Institut Geografii Akademii Nauk SSSR, 106-119.

[10]H. Lister, 1960, "Glaciology 1 Solid Precipitation and Drift Snow," T.A.E. Scientific Report No. 5, Trans-Antarctic Expedition Committee, London

[11]G. Liljequist, 1957, "Energy Exchange of an Antarctic Snow Field," Norwegian - British - Swedish Antarctic Expeditions 1949-1952, Sci Res, 2, part 1c

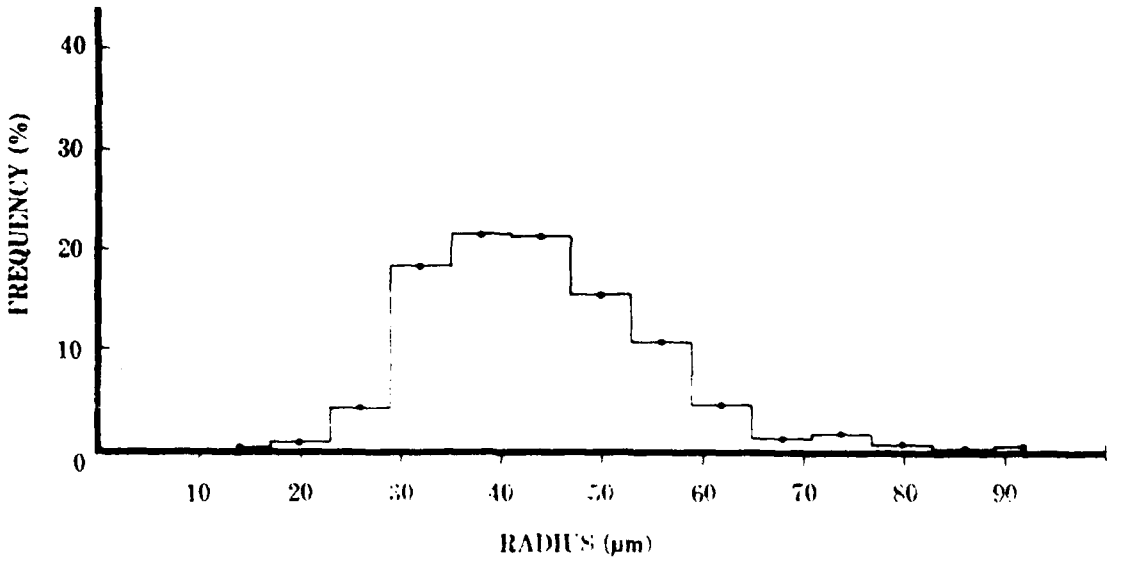

Figure 2. Particle size distribution of blowing snow at 2 meters above the surface (after Budd, Dingle, and Radok, 1966).

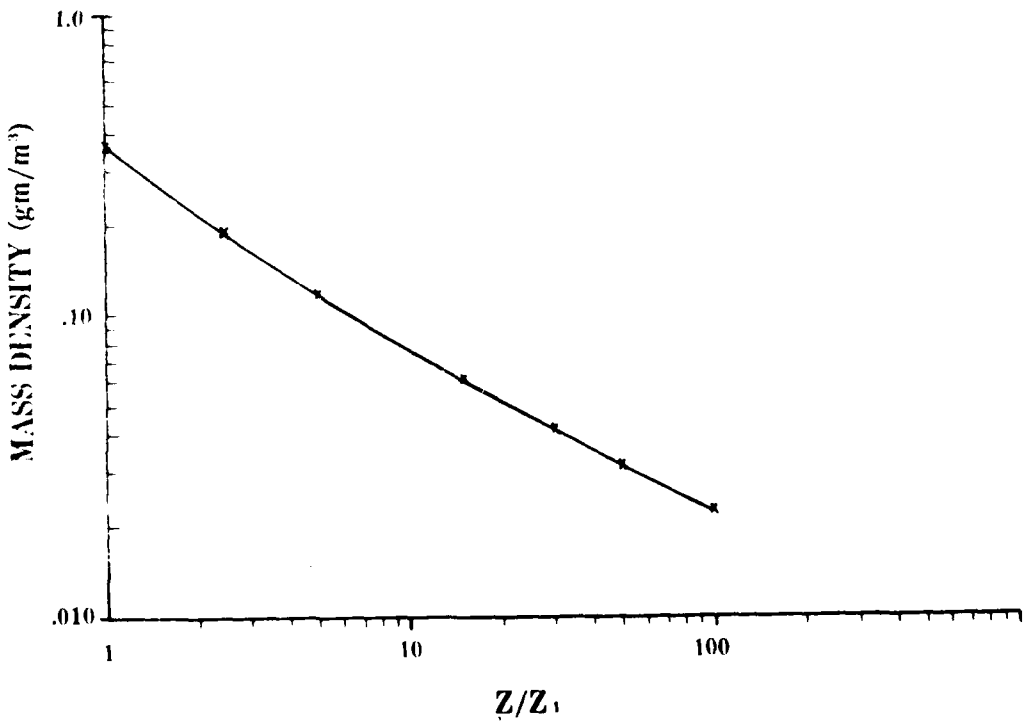

Figure 3. Variation of mass density with $\frac{Z}{Z_1}$ where Z is height above the surface and Z_1 is the reference height, Z_1 = 2 meters.

Empirical relationships between visibility and windspeed were derived by Lister[10] who found at Shackleton in the Antarctic that for visibility less than 1 km:

$$\log V = 1.27825 + 0.31104U - 0.04645U^2 + 0.00117U^3 \text{ (summer)} \quad (6a)$$

$$\log V = 1.55791 + 0.08083U - 0.02585U^2 + 0.00073U^3 \text{ (winter)} \quad (6b)$$

where U = windspeed (meters/second) at 10 m height and log is to the base 10.

The Koschmieder relation

$$\beta_{vis} = \frac{3.912}{V} , \quad (7)$$

where β_{vis} = visible extinction coefficient, may then be used to derive empirical relationships between the extinction coefficient and windspeed:

$$\log \beta_{vis} = -0.68585 - 0.31104U + 0.04645U^2 - 0.00117U^3 \text{ (summer)} \quad (8a)$$

$$\log \beta_{vis} = -0.96551 - 0.08083U + 0.02585U^2 - 0.00073U^3 \text{ (winter)} \quad (8b)$$

These relationships are for a specific location and period of time but should provide an indication of the effects of windspeed on visibility and extinction coefficient at extremely cold temperatures (winter) and somewhat warmer temperatures (summer).

VISIBLE AND INFRARED EXTINCTION

The action of the wind tends to round the windblown snow particles into more nearly spherical shapes than most snow crystals have initially. Because of this action, it is assumed here that Mie calculations for spherical particles would give valid indications of the optical properties of blowing snow. However, this assumption has not been verified with measurements.

Mie calculations confirmed that for the size distribution discussed above single scattering extinction is independent of wavelength and the geometrical optics approximation holds; that is,

$$\beta = 2\int \pi r^2 N(r) dr , \quad (9)$$

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[10]H. Lister, 1960, "Glaciology 1 Solid Precipitation and Drift Snow," T.A.E. Scientific Report No. 5, Trans-Antarctic Expedition Committee, London

where

β = extinction coefficient

$N(r)$ = particles per unit volume per unit size range

r = particle radius

The size distributions found by using equation (3) were used to calculate the variation of β with height for $u_* = 1.0$ m/s. These results were then used to derive an empirical relationship between β and Z; thus,

$$\beta = \beta_1 \left(\frac{Z}{Z_1}\right)^{-\frac{0.596}{u_*}} . \qquad (10)$$

Where β_1 is the extinction coefficient at Z_1.

The variation of β with u_* found by using equation (10) is shown in figure 4.

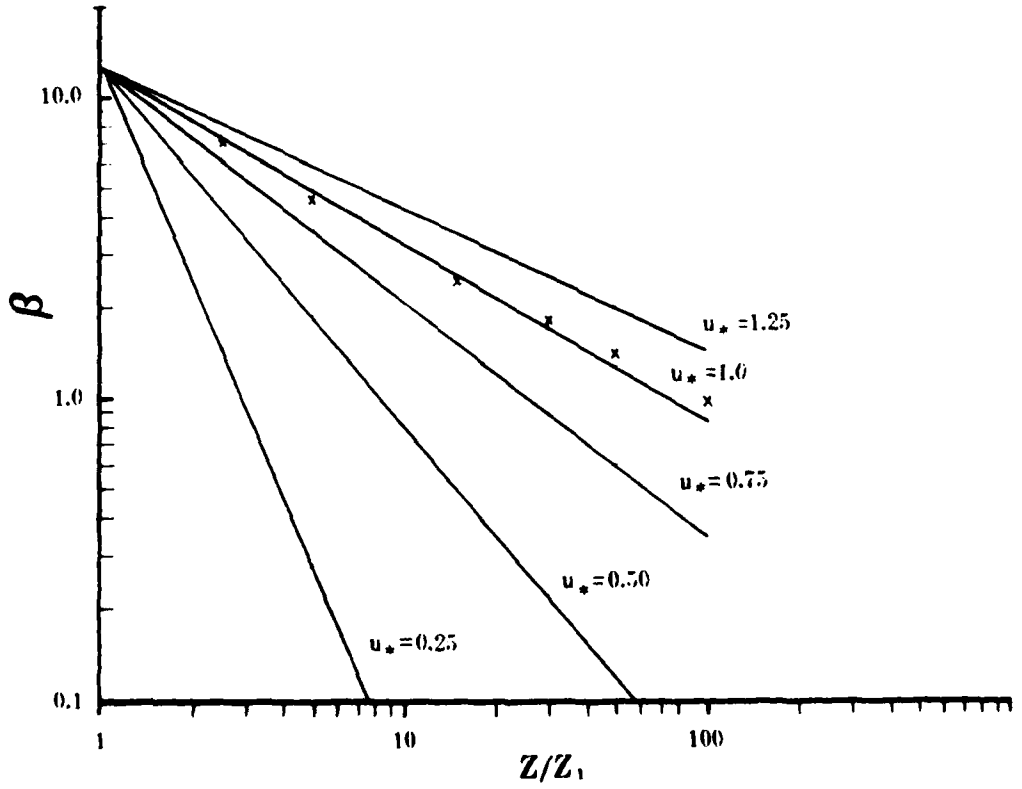

Figure 4. Variation of extinction coefficient with $\frac{Z}{Z_1}$ for various values of friction velocity u_*. X's indicate computed values used in deriving the relationship $\beta = \beta_1 \left(\frac{Z}{Z_1}\right)^{-\frac{0.596}{u_*}}$.

Extinction measurements in falling snow have shown that extinction is not entirely wavelength independent but increases slightly with increasing wavelength. Blowing snow is expected to show a similar effect, which is probably due to increased forward scattering at shorter wavelengths.

NEAR-MILLIMETER WAVE EXTINCTION IN BLOWING SNOW

For wavelengths, λ, in the near-millimeter wave region, the Rayleigh scattering approximation may be used to compute the extinction coefficient, β, when λ is greater than 4 mm. The resulting errors are less than 10 percent. For $\lambda < 4$ mm, the Rayleigh approximation yields extinction coefficients which are lower than those resulting from Mie calculations; and this difference is an order of magnitude for $\lambda = 0.86$ mm. The Rayleigh approximation for the extinction coefficient is given by

$$\beta = \frac{2\pi\kappa}{\lambda\rho} \text{ Im} \left(- \frac{m^2 - 1}{m^2 + 2} \right), \tag{11}$$

where $m = n_r - in_i$ = complex index of refraction and Im indicates the imaginary part of the quantity following.

Ray[12] discusses methods of computing the complex index of refraction for ice in the near-millimeter wave region. He found that the real part, n_r, is 1.78 in the near-millimeter wave region for normal atmospheric temperatures. The imaginary part varies with wavelength and temperature, increasing with either increasing temperature or wavelength. Since for ice

$n_r >> n_i$, $\text{Im} \left(- \frac{m^2 - 1}{m^2 + 2} \right)$ is approximately linearly proportional to n_i; then from equation (11), β is approximately linearly proportional to n_i. For example, since n_i at 0°C (0.0012 at $\lambda = 1$ cm) is about five times n_i at -20°C (0.00024 at $\lambda = 1$ cm), β at 0°C is about five times that at -20°C.

The relationship between the near-millimeter wave extinction coefficient and visibility may be found by combining equations (5) and (11) to give

$$\beta = \frac{2.09}{\lambda V} \text{ Im} \left(- \frac{m^2 - 1}{m^2 + 2} \right), \tag{12}$$

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[12] P. S. Ray, 1972, "Broadband Complex Refractive Indices of Ice and Water," Applied Optics, 11:1836-1844

where λ is in units of millimeters, V is in units of kilometers, and β is in units of kilometers^{-1}. Also, the Koschmieder relation may be used to relate β to the visible extinction coefficient, β_{vis}. Thus,

$$\beta = \frac{53.5\ \beta_{vis}}{\lambda}\ \mathrm{Im}\left(-\frac{m^2 - 1}{m^2 + 2}\right), \tag{13}$$

where λ is in units of millimeters and β and β_{vis} are in units of kilometers^{-1}. Equations (6) and (12) or (8) and (13) may be used to find the relationship between the extinction coefficient and the windspeed:

$$\beta = \frac{53.5}{\lambda}\ \mathrm{Im}\left(-\frac{m^2 - 1}{m^2 + 2}\right) 10^{(-0.68585 - 0.31104U + 0.04645U^2 - 0.00117U^3)} \quad \text{(summer)} \tag{14a}$$

$$\beta = \frac{53.5}{\lambda}\ \mathrm{Im}\left(-\frac{m^2 - 1}{m^2 + 2}\right) 10^{(-0.96551 - 0.08083U + 0.2585U^2 - 0.00073U^3)} \quad \text{(winter)} \tag{14b}$$

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