INFORMATION TO USERS

This manuscript has been reproduced from the microfilm master. UMI films the text directly from the original or copy submitted. Thus, some thesis and dissertation copies are in typewriter face, while others may be from any type of computer printer.

The quality of this reproduction is dependent upon the quality of the copy submitted. Broken or indistinct print, colored or poor quality illustrations and photographs, print bleedthrough, substandard margins, and improper alignment can adversely affect reproduction.

In the unlikely event that the author did not send UMI a complete manuscript and there are missing pages, these will be noted. Also, if unauthorized copyright material had to be removed, a note will indicate the deletion.

Oversize materials (e.g., maps, drawings, charts) are reproduced by sectioning the original, beginning at the upper left-hand corner and continuing from left to right in equal sections with small overlaps. Each original is also photographed in one exposure and is included in reduced form at the back of the book.

Photographs included in the original manuscript have been reproduced xerographically in this copy. Higher quality 6" x 9" black and white photographic prints are available for any photographs or illustrations appearing in this copy for an additional charge. Contact UMI directly to order.

U·M·I

University Microfilms International
A Bell & Howell Information Company
300 North Zeeb Road, Ann Arbor, MI 48106-1346 USA
313/761-4700 800/521-0600

Order Number 9209081

**Remote sensing the susceptibility of cloud albedo to changes in
drop concentration**

Platnick, Steven Edward, Ph.D.

The University of Arizona, 1991

U·M·I

**300 N. Zeeb Rd.
Ann Arbor, MI 48106**

REMOTE SENSING THE SUSCEPTIBILITY OF CLOUD ALBEDO
TO CHANGES IN DROP CONCENTRATION

by

Steven Edward Platnick

A Dissertation Submitted to the Faculty of the
DEPARTMENT OF ATMOSPHERIC SCIENCES

In Partial Fulfillment of the Requirements
For the Degree of

DOCTOR OF PHILOSOPHY

In the Graduate College

THE UNIVERSITY OF ARIZONA

1 9 9 1

THE UNIVERSITY OF ARIZONA
GRADUATE COLLEGE

2

As members of the Final Examination Committee, we certify that we have read
the dissertation prepared by Steven Edward Platnick
entitled Remote Sensing the Susceptibility of Cloud Albedo

to Changes in Drop Concentration

and recommend that it be accepted as fulfilling the dissertation requirement
for the Degree of Doctor of Philosophy.

Benjamin M. Herman
Benjamin M. Herman

9/5/91

Date

Eric A. Betterton
Eric A. Betterton

9/5/91

Date

E. Philip Krider
E. Philip Krider

9/5/91

Date

John Reagan
John Reagan

9/5/91

Date

Phillip N. Slater
Phillip N. Slater

9/5/91

Date

Final approval and acceptance of this dissertation is contingent upon the
candidate's submission of the final copy of the dissertation to the Graduate
College.

I hereby certify that I have read this dissertation prepared under my
direction and recommend that it be accepted as fulfilling the dissertation
requirement.

Benjamin M. Herman
Dissertation Director Benjamin M. Herman

9/5/91

Date

STATEMENT BY AUTHOR

This dissertation has been submitted in partial fulfillment of requirements for an advanced degree at The University of Arizona and is deposited in the University Library to be made available to borrowers under rules of the Library.

Brief quotations from this dissertation are allowable without special permission, provided that accurate acknowledgement of source is made. Requests for permission for extended quotation from or reproduction of this manuscript in whole or in part may be granted by the head of the major department or the Dean of the Graduate College when in his or her judgment the proposed use of the material is in the interests of scholarship. In all other instances, however, permission must be obtained from the author.

SIGNED:

ACKNOWLEDGEMENTS

I wish to express grateful appreciation to my mentor, Sean Twomey, for his advice, guidance, and support during the three years of this work. And also for the many stimulating and enjoyable discussions I was able to share with him regarding the nature of the atmosphere as well as the world at large. Cheers.

I wish to especially thank professor Ben Herman for his careful reading of this manuscript and his help in its preparation, and to professors Eric Betterton and E. Philip Krider for their generous time.

My fellow colleagues in the Department of Atmospheric Sciences have been of continual help to me during these years. I wish to especially thank David Flittner for allowing me to use him as a sounding board during the many months I was without mentor and for his help in obtaining AVHRR calibration data. Special thanks to Mike Leuthold for his tireless computer help and to Joel Mozer, Ali Shambayat, and Peter Pilewskie.

Finally, I wish to thank Alvin, Betty, Michael, Karen, and Sadie for a lifetime of unwavering support and encouragement, and all the loved-ones and friends who helped make this possible.

TABLE OF CONTENTS

LIST OF ILLUSTRATIONS	8
LIST OF TABLES	14
ABSTRACT	16
1. INTRODUCTION	17
2. CLOUD SUSCEPTIBILITY	20
2.1 A Brief Introduction to Cloud Radiative Transfer	20
2.1.1 Cloud Physical Parameters	21
2.1.2 Optical Parameters	24
2.1.3 Cloud Albedo	29
2.2 Susceptibility	32
2.2.1 Broadband Albedo	37
2.2.2 Ship Tracks	39
2.2.3 Global Albedo	40
2.2.4 The Influence of Liquid Water Content	41
2.2.5 Other Effects of Pollution on Albedo	43
2.3 The Remote Sensing of Susceptibility	45
2.4 Marine Stratus Measurements	51
3. THE AVHRR INSTRUMENT	62
3.1 Overview of Available Satellites and Instruments	62
3.1.1 The NOAA Polar Orbiter	63
3.1.2 Landsat	65
3.1.3 GOES	66
3.2 AVHRR Data	68
3.2.1 AVHRR Channel Spectral Response	68
3.2.2 AVHRR Data Formats	70

TABLE OF CONTENTS - Continued

3.2.3	AVHRR Channel Counts and Physical Quantities	72
3.2.4	Digitization Error	76
3.3	Channel Calibration	79
3.4	Geometric Considerations	85
3.4.1	Satellite Zenith Angle	86
3.4.2	Relative Azimuth Angle	87
3.4.3	Sensitivities of μ , ϕ , to the Geometry	88
3.4.4	Ground Resolution	89
4.	RADIATIVE TRANSFER	102
4.1	Mie Calculations for the AVHRR Channels	102
4.1.1	The Effect of Drop Size Distribution	103
4.1.2	Monochromatic Examples	107
4.1.3	Averages Over Wavelength	113
4.2	The Doubling Code	115
4.2.1	Azimuthally Independent Formulation	116
4.2.2	Azimuthally Dependent Formulation	120
4.2.3	Computational Checks	122
4.2.4	Emission	124
4.3	The Library Files	125
5.	ATMOSPHERIC EFFECTS ON REMOTE SENSING	143
5.1	LOWTRAN Radiative Transfer Code	145
5.2	Atmospheric Transmission	151
5.3	Atmospheric Scattering	155
5.3.1	Gauss-Seidel Calculations	158
5.3.2	Comparisons with HRPT Clear Sky Data	164

TABLE OF CONTENTS - Continued

5.4	Atmospheric Emission	168
5.5	Summary of AVHRR Atmospheric Corrections	174
5.7	Uncertainty Analysis	176
6.	THE REMOTE SENSING ALGORITHM AND SOME RESULTS FOR MARINE STRATUS CLOUDS	208
6.1	The Remote Sensing Algorithm	212
6.1.1	Sensitivity to Algorithm Parameters	215
6.1.2	Error Calculations Based on Chapter 5 Results	221
6.2	Comparison with In Situ Measurements	222
6.3	Marine Stratus and Ship Tracks	228
6.3.1	Marine Stratus Near the West Coast of the United States	229
6.3.2	Some Observations on Ship Tracks	231
6.3.3	Marine Stratus Near Southern Africa	236
6.4	California Valley Fog	239
6.5	Conclusions	241
	LIST OF REFERENCES	256

LIST OF ILLUSTRATIONS

Figure	Page
2.1 Schematic representation for a cloud	56
2.2a Cloud albedo for AVHRR channel 1	57
2.2b Cloud albedo for NIR atmospheric window wavelengths at a drop radius of 10 μm	57
2.3 Two stream approximation for the part of Equation 2.10	58
2.4a Surface of susceptibility versus radius and optical thickness using the two stream approximation	59
2.4b Surface of susceptibility versus radius and albedo using the two stream approximation	60
2.5 The effect on albedo of changing cloud drop concentration by a factor amount	61
3.1a Relative spectral response of AVHRR channels 1 and 2	93
3.1b Relative spectral response of AVHRR channel 3	93
3.1c Relative spectral response of AVHRR channels 4 and 5	94
3.2 Temperature error for AVHRR channel 4 due to digitization error in detected intensity	95
3.3a In-flight calibration coefficients for NOAA-11 channels 1	96
3.3b In-flight calibration coefficients for NOAA-11 channels 2	96
3.4 Definition of relative azimuth angle between the satellite and the sun	97
3.5 Geometry used in calculating the satellite zenith angle	98
3.6 Geometry used in defining the absolute solar and satellite azimuth angles	99

LIST OF ILLUSTRATIONS - Continued

Figure	Page
3.7 Geometry used in calculating the absolute satellite azimuth angle . . .	100
3.8 Ratio of linear IFOV for a pixel compared with that for a pixel at nadir	101
4.1a Asymmetry parameter versus mean radius for AVHRR channel 1 . .	131
4.1b Asymmetry parameter versus mean radius for AVHRR channel 3 . .	131
4.2a Absorption versus mean radius for AVHRR channel 3	132
4.2b Absorption versus mean radius for AVHRR channel 4	132
4.3a Extinction efficiency versus mean radius for AVHRR channel 1 . . .	133
4.3b Extinction efficiency versus mean radius for AVHRR channel 3	133
4.3c Extinction efficiency versus mean radius for AVHRR channel 4	133
4.4 Vapor absorption coefficient for AVHRR channel 3	134
4.5a Spectral average of asymmetry parameter versus mean radius for AVHRR channel 1	135
4.5b Spectral average of asymmetry parameter versus mean radius for AVHRR channel 3	135
4.5c Spectral average of asymmetry parameter versus mean radius for AVHRR channel 4	135
4.6a Spectral average of single scattering albedo versus mean radius for AVHRR channel	136
4.6b Spectral average of single scattering albedo versus mean radius for AVHRR channel 4	136
4.7 Schematic for fields involved in doubling calculations	137
4.8 Components to emission from cloud and surface	138

LIST OF ILLUSTRATIONS - Continued

Figure	Page
4.9a Azimuthally averaged reflectance for AVHRR channel 1	139
4.9b Azimuthally averaged reflectance for AVHRR channel 3	139
4.10a Azimuthally dependant reflectance for AVHRR channel 1	140
4.10b Azimuthally dependant reflectance for AVHRR channel 3	140
4.11 Ratio of reflected solar intensity to thermal emission for AVHRR channel 3	141
4.12 Scene temperature inferred with AVHRR channel 4 versus mean radius	142
5.1 Schematic for the effect of the atmosphere on the satellite signal . . .	185
5.2a Spectral transmission in AVHRR channel 1	186
5.2b Spectral transmission in AVHRR channel 2	186
5.2c Spectral transmission in AVHRR channel 3	187
5.2d Spectral transmission in AVHRR channel 4	187
5.2e Spectral transmission in AVHRR channel 5	188
5.3a Atmospheric transmission in AVHRR channel 1 from LOWTRAN . .	189
5.3b Atmospheric transmission in AVHRR channel 2 from LOWTRAN . .	189
5.3c Atmospheric transmission in AVHRR channel 3 from LOWTRAN . .	190
5.3d Atmospheric transmission in AVHRR channel 4 from LOWTRAN . .	190
5.3e Atmospheric transmission in AVHRR channel 5 from LOWTRAN . .	191
5.4a Intensity of atmospheric scattering in AVHRR channel 1 from LOWTRAN	192

LIST OF ILLUSTRATIONS - Continued

Figure	Page
5.4b Intensity of atmospheric scattering in AVHRR channel 2 from LOWTRAN	192
5.4c Intensity of atmospheric scattering in AVHRR channel 3 from LOWTRAN	193
5.5 Azimuthal reflectance using Gauss-Seidel code	194
5.6a Measure reflectance for clear sky over ocean for AVHRR channel 1 .	195
5.6b Measure reflectance for clear sky over ocean for AVHRR channel 2 .	195
5.7a Intensity of atmospheric emission in AVHRR channel 3 from LOWTRAN	196
5.7b Intensity of atmospheric emission in AVHRR channel 4 from LOWTRAN	196
5.7c Intensity of atmospheric emission in AVHRR channel 5 from LOWTRAN	197
5.8 Uncertainty in atmospheric transmission versus maximum error in reflectance for AVHRR channel 1	198
5.9 Uncertainty in optical thickness versus maximum error in reflectance for AVHRR channel 1	198
5.10 Uncertainty in atmospheric scattering versus maximum error in reflectance for AVHRR channel 1	199
5.11a Uncertainty in atmospheric transmission versus maximum error in reflectance for AVHRR channel 3, $T_s=280$ K	200
5.11b Uncertainty in atmospheric transmission versus maximum error in reflectance for AVHRR channel 3, $T_s=290$ K	200
5.12a Uncertainty in optical thickness versus maximum error in reflectance for AVHRR channel 3, $T_s=280$ K	201

LIST OF ILLUSTRATIONS - Continued

Figure	Page
5.12b Uncertainty in optical thickness versus maximum error in reflectance for AVHRR channel 3, $T_s=290$ K	201
5.13 Uncertainty in atmospheric scattering and emission versus maximum error in reflectance for AVHRR channel 3	202
5.14 Uncertainty in atmospheric transmission versus maximum error in temperature for AVHRR channel 4	203
5.15 Uncertainty in atmospheric emission versus maximum error in temperature for AVHRR channel 4	204
5.16a Range in error in retrieved reflectance based on the uncertainty in transmission and scattering for AVHRR channel 1	205
5.16b Range in error in retrieved reflectance based on the uncertainty in transmission and scattering for AVHRR channel 2	205
5.16c Range in error in retrieved reflectance based on the uncertainty in transmission, scattering, and emission for AVHRR channel 3, $T_s=290$ K	206
5.16d Range in error in retrieved reflectance based on the uncertainty in transmission, scattering, and emission for AVHRR channel 3, $T_s=280$ K	206
5.16e Range in error in retrieved temperature based on the uncertainty in transmission and emission for AVHRR channel 4	207
6.1 Schematic diagram of the forward problem	243
6.2 Schematic diagram of the inverse problem	244
6.3a Isolines of residual error with $W_1=1$ and $W_3=2$ for thin stratus . . .	245
6.3b Isolines of residual error with $W_1=1$ and $W_3=2$ for moderately thick stratus	245
6.4a Isolines of residual error with $W_1=1$ and $W_3=4$ for thin stratus . . .	246

LIST OF ILLUSTRATIONS - Continued

Figure	Page
6.4b Isolines of residual error with $W_1=1$ and $W_3=4$ for moderately thick stratus	246
6.5 NOAA-9 LAC, 30 June 1987, AVHRR channel 1	247
6.6 Retrieved radius and optical thickness for the image of Figure 6.5 . .	248
6.7 NOAA-10 HRPT, 10 July 1987, AVHRR channel 1	249
6.8a Retrieved radius and optical thickness for the image of Figure 6.7 for a northern section	250
6.8b Retrieved radius and optical thickness for the image of Figure 6.7 for a southern section	250
6.9 NOAA-11 HRPT, 20 March 1989, AVHRR channel 1	251
6.10 NOAA-11 HRPT, 2 March 1990, AVHRR channel 1	252
6.11 NOAA-11 GAC, 4 January 1989, AVHRR channel 1 and 3	253
6.12 NOAA-11 GAC, 22 January 1989, AVHRR channel 1 and 3	254
6.13a Retrieved radius and optical thickness for California valley for on 20 December, 1989	255
6.13b Retrieved radius and optical thickness for California valley for on 11 January, 1990	255

LIST OF TABLES

Table	Page
2.1 Range of single scattering albedos for NIR atmospheric windows . . .	48
2.2 Range of critical optical thicknesses using the Van de Hulst approximation for NIR atmospheric windows	49
2.3 Summary of some microphysical measurements made in maritime clouds	55
3.1 Characteristics of the NOAA TIROS-N series satellite and the AVHRR instrument	69
3.2 Spectral half-power bandwidths for the AVHRR channels	71
3.3 AVHRR channels 1 and 2 calibration coefficients used in this study .	83
3.4 Sensitivity of satellite geometry to earth location and satellite height	90
4.1a Optical parameters for AVHRR channel 3 for three size distributions	108
4.1b Optical parameters for AVHRR channel 1 for three size distributions	109
4.2 Cloud drop size distributions used for the analysis of Table 4.1	110
4.3 The effective single scattering albedo for AVHRR channel 3 including water vapor absorption	113
4.4 Wavelength and channel response for AVHRR channels 1, 2, 3, and 4 used for band averaged calculations	115
5.1a Midlatitude summer model atmosphere used by LOWTRAN	149
5.1b Midlatitude winter model atmosphere used by LOWTRAN	150
5.2 Transmittances in the AVHRR channels for LOWTRAN and the line-by-line calculation of Saunders and Edwards	154

LIST OF TABLES

Table	Page
5.3 Species optical thicknesses used in Gauss-Seidel calculations	160
5.4a Gauss-Seidel calculations for a path from sea level to space	162
5.4b Gauss-Seidel calculations for a path from a 2 km level to space	163
5.5 The maritime model atmosphere used by LOWTRAN for calculating emission in AVHRR channels 3 and 4	171
5.6 Summary of atmospheric effects obtained from LOWTRAN	175
6.1a Sensitivity of retrieved parameters to varying surface temperature, surface albedo, atmospheric corrections, and channel 1 calibration for thin stratus	218
6.1b Sensitivity of retrieved parameters to varying atmospheric corrections and channel 1 calibration for moderate stratus	219
6.1c Sensitivity of retrieved parameters to varying atmospheric corrections and channel 1 calibration for thick stratus	219
6.2 Comparison of NOAA-10 retrieved parameters with the measurements of Radke et al. (1989)	228
6.3a Retrieved parameters for out-of-track regions for NOAA-11, HRPT, 2 March 1990	234
6.3b Retrieved parameters for in-track regions for NOAA-11, HRPT, 2 March 1990	235

ABSTRACT

The role of clouds in reflecting solar radiation to space and thereby reducing surface heating is of critical importance to climate. Combustion processes that produce greenhouse gases also increase cloud condensation nuclei (CCN) concentrations which in turn increase cloud drop concentrations and thereby cloud albedo.

A calculation of *cloud susceptibility*, defined in this work as the increase in albedo resulting from the addition of one cloud drop per cubic centimeter (as cloud liquid water content remains constant), is made through satellite remote sensing of cloud drop radius and optical thickness. The remote technique uses spectral channels of the Advanced Very High Resolution Radiometer (AVHRR) instrument on board the NOAA polar orbiting satellites.

Radiative transfer calculations of reflectance and effective surface and cloud emissivities are made for applicable sun and satellite viewing angles, including azimuth, at various radii and optical thicknesses for each AVHRR channel. Emission in channel 3 (at $3.75\ \mu\text{m}$) is removed to give the reflected solar component. These calculations are used to infer the radius and optical thickness giving the best match to the satellite measurements. The effect of the atmosphere on the signal received by the satellite is included in the analysis.

Marine stratus clouds are a focus of this work. As well as being important modifiers to climate, they are cleaner than continental clouds and so likely to be of higher susceptibility. Analysis of several stratus scenes, including some containing ship tracks, supports this expectation.

CHAPTER 1

INTRODUCTION

With much attention being given to CO₂ and other greenhouse gases, it is important to appreciate that water is the most critical radiative species in the atmosphere, both as a vapor and as liquid or ice in clouds. Clouds have been recognized as the most significant modulators to radiative processes in the atmosphere. Reflection of solar radiation in the shortwave reduces the energy absorbed at the surface and in the lower atmosphere. In the longwave, clouds reduce the effective radiating temperature of the atmosphere and so less infrared energy is radiated to space. The balance between these two competing effects has been a cause for much study. Arking (1991) presents a review of various methods used to determine the overall radiative effect of clouds compared with a clear sky earth. His calculations, based on satellite observations reported by various researchers, all indicate that on a global scale clouds have a cooling influence. The magnitude of the so called *cloud forcing* varies by a factor of two between the studies.

The *cloud sensitivity* (Arking 1991), defined as the change in energy absorbed by the climate system to changes in a cloud parameter, is meaningful for climate change. Important parameters include the macroscopic (such as cloud amount, or cloud cover fraction, and cloud height and thickness) and the microscopic (cloud liquid water content, drop size, and phase). Recognition that cloud parameter changes might follow

climate changes introduces a feedback between the two which could be either positive or negative. Present models only seem to underscore the critical nature of the cloud problem. A parameterization which links cloud liquid water content with increases in global temperatures (Summerville and Remer 1984) indicates a negative feedback. Arking's review of 19 global circulation models (GCMs) based on a study by Cess et al. (1990) shows both positive and negative feedback for cloud amount. An overview of cloud parameterization concerns in GCMs is given by Fouquart et al. (1990) and by Arking.

In typical GCMs, only cloud amount and height are allowed to respond to climate changes, while cloud optical properties are fixed. An important cloud microphysical parameter, not incorporated into GCMs, is drop size. It will be shown in chapter 2 that cloud reflectance is partially dependent on drop size which is in turn linked with cloud condensation nuclei (CCN) concentrations, a subset of the atmospheric aerosol, present during cloud development (Twomey 1974). CCN concentrations are variable, having both natural and anthropogenic sources. Among the anthropogenic sources are combustion processes that also release CO_2 , a major greenhouse gas (excluding water vapor). The overall effect of increasing CCN is to increase cloud brightness which results in cooling. There is no compensating outcome in the infrared. In light of these concerns, it is useful to define a quantity representing the sensitivity of cloud reflectance to changes in CCN concentration. This quantity is referred to as *cloud susceptibility* and is discussed in chapter 2. Unlike cloud amount, modification of cloud reflectance by CCN does not immediately constitute a feedback. That is, the effect does

not depend on the development of any actual climate change and can occur quite free from any such controversy.

As expected, not all clouds are equally susceptible; the determining factors turn out to be cloud optical thickness and drop size which can be inferred through solar reflection measurements at absorbing and non-absorbing wavelengths. In situ studies are of course possible but generally more costly and difficult to conduct over large spacial and temporal scales. Since a global understanding of susceptibility is ultimately desired, a satellite remote sensing scheme has been developed. Chapter 3 discusses the instrument used in this study, the Advanced Very High Resolution Radiometer (AVHRR) aboard the NOAA TIROS-N polar satellites. The radiative transfer theory necessary for inferring optical thickness and drop size from reflection measurements is presented in chapter 4. The atmosphere, primarily above the cloud, will modify the observed reflectances at the satellite sensor and should be accounted for when inferring cloud properties. Approximations for the atmospheric effect in the AVHRR channels is given in chapter 5. This study primarily investigates the susceptibility of maritime stratus clouds. Results for a number of these stratus scenes are presented in chapter 6.

CHAPTER 2

CLOUD SUSCEPTIBILITY

Reflection of solar radiation by clouds is a vital and important factor in climate. This chapter studies the sensitivity of cloud albedo to changes in drop concentration (almost linearly related to CCN concentration), termed *susceptibility*. A practical understanding of cloud radiative transfer as applied to reflectance is needed at this point and is presented in the first section. This will be achieved without the detailed results of Mie theory and radiative transfer theory which is deferred until chapter 4. After susceptibility is defined and developed, the remote sensing of this parameter is examined, and found to depend on the ability to deduce cloud optical thickness and drop radius. For a variety of reasons, to be addressed later, this study will concentrate on remote sensing the susceptibility of marine stratus clouds. The chapter concludes with a summary of microphysical studies done on these type of clouds.

2.1 A Brief Introduction to the Basics of Cloud Radiative Transfer

A cloud can be thought of approximately as shown in the schematic illustration of Figure 2.1. Consider the cloud to be horizontally homogeneous and plane parallel, and if not vertically homogeneous than at least capable of being represented by some vertical weighting of the necessary descriptive parameters. These assumptions can only

be valid over some appropriate scale. Synoptic and mesoscale meteorological conditions and even local influences effect the larger scales while the smallest scales are governed by turbulence. The effects of inhomogeneities on cloud radiative properties will be discussed in chapter 6. Solar radiation shown incident at the top of the cloud is effectively plane parallel (the solar disk as seen from earth spans less than a half a degree of arc length). The cosine of the solar zenith angle, μ_0 , is the fundamental geometrical variable, not the zenith angle itself. The interaction of solar radiation with the cloud drops gives rise to diffuse reflection and transmission, not necessarily isotropic as implied in the figure. Also shown in the diagram is directly transmitted radiation which passes through the cloud without being effected by the drops, and reflection off the underlying surface, eg. ocean or land, which can modify the reflection as seen from cloud top. The cloud drops are assumed to be liquid water in this study.

2.1.1 Cloud Physical Parameters

The ultimate fate for a given wavelength of the incident radiation is dependent on three *physical* cloud parameters (r , N , and h) as indicated in the figure. The cloud drop density, N , is examined first. Its value has been found to range from tens per cm^3 for very clean air to thousands per cm^3 for continental or polluted air. Each cloud drop originally forms on a sub-micron nucleus (thought to typically be a few hundredths of a micron in radius) which because of its solubility, or at least partial solubility, can be activated and grown at small vapor supersaturation, normally a few tenths of a percent (see Twomey and Wojciechowski 1969 for a global summary). The supersaturation

necessary for activation (i.e., unstable growth) for a purely soluble particle is dependent on the particle mass. The larger the mass the greater the quantity of soluble material dissolved in the drop. And according to Raoult's law, the greater the dissolved solute the lower the vapor pressure necessary for equilibrium. This influence acts against the opposite effect of the drop's curvature which raises the vapor pressure necessary for drop equilibrium. The competition between the two gives rise to the well known Köhler curve which determines, for a given CCN mass and solute dissociation factor, a critical supersaturation above which activation begins; the larger the CCN the smaller the critical supersaturation. The origin of CCN are thought to be primarily sulfates formed by gas to particle conversion (first inferred by Dinger et al. 1970; Hobbs 1971). The final number of cloud drops per unit volume, N , is approximately proportional to the number density of CCN present during cloud formation, the primary variables being updraft velocity and CCN densities (Twomey 1959). If the cloud develops with large vertical velocities, CCN with larger critical supersaturations (smaller mass) will have a chance to be activated in the cooling parcel before vapor is depleted by drops activated at lower supersaturations (ie., higher temperatures). Consequently the final drop density would be larger than for the condition of small updrafts where initially activated CCN have time to grow and consume the available vapor before further cooling activates more CCN. The exact empirical relation between CCN density and critical supersaturation will also affect the final drop density. Calculations based on experimental data of CCN densities indicated that typically $N \propto \text{CCN}^{.8}$ for clean maritime air (Twomey and Wojciechowski 1969). Direct measurements of both CCN

and N by Twomey and Warner (1967) gave an approximate linear fit.

Continental air is found to have larger numbers of CCN than clean oceanic air masses. For critical supersaturations below 1%, the difference in CCN densities for the two air masses can easily be greater than an order of magnitude (Twomey and Wojciechowki 1969). Natural sources of sulfur needed for CCN production include volcanic and biological activity. Over the oceans dimethylsulphide (DMS) excreted by phytoplankton is believed a significant contributor (Charlson et al. 1987). Combustion processes are also found to be a source for CCN. Squires (1966) extrapolated measurements of CCN originating from Denver, Colorado to infer that the ratio of anthropogenic to natural CCN concentrations in the United States from fossil fuel burning (at 1963 emission rates) would be about 14% but could approach 100% in heavy industrialized areas. Warner and Twomey (1967) measured large increases in CCN levels and concurrent number densities downwind from cane fires in maritime air off Queensland, Australia. Hobbs et al. (1980) made measurements of elevated CCN in power plant plumes and found order of magnitude increases of cloud drop numbers in clean marine stratus intercepted by the plume. Gorbinet et al. (1985) measured increases by factors of two to three in CCN concentrations near industrial activity.

The two remaining physical cloud parameters shown in figure 2.1 are drop radius and cloud thickness. Non-precipitating clouds generally show a narrow size distribution, $n(r)$, with a mean radius of a few microns to as high as 20 microns (see for example Warner 1969; Noonkester 1984). Drop radius can be linked with cloud drop number by the cloud liquid water content, defined as the mass of liquid per volume

of air,

$$\begin{aligned}
 W &= \frac{4}{3}\pi\rho_l \int r^3 n(r) dr \\
 &= \frac{4}{3}\pi\rho_l \overline{r^3} N .
 \end{aligned}
 \tag{2.1}$$

For narrow size distributions $\overline{r^3} \approx \overline{r}^3$. Though liquid water content in a cloud may vary considerably, spacial averages in marine clouds throughout the world have a typical value of .3 g/m³. Thicknesses for marine stratus clouds, normally capped by a temperature inversion (at about 950 to 850 mb), are around a few hundred to a thousand meters. Each of the physical parameters, and liquid water content, are highly variable on a small scale (e.g., Curry 1986) but it is generally assumed that each can be suitably averaged over some appropriate physical distance for radiative transfer calculations. Values for these parameters obtained from several measurement programs for marine clouds are summarized in section 2.4 at the end of this chapter.

2.1.2 Optical Parameters

Through a *physical* description it is possible to predict the radiative properties of the cloud, however the fundamental basis of light interaction with the drops is more easily couched in terms of *optical* parameters. Consider first the interaction of radiation with a single drop. Incident energy will be scattered by the drop and redirected into other directions. Depending on the wavelength, some of that energy may also be absorbed by the liquid within the drop. The ratio of the particle size to the wavelength

of radiation is a fundamental variable and normally expressed as the *size parameter*, x , which is defined as 2π times the above ratio. For a cloud drop with $10\ \mu\text{m}$ radius and for visible light, say $.6\ \mu\text{m}$, $x \approx 105$. The description of the interaction for spherical particles whose size parameters are comparable with or larger than unity is known as Mie theory (see Van de Hulst 1981) and is applicable to the cloud drop problem. Rayleigh theory is the limiting case for size parameters much less than unity, such as for light scattering from molecules. The total radiation that is redirected (scattered) or absorbed by the drop is considered as extinction or attenuation of the incident beam and is expressed in terms of an effective cross section $\sigma_{\text{ext}} = Q_{\text{ext}} \pi r^2$ such that the total extinction is the product of σ_{ext} and the incident plane-parallel flux density (e.g., W/m^2 , also known in the literature as irradiance). Q_{ext} is the extinction efficiency which in the limit of large size parameter approaches a value of 2. Efficiencies can also be defined for scattering and absorption of radiation only, Q_{sca} and Q_{abs} respectively.

The necessary optical parameters are now presented. The ratio of scattered energy to total extinction, or $Q_{\text{sca}}/Q_{\text{ext}}$, is referred to as the *single scattering albedo* and for historical reasons (the ratio being the first coefficient in the Legendre function expansion of the phase function) is given the symbol ω_0 . The single scattering albedo is determined from the absorption of bulk liquid water and the drop radius. In the visible, $\omega_0 = 1$ (no absorption) for cloud drops and generally decreases (increasing absorption) as wavelength increases into the near infrared (NIR). The distribution of scattered energy from the drop is given by the *phase function* $p(\Theta)$ where Θ is the scattering angle, defined such that $\Theta = 0$ is the forward scattered direction and $\Theta = \pi$ is

the backscattered. Azimuthal symmetry exists about the axis defining these two directions. In this work the phase function will be normalized such that its integration over 4π solid angle is one, that is, scattered radiation must, by definition, be scattered somewhere. Other authors may normalize it to ω_0 in which case it represents the redistribution of attenuated energy. For $x > 2$ scattering in the forward direction is many orders of magnitude larger than in other directions although there is some peaking in the back direction as well (see for example plots shown by Twomey 1977). In general, the larger the size parameter the more detail (e.g., the number of lobes in the pattern is $\approx x$) found in the phase function and the greater the forward peak. In the Rayleigh limit the phase function is relatively isotropic having a peanut or dumbbell-like shape. The physics describing ω_0 and $p(\Theta)$, and the radiative transfer equation itself, is based on the wave theory of classical electromagnetics. However a photon interpretation often has an easier description and can be useful for calculations (e.g., Monte Carlo techniques). In this view, $1-\omega_0$ can be seen as the probability that a photon interacting with a drop will be absorbed. If it is not absorbed, then $p(\Theta)$ becomes the probability function for the new direction that the scattered photon will take.

The single scattering albedo and phase function describe the extinction of radiation from a single drop, i.e., single scattering. In the atmosphere however a collection of scatterers is present and distributed through space and multiple scattering can take place before a photon finds its way to a detector. This is the general case for scattering in clouds. A photon incident on a collection of drops "sees" an extinction optical path, τ_{ext} , defined as

$$\begin{aligned}
\tau_{ext} &= l \int Q_{ext} \pi r^2 n(r) dr \\
&= Q_{ext} \pi \overline{r^2} N l \\
&= k_{ext} l .
\end{aligned}
\tag{2.2}$$

for the general case of a size distribution where l is a physical length. This is essentially the total effective cross-section per area presented to the incident beam. k_{ext} is the extinction coefficient. Of course, Q_{ext} can vary over the radius limits and an appropriate average for it is necessary to place it outside the integral. An integration over l would be needed if $n(r)$ is a function of position. As with any classic energy attenuation law where $dE/dl = -k_d E$, $e^{-\tau_{ext}}$ gives the fraction of energy making it through the path without being attenuated or equivalently, the probability that a photon passes through the path without being scattered or absorbed. Strictly speaking, τ is valid for a single wavelength only because a sum of exponentials is not, in general, an exponential. The total optical path encountered by a photon incident from the zenith on a plane parallel cloud layer is usually referred to as *optical thickness*. The term *optical depth* will be used when referring to optical positions within the layer. Optical paths and extinction coefficients for scattering and emission alone can also be defined. For optical depths much less than one, single scattered photons dominate. Optical paths greater than about one generally imply significant multiple scattering although no clear boundary exists and will ultimately depend on the application. For a cloud of optical thickness equal to 10, the average number of scatterings for reflected photons is around 20 in the visible. These escaping photons will have little "memory" of their original

direction or of the details of the phase function. In these cases an average measure of the asymmetry of the phase function is often all that is needed to adequately describe the radiative properties of the cloud. The *asymmetry parameter*, g , is defined as the average cosine of the scattering angle of the phase function,

$$g = \langle \cos\theta \rangle = \frac{\int_{-1}^1 p(\theta) \cos\theta d(\cos\theta)}{\int_{-1}^1 p(\theta) d(\cos\theta)} \quad (2.3)$$

The denominator is usually normalized to unity as previously mentioned. For isotropic scattering $g=0$ and increases with size parameter to about .85 for the forward peaking found in typical cloud drops in the visible. The spread of radii in $n(r)$ also serves to smear out the details of the scattering. Reviews of scattering theory and graphs of the optical parameters for a variety of size parameters and absorptions are shown in Hansen and Travis (1974) and Twomey (1977). A summary of $Q(r)$, $\omega_0(r)$, and $g(r)$ for wavelengths used in the satellite cloud remote sensing of this paper is given in chapter 4.

A not so subtle assumption used in the preceeding discussion, and in radiative transfer in general, is that multiple scattering is incoherent. For this to be valid the scatterers must be in the far field of each others radiation field and they should be randomly distributed in space to cancel any phase coherence (Van de Hulst 1963). Both criteria are met by clouds and so the use of energy (or intensity or flux) instead of a complete wave description is justified. Though the scattered radiation is in general partially polarized, only the total radiation will be considered in this study.

2.1.3 Cloud Albedo

The cloud radiative property of concern in this chapter is the reflection of solar radiation. More precisely it is the albedo, the fraction of incident radiation flux redirected back towards space, that is desired. Cloud albedo is a function of the wavelength dependent optical parameters τ , g , and ϖ_0 , as well as the geometric variable μ_0 . Albedo initially increases rapidly with τ because of the larger cross-section available for scattering. As the optical thickness increases further, it becomes more unlikely that a photon can find its way to cloud base before being reflected and albedo asymptotically approaches some limit. For $\varpi_0=1$ this limit is one, since for infinite τ and no absorption, a photon's fate is to eventually be scattered out the cloud top (given enough scatterings of course). As ϖ_0 decreases, absorption begins to limit the number of scattering events a photon may survive and the albedo approaches some value less than one. The asymmetry parameter indicates how difficult it is for an incident photon to be turned around back towards the top of the cloud by scattering. If $g=1$ then a photon could only be scattered in the forward direction and would never be redirected out the top of the cloud. As g decreases photons need, on average, fewer scatterings to be turned around and the likelihood of reflection therefore increases. To summarize, albedo is a function τ , ϖ_0 , and g^{-1} . The graph of Figure 2.2a shows cloud albedo versus optical depth for various drop sizes. The calculation was made for the spectral response of the satellite sensor used in this study. The channel is in the visible ($\varpi_0=1$) and so the radius dependence is due solely to changing g (g is about 0.75 at $1 \mu\text{m}$ and increases to 0.87 at $20 \mu\text{m}$). Because of the size parameter dependence of Q_{ext} ,

knowing the mean cloud drop radius or $n(r)$ and geometrical thickness is not enough to specify optical thickness. The abscissa of the plot is the actual wavelength dependent optical thickness multiplied by $2/Q_{\text{ext}}$ (2 being a typical value for extinction efficiency). This scaling is useful when graphically trying to compare albedos at two different wavelengths from a model of the microphysics only without having to lookup or otherwise determine Q_{ext} . Figure 2.2b shows results for various values of single scattering albedo.

Various combinations of τ , g , and ω_0 can be made to give the same albedo. Van de Hulst (1980) showed that two different clouds, having values of τ , g , ω_0 for one and τ' , g' , and ω_0' for the other, will have very nearly the same albedo (and other radiative properties as well) if

$$\begin{aligned} (1-g')\omega_0'\tau' &= (1-g)\omega_0\tau \\ (1-\omega_0')\tau' &= (1-\omega_0)\tau \end{aligned} \quad (2.4)$$

The first equation states that the scattering optical thicknesses, weighted by $(1-g)$, must be equal. The weighting is necessary since two clouds can have the same scattering optical thickness but different g and consequently different albedos. As g increases the weighting gives a smaller effective optical thickness to account for the reduced albedo. The second equation is a statement that the absorption optical thicknesses are equal; g is not a consideration in absorption and so no weighting is needed. These relations are intuitively satisfying but must be verified numerically. Supposed a reference cloud is represented by the primed parameters and that calculations can be made for a fixed g'

and combinations of τ' and ω_0' (these two parameters alone give a unique albedo).

Then for an actual cloud with parameters τ, g , and ω_0 , calculations using g' and

$$\begin{aligned}\tau' &= \tau \left[1 + \omega_0 \left(\frac{1-g}{1-g'} - 1 \right) \right] \\ \omega_0' &= \frac{\omega_0}{\omega_0 + \frac{1-g'}{1-g} (1-\omega_0)}\end{aligned}\tag{2.5}$$

will give a close approximation to the correct albedo of the cloud. If the reference cloud has isotropic scattering ($g'=0$) and $\omega_0=1$ then the two equations simply reduce to $\tau'=\tau(1-g)$ and $\omega_0'=1$. So albedo calculations for isotropic and conservative scattering can be used to give albedo for asymmetric scattering by this simple substitution for τ' . The two stream approximation for isotropic, conservative scattering (see for example Bohren 1987 for a review) is

$$A = \frac{\tau'}{2 + \tau'} = \frac{\tau(1-g)}{2 + \tau(1-g)}\tag{2.6}$$

where A is the albedo. This is an approximation to the extent that radiation is confined to downward and upward directions only ($\mu=\pm 1$) but yet provides suprisingly good results and has the advantage of being analytically easy to manipulate. The two stream solution for $g=0.85$ is shown in Figure 2.2a for comparison.

If the radiation is reflected back into a particular direction, such as towards a satellite or aircraft sensor, then the details of the phase function may be necessary. This direction specific reflection is usually specified in terms of a bidirectional

reflectance, denoted by R , which is the albedo that would exist if the detected radiation represented an isotropic reflection. The power per solid angle passing through a unit perpendicular area is intensity, I , (or often radiance in the literature) and can be inferred from the sensor measurement. Then bidirectional reflectance is $R = \pi I / \mu_0 F_0$ where F_0 is the plane parallel solar flux density incident on the cloud. The scaling relations of equation 2.4 can still be used for bidirectional reflectance calculations when total energy into a specific zenith angle is required or when reflected photons have undergone enough scatterings to smooth out the details of the phase function.

2.2 Susceptibility

Twomey (1974) discussed the following link between pollution and cloud albedo. Since combustion processes are known to be prolific sources of CCN, a cloud forming in a polluted air mass will end up with larger concentrations of cloud drops than for the same cloud developing under identical circumstances in cleaner air (see previous section). From the definition of optical depth, an increase in N leads to an increase in τ ($\tau \propto r^2 N$) which in turn leads to an increase in albedo. But this is not the complete story. It is reasonable to assume that clouds forming under the same set of circumstances but with different CCN amounts will have the same supply of vapor available for drop growth. For such a case the liquid water content of the mature clouds would be expected to be equivalent. Drop radius is inversely related to N for constant liquid water content indicating that drop sizes in the polluted cloud will be smaller than for the clean cloud. A decrease in drop radius serves to increase τ and

likewise albedo. With constant liquid water content,

$$\tau = Q_{ext} \pi h \left(\frac{\frac{3}{4}W}{\rho_l \pi} \right)^{\frac{2}{3}} N^{\frac{1}{3}} . \quad (2.7)$$

assuming that $\bar{r} = (\frac{3}{4}W/\rho_l \pi)^{\frac{1}{3}}$. Grassl (1982) analyzed this common assumption with Mie calculations and found it to give errors of <5% in τ for all size distributions considered, with the exception of those associated with fog. The end result of these competing effects is an optical depth which will increase with $N^{\frac{1}{3}}$.

The question then arises as to the significance of this albedo change. Continental air is generally particle-rich from natural sources and so pollution might not have an important effect on albedo. Clean maritime air with its much lower CCN concentrations is more susceptible to pollution. From figure 2.2 it is also clear that the change in albedo resulting from such a change in τ depends heavily on the original value of τ in clean air. The larger τ , the smaller $\frac{dA}{d\tau}$ and of course, in the limit, a cloud reflecting

all incident energy cannot be made any brighter. At very small optical depths $\frac{dA}{d\tau}$ is

large. But ultimately it is not the change in albedo with optical depth that is of interest but rather its change with N . This is a more complicated relation because τ is not linear in N and $\frac{d\tau}{dN}$ is a function of both τ and N . Also, in general, the corresponding

change in radius effects the drop's single scattering albedo and asymmetry parameter.

It is useful then to define a parameter that will characterize the extent of the link between changing cloud drop concentrations and the resulting change in albedo. The differivative $\frac{dA}{dN}$ represents such a link (Twomey 1989). Since, in general,

$A=A(\tau, \omega_0, g)$, the derivative can be expressed in the form

$$\frac{dA}{dN} = \frac{\partial A}{\partial \tau} \frac{d\tau}{dN} + \frac{\partial A}{\partial \omega_0} \frac{d\omega_0}{dN} + \frac{\partial A}{\partial g} \frac{dg}{dN} \quad (2.8)$$

and under the condition of constant liquid water content is termed *cloud susceptibility*. Being a derivative, susceptibility cannot immediately be applied to non-differential changes but does provide a practical measure for the sensitivity of albedo to changes in drop concentration. Note that all terms are wavelength-dependent. The wavelengths of primary interest will be in the visible where about one-half of solar flux occurs and absorption is very small. In the first term, $\frac{d\tau}{dN}$ can be shown to be equal to:

$$\frac{\tau}{3N} = (4\pi\rho_l/9W) \tau \bar{r}^3$$

with the assumption that

$$\bar{r}^2 \approx (\bar{r}^3)^{2/3}.$$

From now on no distinction will be made between the different moments of the radius.

Unless using approximations such as the two stream, numerical values for $\frac{dA}{d\tau}$ will be needed.

In the near infrared where absorption is weak, a geometrical optics approximation gives

$$\varpi_o \approx 1 - Ck_B r$$

(Twomey and Bohren 1980). $C = .85$ for a real refractive index of 1.33, typical at these wavelengths, and k_B is the bulk absorption coefficient of liquid water. This allows the second term of the equation to be expressed as

$$\frac{\partial A}{\partial \varpi_o} \frac{.85k_B}{\left(\frac{3W}{4\pi\rho_l}\right)^{\frac{1}{3}}} r^4 \quad (2.9)$$

Analytic approximations can be made for $\frac{\partial A}{\partial \varpi_o}$ for deep layers (Twomey and Bohren 1980) but otherwise numerical differentiation is needed. Other than expressing

$$\frac{dg}{dN} = \frac{\partial g}{\partial r} \frac{dr}{dN}$$

little more can be done with the third term.

A greater understanding of susceptibility can be had by studying the special case of conservative scattering which is applicable in the visible. There $\varpi_o = 1$ and g is approximately constant with radius (see for example Chapter 4 or Grassl 1982), the last two terms in equation 2.8 can be neglected and susceptibility reduces to

$$\left(\frac{dA}{dN}\right)_{W=const.} = \frac{\partial A}{\partial \tau} \frac{\tau}{3N} = \left[\frac{4\pi\rho_l}{9W} \right] \tau \frac{\partial A}{\partial \tau} r^3 \quad (2.10)$$

where either N or r can be used as an independent variable. Of obvious note is the r^3

dependence in the second form. The term $\tau \frac{\partial A}{\partial \tau}$ is given by the two stream approximation as

$$\begin{aligned} \tau \frac{\partial A}{\partial \tau} &= \tau \left(\frac{1-g}{2+(1-g)\tau} - \frac{(1-g)\tau}{(2+(1-g)\tau)^2} \right) \\ &= \frac{A(1-A)}{3} \end{aligned}$$

and is shown as a function of both albedo and scaled optical thickness in Figure 2.3. The peak is at an albedo of 0.5 and an optical thickness of about 13. These curves multiplied by r^3 (and a term containing liquid water content) become susceptibility. An important point should be made with regard to equation 2.10; as might be intuitively expected the percent increase in existing albedo monotonically increases as $\tau \rightarrow 0$, however the net change in albedo becomes smaller. It is this consideration of net albedo change, which is ultimately "seen" by the climate system, and not percent change that accounts for a maximum at $A=0.5$. Surfaces of susceptibility for the two stream approximation are found in Figure 2.4a for $W = 0.3 \text{ g/m}^3$ versus both τ and r . In Figure 2.4b, τ has been mapped into albedo. The two stream approximation can also be used to show albedo changes due to non-differential changes in drop concentration. Consider a cloud having an albedo A for some drop concentration N . If N is changed by some factor χ (i.e., $N \rightarrow \chi N$) then $\tau \rightarrow \chi^{1/3} \tau$ and

$$\Delta A = \frac{A(1-A) (\chi^{1/3} - 1)}{A(\chi^{1/3} - 1) + 1} \quad (2.12)$$

At $A=0.5$, a doubling in N will increase albedo to a value of about 0.56. Results for

other values of χ are shown in figure 2.5. Notice that the peak ΔA always occurs at $A = 0.5$ just as does the differential equivalent susceptibility.

2.2.1 Broadband Albedo

Broad-band albedo found by integration over the solar spectrum is important in studying the earth's radiation budget. Such an integration would have to be taken from the near ultraviolet into the near infrared to capture most of the sun's energy (it is estimated that 95% of solar energy lies within .3-2.5 μm , Wydick et al. 1987). Likewise susceptibility of the broad-band albedo to changes in drop concentration would involve the calculation

$$\frac{d}{dN} \left[\frac{\int_{\lambda} A(\tau, \varpi_0, g, \lambda) F_0(\lambda) d\lambda}{\int_{\lambda} F_0(\lambda) d\lambda} \right] = \frac{\int_{\lambda} \frac{dA(\tau, \varpi_0, g, \lambda)}{dN} F_0(\lambda) d\lambda}{\int_{\lambda} F_0(\lambda) d\lambda} \quad (2.13)$$

where all terms in equation 2.8 would have to be addressed. Accounting for atmospheric effects (see chapter 5) above cloud top is also needed in a complete study. An easier, though approximate, approach is to find a relation for broad-band albedo based on the albedo for a visible wavelength and apply this to susceptibility.

For use with narrow-band satellite channels, several studies have attempted to find a regression between narrow-band albedo in the visible and the integrated broad-band calculation. The spectral albedo of clouds decreases into the near infrared due to increasing absorption by liquid water, so it is expected that visible albedo will exceed the broad-band. The difference between the two will depend on cloud optical thickness

and microphysics. Wydick et al. (1987) made numerical and empirical studies. Numerical results for $\mu_0 = 1$, including atmospheric scattering and absorption, gave the linear approximation $\langle A \rangle_\lambda = .8 A_{\text{vis}} + .78\%$ for channel 1 of the Advanced Very High Resolution Radiometer (AVHRR) satellite instrument (used in this study and discussed in more detail in the following chapter) as a fit for all cloud optical thicknesses considered. Apart from the small offset, A_{vis} overestimates broad-band albedo by 25%. An assumption that bidirectional reflectance is equivalent to albedo was used. Empirical comparisons of the AVHRR with the broad-band Earth Radiation Budget (ERB) satellite showed more spread but could only be made with coincident views of the two satellites. Shine et al. (1984) made a similar study for the AVHRR channel 1. For clouds with optical thicknesses of 10, 20, and 30, broad-band albedo was over-estimated by 12%, 17%, and 20% respectively using a mid-latitude model with $\mu_0 = 1$. For $\mu_0 = 0.5$ the over-estimation showed less spread with respect to optical thickness; 14%, 16%, and 18% respectively. Results from Laszlo et al. (1988) are similar. If a linear regression is appropriate then the integrated susceptibility becomes

$$\frac{d\langle A \rangle_\lambda}{dN} = \alpha \frac{dA(\lambda)}{dN} \quad (2.14)$$

with α being the slope of the fit. Further mention of susceptibility will imply the narrow-band visible value. Equation 2.14 may be used as an approximation for inferring changes in broad-band albedo.

2.2.2 Ship Tracks

A spectacular example of cloud albedo modification was first identified by Conover (1966). Puzzling anomalous linear cloud features had been occasionally seen with early TIROS satellite pictures. Conover identified these as "tracks" left by ocean-going vessels whose exhaust, being a source of CCN, is carried aloft as a buoyant plume into the stratus deck. Direct sightings by aircraft (Conover 1969) of ship tracks behind vessels helped confirm this theory. These tracks may be hundreds of kilometers long and as much as 50 km in width. Also noticed in those years was that ship tracks are mostly absent from stratus although ship traffic is certainly always present. Either some type of meteorological influence must be at work, the susceptibility of the stratus is critical, or both. Tracks often exist in what appears to be a mature cloud (reflectance apparently unchanged during several hours of GOES observation) where activation has presumably ended and cloud supersaturation would be expected to be too small to activate a new input of CCN. In the absence of a new source of vapor, some mechanism(s) must be at work to allow the ship's CCN to scavenge liquid water away from existing drops; for example the evaporation of existing drops from entrainment by the warm plume which subsequently cools to allow activation. Otherwise there would have to exist a continual process of cloud development, balanced by removal, to give both the appearance of mature clouds and provide a mechanism for activating ship produced CCN. Several general studies of stratus development have been made (see for example Telford and Chai 1984, Koracin and Rogers 1989) but the answer to the above remains uncertain. The important point to be made here is that anthropogenic

modification of cloud albedo is more than just a theoretical concern, but readily validated in relatively common ship tracks. Recent studies of ship tracks include Twomey et al. (1987) and Coakley et al. (1987). Radke et al. (1980) measured factor of two changes in drop concentration within tracks which were concurrently observed with satellites. These tracks provide a useful test for remote sensing cloud microphysics and susceptibility in chapter 6.

2.2.3 Global Albedo

Thus far the susceptibility of a specific cloud was considered and shown to be a function of drop radius and optical thickness (or albedo if the g dependence on radius is ignored). So the global effect of increasing CCN concentration depends on geographical and temporal distributions of these two parameters. If probability distributions for cloud albedo as a function of geographical location existed then estimates could be made of global susceptibility. Cloud climatologies based on satellite imagery are currently being developed and can begin to provide information on optical thickness distributions, at least with coarse geographic resolution (Rossow and Schiffer 1991). Remote inference of radius is more difficult and is undertaken in this study. The net contribution of cloud albedo to global albedo also depends on the reflectance of the underlying surface for clouds with significant transmission. Ignoring surface effects, at a given location

$$\left. \frac{dA}{dN} \right)_W = \int_{\tau} \int_r \left. \frac{dA}{dN}(\tau, r) \right)_W p(\tau, r) d\tau dr \quad (2.15)$$

where $p(\tau, r)$ is a joint probability for the temporal distribution of τ and r . If liquid water content is variable then it would have to be included in the probability. Finally, an integration over the earth's surface is needed to complete the analysis with plane-parallel albedo converted to spherical albedo

$$2 \int \mu_o A(\mu_o) d\mu_o .$$

Though the information necessary for such analysis is not, and may never be, completely available it is important to be aware of the complexity of the detailed analysis.

2.2.4 The Influence of Liquid Water Content

A note on liquid water content is in order. Though marine stratus are notorious for their longevity, Albrecht (1989) has reported measuring drizzle drops ($> 40 \mu\text{m}$ radii) in these clouds and concluded that precipitation must be occurring. He theorizes that increases in CCN in clean maritime clouds which decrease the mean drop size will decrease the production of drizzle drops. Consequently the clouds would contain a higher liquid water content, i.e., less drizzle leaving the cloud. Such a connection, if generally true of marine stratus, would suggest that defining susceptibility for a process of constant liquid water content should be modified to allow for an increase. This would serve to further enhance cloud brightening. Measurements by Radke et al. (1989) showed increased liquid water content inside ship tracks compared with cloud

areas away from the tracks. Although it could be argued that the development of the track can certainly be affected in other ways by the ship's passage, it is useful to look at this suggestion.

Consider susceptibility (in the visible, equation 2.10) for such an assumption. it seems reasonable to assume some type of power law to relate liquid water content and drop concentration of the form $W = \alpha N^\beta$. With $0 < \beta < 1$ the increase in liquid water content with N of the mature cloud tapers off as N increases, that is, it becomes harder to have drizzle rain out with large N . This is about the only characteristic such a relation should be certain to have based on Albrecht's hypothesis. Measurements in and out of ship tracks, shown in Figure 3b from Radke et al., give $W \approx 0.3 \text{ g/m}^3$ and 0.5 g/m^3 for $N \approx 30 \text{ cm}^{-1}$ and 100 cm^{-1} respectively ($\alpha \approx 0.07 \text{ g/m}^3$ and $\beta \approx 0.42$). Then

$$r^2 = \left(\frac{3\alpha}{4\pi\rho_1} N^{\beta-1} \right)^{\frac{2}{3}}$$

which can be substituted into equation 2.2 for τ . Taking the derivative $\frac{\partial \tau}{\partial N}$ gives a susceptibility of

$$\begin{aligned} & \frac{\partial A}{\partial \tau} \frac{\tau}{3N} (1 + 2\beta) \\ &= \left[\frac{4\pi\rho_1}{9W} \right] \tau \frac{\partial A}{\partial \tau} r^3 (1 + 2\beta) . \end{aligned} \tag{2.16}$$

The results are identical with equation 2.10 with the exception of factor $(1+2\beta)$. For

constant liquid water content, $\beta=0$ and the two equations are identical. With the Radke et al. data, the factor becomes about 1.8 indicating that susceptibility would be increased by 80% over the constant liquid water assumption for all initial values of W . As for equation 2.10, knowledge of the initial liquid water content is still needed for the expression containing radius. In light of the critical dependencies of susceptibility on radius (or N) and optical depth, this underestimation of 44% seems minor. Including the uncertainty involving the need to even account for changing liquid water content suggests that the original definition of susceptibility is both adequate and preferable.

2.2.5 Other Effects of Pollution on Albedo

Clouds not only affect the radiative balance of the earth by reflecting radiation in the shortwave but also by emission in the longwave. Therefore the effect of increasing CCN amounts on the infrared properties of the cloud should also be mentioned. If a cloud is already black in the infrared and at a constant temperature with height then CCN can have no effect. A study by Grassl (1982) considered a cloud with a non-zero temperature lapse rate and with various size distributions. As CCN increase, optical depth increases and the effective layer that governs emission will move upwards toward cloud top. This layer will be at a lower temperature and so emission will decrease. This would act to warm the earth-atmosphere system and partially compensate the cooling from enhanced solar reflection. But as optical depths increase, drop sizes decrease, and through much of the thermal infrared this results in a decrease in ϖ_0 for most size distributions studied (opposite to the dependence for near infrared

wavelengths) and so the possibility exists that with enhanced drop absorption, emissivities will increase and act to support cooling. Grassl studied these two competing effects and found that the latter dominated. The net effect of increasing CCN on the infrared balance was to cause a cooling of the earth-atmosphere system of about 3% of what was contributed by the increase in solar albedo. The consequences of altered cloud albedo therefore far outweigh any influence in the IR.

CCN are but a subset of the atmospheric aerosol. Pollution is also a source of carboniferous aerosol which absorbs in the visible and is inactivated in cloud development. This aerosol would also be present in regions of the atmosphere where the cloud does not form. The resulting increase in absorption coefficient serves to lower albedo. Twomey (1977b) considered the change of albedo with optical thickness (not with N) and assumed a link between CCN and absorption optical thickness. As pollution increased optical thickness, a decrease in the enhanced visible albedo of the cloud-atmosphere system occurred only for clouds with large optical thicknesses. The effect of increasing aerosol absorption can only be seen in clouds which are optically thick because enhanced brightness from increasing the thickness is asymptotically limited. That is, the brightest clouds can't be made brighter. The situation was also studied by Grassl (1982) for various size distributions using measurements of absorption coefficient from a polluted industrial area assumed to represent an upper limit to global aerosol absorption. As an example, aerosol absorption lowered albedo by a few percent when the clean cloud albedo was about 70%.

2.3 The Remote Sensing of Susceptibility

An aircraft or satellite sensor flying over a cloud can in principle measure bidirectional reflectance at various wavelengths and infer cloud optical thickness and mean drop radius. As described in Chapter 4, full radiative transfer calculations for bidirectional reflectance can be made for any applicable geometry that would be encountered. The general dependence of bidirectional reflectance on the optical parameters is similar to albedo. Optical thickness can be determined from reflectance measurements in the visible (Figure 2.2a) though some uncertainty is caused by variation of the asymmetry parameter with drop radius. Strictly speaking, because of this uncertainty in g , it is only the scaled optical thickness, τ' , which can be inferred when using only visible wavelengths. For scaling with respect to a conservative and isotropic scattering cloud, $\tau' = (1-g) \tau$ is the value of the scaled optical thickness necessary to account for the measured reflectance. If an additional channel(s) can be used to determine drop radius, and thereby g , the true optical thickness can be calculated. Single scattering albedo is uniquely related to drop radius for a given wavelength and generally decreases for liquid water drops into the near infrared. The greater absorption asymptotically limits reflectance. At least for large optical depths it is clear from Figure 2.2b that measured reflectances for a wavelength with $\omega_0 < 1$ can be equated with a single scattering albedo which in turn, for that wavelength, implies a unique cloud drop size. Hansen and Travis (1974) show that it is actually the *effective* radius which is sensed for a size distribution of drops. This will be discussed in Chapter 4. Remote sensing radius and optical thickness is then sufficient to calculate

susceptibility as long as some assumption regarding the value of liquid water content is made. Or, more exactly, the product of susceptibility and liquid water content can be calculated. Drop concentration cannot be sensed directly and so the radius dependent form of susceptibility needs to be used.

The wavelengths of choice are those in the so-called near infrared windows where water vapor absorption is at a minimum (but not negligible). Transmission windows exist at about 1.0, 1.2, 1.6, 2.2, and 2.75 μm in the near infrared. Table 2.1 lists the range of ϖ_0 at each of the window wavelengths for drop radii between 6 and 20 μm (the 3.75 μm wavelength is referred to by some as being in the mid infrared). The approximation that $\varpi_0 \approx 1 - .85_{\text{Pr}}$ (Twomey and Bohren 1980) has been used and is valid for weakly absorbing wavelengths. At 3.75 μm , liquid water bulk absorption is an order of magnitude larger than at 2.2 μm . With this large absorption the approximation still is adequate, however Mie calculations show that ϖ_0 varies with approximately r^{-5} for the radii considered. For this wavelength, the actual Mie values are shown along side that of the approximation.

With ϖ_0' as the scaled optical thickness for isotropic scattering, Van de Hulst (1964) introduced

$$\tau_c = \frac{2}{\sqrt{1 - \varpi_0'}}$$

as the approximate minimum optical thickness for which the asymptotic limit to reflectance occurs. A more involved approximation is given by Sato et al. (1977), but the exact thickness where the asymptotic limit is considered to be reached is of course

arbitrary anyway. The Van de Hulst approximation is said to be within 2% of the limit. For thicknesses greater than this critical value, reflectance is a function of radius only. The ideal choice for the absorbing wavelength would be one which was influenced by radius only. This would complement the visible channel measurement which primarily gives information about optical thickness. The $3.75 \mu\text{m}$ wavelength comes closest to that case because of its large absorption. Ideally, the choice of wavelengths would include all those just mentioned to insure that some channels were influenced by both optical thickness and radius. Table 2.2 gives calculations for critical thickness for the Van de Hulst approximation using the range of ω_0 for the wavelengths of Table 2.1. The values of g at $6 \mu\text{m}$ and $20 \mu\text{m}$ needed for converting ω_0 to ω_0' and then τ_c' to τ_c came directly from Mie calculations. From a practical matter (e.g., sensor calibration, atmospheric effects) the reflectance limit occurs well before the 2% error of the approximation is reached. Later graphs of reflectance for AVHRR channel 3 (centered at $3.75 \mu\text{m}$) suggest a critical optical thickness of 4 to 7 for a radius of $20 \mu\text{m}$ and $6 \mu\text{m}$ respectively, about half of the value given by the approximation. So the results of Table 2.2 appear to be conservative.

The idea of using near infrared absorption to infer drop radius was originally applied to clouds on Venus by Pollack et al. (1974, 1978). Twomey and Cocks (1982, 1989) first applied the technique to terrestrial clouds using aircraft borne sensors. Other aircraft studies have been conducted by Stephens and Platt (1987), Foot (1988), and Nakajima et al. (1991). Two or more near infrared wavelengths were used though none of the investigators made measurements in the $3.75 \mu\text{m}$ window. In situ cloud

measurements were simultaneously obtained with the typical result being that drop radius was overestimated by $2\text{ }\mu\text{m}$ to $5\text{ }\mu\text{m}$. Rozenberg et al. (1974) were the first to report satellite results with near infrared channels aboard the U.S.S.R. Cosmos satellite

Table 2.1 The range of single scattering albedos, using the approximation given in the text, available for remote sensing cloud drop radius at four of the near infrared wavelengths. The visible is shown for reference. The range of possible radii is limited to $6\text{ }\mu\text{m}$ (the first entry for ω_0) to $20\text{ }\mu\text{m}$ (the second entry). The values in parenthesis for $3.75\text{ }\mu\text{m}$ are the actual limits of ω_0 calculated with Mie code. Liquid water absorption coefficients are from Hale and Querry (1973), size distributions for the calculations had a dispersion of .20.

Range of ω_0 for $6\text{ }\mu\text{m} < r < 20\text{ }\mu\text{m}$		
Wavelength(μm)	$k_{\text{Bulk}}(\text{cm}^{-1})$	ω_0
.65	10^{-3}	1.0000-.9999
1.2	1.04	.9995-.9982
1.6	6.2	.9968-.9895
2.2	16.5	.9916-.9720
3.75	119.0	.9393-.7977 (.9247-.8196)

Table 2.2 The range of critical optical thickness using the Van de Hulst approximation. The first entry in each column corresponds to a drop radius of 6 μm and the second entry to 20 μm . Single scattering albedos were taken from table 2.1. Calculations are based on a size distribution with dispersion equal to .20.

$\lambda(\mu\text{m})$	g(Mie code)	ω_0'	τ_0'	τ_0
1.2	.8372-.8699	.9969-.9863	36.13-17.11	221-130
1.6	.8299-.8680	.9815-.9256	14.67-7.33	85.1-52.0
2.2	.8262-.8721	.9535-.8162	9.28-4.66	51.3-30.6
3.75	.7116-.8711	.7798-.3693	4.26-2.52	12.46-8.80

and came to a similar conclusion. Inferring larger drops implies that the observed ω_0 is lower than would be expected from calculations based on measured drop sizes. This has been termed anomalous absorption and is yet to be understood. Stephens and Tsay (1990) give a review of the measurements and comment on proposed causes of the anomaly. Suggested causes include continuum vapor absorption in the windows and cloud inhomogeneities. It is not clear whether radius inferred from a $3.75 \mu\text{m}$ channel would suffer from a similar anomaly.

The ability to remotely sense drop radius from satellites is limited to the current orbiting channels. The NOAA polar orbiter AVHRR instrument and the Landsat Thematic Mapper (TM) are two candidates. While the TM has channels in most of the near infrared windows except $3.75 \mu\text{m}$, the AVHRR has only a $3.75 \mu\text{m}$ channel available. The AVHRR is aboard two NOAA satellites placed in orbits which allow daily local passes in the morning and afternoon for every location on earth. The sole Landsat is placed into an orbit which only allows repeat coverage every 16 days. Arking and Childs (1985) used the AVHRR for remote sensing cloud drop radii and optical depths though no in situ cloud measurements were used for validation. Grainger (1990) used the AVHRR for studying orographic effects on drop sizes. In a manner similar to these works, the visible and $3.75 \mu\text{m}$ channels of the AVHRR are used in this study for remote sensing susceptibility and are explained in some detail in the next chapter.

It is worth noting that the researchers using a $3.75 \mu\text{m}$ channel are those who feel a need for a daily coverage satellite and have found the AVHRR to be the most

practical choice. A few disadvantages for using this wavelength are worth pointing out. With increased absorption, there is less reflection and so less signal available. The ability to distinguish large radii from one another is more difficult because of the small differences in absolute reflectance. Also, as mentioned above, the $3.75\ \mu\text{m}$ channel shows a weaker dependence on radius than do the less absorbing wavelengths. Perhaps the biggest concern is that emitted radiation at this wavelength is comparable to, and can even exceed the reflected radiation. This crossover between emission and reflection results from the longer wavelength, larger cloud emissivities, and the small reflection. It is most likely the last reason which has kept researchers operating with aircraft sensors from including this wavelength. However, the large absorption does give it one very important advantage. The $3.75\ \mu\text{m}$ signal is almost entirely dependent on radius, with very little optical thickness sensitivity except for the thinnest clouds. Shorter wavelengths have substantial optical thickness dependencies over much of the expected range. The substantial radius information content of a $3.75\ \mu\text{m}$ channel makes it of special interest for remote sensing regardless of it being a major contender for satellite studies.

2.4 Marine Stratus Measurements

There are several reasons for studying marine stratus and stratocumulus. These clouds are common throughout all oceans of the world and have important influences on climate. Modification to these clouds from pollution is of major interest. Along the mid-latitude western coast of the continents stratus is particularly extensive and uniform. This uniformity approaches the homogeneous, plane-parallel cloud assumption of the

radiative transfer calculations as well as any cloud type can be expected. Being capped by a temperature inversion, they are primarily low clouds and certain to be liquid water, except perhaps at higher latitudes.

A summary of several microphysical measurement studies made in these clouds is provided in Table 2.3. The typical method for measuring size distribution is through an optical scattering instrument such as the Forward Scattering Spectrometer Probe. The work by Warner (1969), which captured drops by impaction on slides, is the only exception in the table. Often the liquid water content is calculated directly from the drop distribution, or occasionally measured directly with instruments such as a hot-wire probe. Individual measurements normally were carried out in localized areas; many oceans are represented in the table but measurements were almost exclusively coastal. It is impossible to infer a climatology from these studies but the general agreement suggests that the liquid water content lies within a relatively narrow bound about a nominal value of 0.3 g/m^3 . The spread in mean radius is also relatively constrained. All data represent some averaging over appropriate scales.

Climatologies of marine stratus albedo are beginning to be developed with satellite observations. Heck et al. (1990) used GOES to analyze stratus in the northeastern and southeastern Pacific (subtropical to mid-latitude zones) for July 1983-1989, January 1984-1989, April and October 1985-1987. The North and South Atlantic was studied with METEOSAT (European Space Agency geosynchronous satellite) over similar dates as for GOES. In the study, areas with low cloud tops and little midlevel or high clouds were considered stratus and used in the statistics. A monthly climatology

for July 1985 based on these data showed that the waters off the western coast of South America and Africa had the greatest mean stratus cloud amounts (they report that it is typical for the winter hemisphere to have greater cloud amounts). However, estimates of mean optical thickness during the same month showed the region off the southern California coast as having the largest values ($\tau > 12$ was the largest optical depth bin considered). It is at an optical thickness of about 13 that the term $\tau \frac{\partial A}{\partial \tau}$ in susceptibility has its maximum. Over extensive areas (about 30° of latitude and 40° of longitude) the mean optical thickness was typically greater than 6. The mean cloud heights, averaged over all years considered, indicated the lowest heights near the coasts (500-1000 m) gradually increasing away from the coasts. Patterns for each region indicated maximum cloud amounts around sunrise and early morning with minima during mid to late afternoon. This suggests that remote sensing marine stratus with the AVHRR, which has a local morning or afternoon pass only, could be misleading if only one pass were to be used. An eight month sample of stratus at San Nicolas Island (off southern California coast) during 1987 by Fairall et al. (1990) showed that the most frequently observed albedo was around 50%, where maximum susceptibility (for a fixed radius) occurs.

Based on these results, stratus with typical optical thicknesses from 6 to 12, drop radii from 5 to 15 μm , and liquid water content from 0.1 to 0.5 g/m^3 would be expected to have susceptibilities ranging from about 10^{-5} to 10^{-2} cm^3 . Since

susceptibility is defined for absolute changes in albedo, a value of 10^{-2} implies that adding a single drop per cm^3 to the existing cloud will add an additional 1% to its albedo!

PAPER	DATE	LOCATION	CLOUD TOP - BASE	LWC (g/m ³)	$\bar{r}$ (μ m)	$\sigma/\bar{r}$	r_{eff} (μ m)	V_{eff}	N (cm ⁻¹)
Warner (1969)	4/59 - 12/67	Eastern Australian Coast	(90 clouds studied)	Typically 0.4, up to 1.0 with strong convection	12 (base) 15 (top)	0.20 0.45 0.21 (avg.)			
Slingo, Nicholls, & Schmelz (1982)	8/8/78 (1 day only)	~60°N, 13°W	Inversions at ~920mb, 900m	~0.5 (avg.)	5 (base) 11 (top)	0.40 0.12			100 100
Noonkesler (1984)	5/81	San Diego Coast		0.22 (base) 0.29 (top)	4.8 5.8	0.44 0.45			296 228
Albrecht, Randall, & Nichols (1988)	Summer 1987 (fire)	California Coast		0.1 - 0.5 g/m ³					
Duda, Stephens, & Cox (1991)	7/7/87 7/10/87 (fire)	San Nicolas Island	945mb (base) 890mb (top)	0.50 max 0.14 max	8.0 5.4	0.51 0.79	11.41 11.54	0.086 0.149	50 - 100
Rawlings & Foot (1990)	6/30 - 7/18 (5 days fire)	So. California Coast					7.3 - 9.2 7.3 - 7.8 3.6 - 7.8 4.1 - 10.2 4.2 - 5.4		
Tsay & Jayaweera (1984)	6/13/80 6/20/80 6/30/80	Beaufort Sea	251m 233m ~300m (thickness)	0.09 0.21 0.14	5.5 4.7 5.0				73 222 138
Nakajima, King, Spinhrne, & Radke (1991)	7/10/87 (fire)	California Coast					5 - 15		
Felgelson (1978) (USSR data summary)		USSR		~0.26					

Table 2.3 Summary of microphysical measurements made on maritime clouds.

Figure 2.1 Schematic representation for a cloud.

Figure 2.2 Albedo for (a) AVHRR visible channel 1. The albedo shown is the average over the spectral response of the channel and g is the channel averaged asymmetry parameter for the given drop radius. The two stream approximation is also shown. (b) Albedos for various single scattering albedos representative of the near infrared wavelengths used in cloud remote sensing, $g=0.86$ in each case. These last calculations were made without regard to actual wavelengths or radii so the $2/Q_{ext}$ normalization is unnecessary. The cosine of the solar zenith angle is 0.85.

Figure 2.3 Two stream approximation for the term $\tau \frac{\partial A}{\partial \tau}$ in Equation 2.10 for susceptibility.

Figure 2.4a Calculated surface of susceptibility versus radius and optical thickness, using the two stream approximation with a liquid water content of 0.3 g/m^3 .

Figure 2.4b Calculated surface of susceptibility versus radius and albedo, using the two stream approximation with a liquid water content of 0.3 g/m^3 .

Figure 2.5 The effect on albedo of changing cloud drop concentration from $N \rightarrow \chi N$ for various factors χ using the two stream approximation.

CHAPTER 3

THE ADVANCED VERY HIGH RESOLUTION RADIOMETER

Remote sensing of cloud optical thickness and drop radius is made possible through solar reflection measurements at a non-absorbing (visible) wavelength and at some combination of absorbing (near infrared) wavelengths. As mentioned in chapter 2, the visible and $3.75\ \mu\text{m}$ channels of the Advanced Very High Resolution Radiometer (AVHRR) aboard the NOAA polar orbiting satellites are used in this study. Though some justification was given for the use of this satellite instrument, a more complete explanation is required. A brief review of the history and capability of available satellite instruments is given at the beginning of this chapter. A detailed description of the AVHRR instrument, data set, calibration, and flight geometry follows.

3.1 Overview of Available Satellites and Instruments

The sun-synchronous and geosynchronous orbits are two primary ones used for meteorological observations. The sun-synchronous satellite is placed in a near polar orbit where the plane of the orbit precesses about the earth once during the course of a year (i.e., sun-synchronous) so that the plane always has a constant reference angle to the sun. During an orbit, the earth's rotation causes the surface to pass beneath the satellite. One daytime and nighttime pass is then made over every area at about the same local time. The orbits are typically inclined at about 10 degrees (with respect to the poles) so practically the entire surface of the earth, with a possible exception of the

poles, can be imaged if the particular instrument has sufficient scan capability. The height of an orbit is typically hundreds of kilometers and orbit time is on the order of one and a half hours. Examples of such satellites include the NOAA polar orbiters, Landsat, and the Nimbus experimental satellite among others. The geosynchronous (or geostationary) satellite is at a height of about 35,700 kilometers with an orbital period of one day such that it is stationary over a fixed point on the equator. This excludes the poles, and higher latitudes from being imaged depending on acceptable viewing angles. About five equally spaced satellites would be needed to have complete longitudinal coverage. Examples include the U.S. GOES, the European Space Agency METEOSAT, and the Japanese GMS. The complete earth disk can be imaged in about one half an hour and so can provide temporal coverage of cloud development and weather systems.

3.1.1 The NOAA Polar Orbiter

The first generation of these satellites began as the TIROS (television infrared observation satellite) series. TIROS I, launched in April 1960, had a narrow and a wide field of view television camera and was the first satellite to transmit television pictures of the earth. TIROS II, launched the following November, included an infrared radiometer. Over the next five years, eight more TIROS series satellites were launched that eventually included real time automatic picture transmission capability (APT) to ground stations. The next generation of weather polar satellites (from 1966 to 1969) were considered "operational" and designated as ESSA (Environmental Science Service

Administration). The very high resolution radiometer (VHRR) was added to the improved TIROS operational system (ITOS), which became the third generation of TIROS, launched from 1970 to 1976. This instrument contained a visible and thermal infrared channel with resolution of about 1 km at nadir. The fourth and latest generation is called TIROS-N after the first satellite in the series and includes the AVHRR instrument, an improved version of the VHRR containing four to five channels. Subsequent satellites were designated NOAA-6, NOAA-7, in ascending order up to NOAA-11 launched in November of 1988. The even numbered satellites have morning daytime passes and the odd numbered have afternoon passes. A detailed history of TIROS is given by Schnapf (1985).

The AVHRR/2 (referred to here simply as the AVHRR), starting with NOAA-7, is a cross-track scanning radiometer at nominal wavelengths of 0.65, 0.85, 3.75, 10.75, and 12.0 μm referred to as channels 1 through 5 respectively. The 3.75 μm channel provides for the remote sensing of cloud drop size while both of the first two channels are effectively non-absorbing and provide optical thickness information (the next generation of the AVHRR is to include a 1.6 μm channel which would provide an additional absorbing wavelength). Resolution is about 1 km at nadir. The fifth channel is included on the afternoon satellites only. In addition to the AVHRR, the TIROS-N series includes the TOVS (TIROS operational vertical sounder) which itself consists of three instruments, the high resolution infrared sounder (HIRS/2), the stratospheric sounding unit (SSU), and the microwave sounding unit (MSU). The HIRS/2 contains 20 bands, one in the 3.75 μm window. But, being primarily a sounding instrument,

resolution is limited to about 20 km and is therefore of little use for remote sensing cloud microphysics.

The AVHRR has the advantage of containing useful wavelengths for remote sensing at relatively high resolution. Though only two daytime passes are possible (if two satellites are used), the satellite is placed into an orbit such that for the AVHRR's $\pm 55.4^\circ$ scan angle, complete earth coverage is possible. This scan angle corresponds to over 2000 kilometers of surface imaged during one scan. Details of the scan and resulting data sets are presented later.

3.1.2 Landsat

Programs for land remote sensing began in the late 1960's with the first three Landsats (Landsat-1, 2, and 3) launched in 1972, 1975, and 1978 and being based on instruments flown on the Nimbus research satellites. These satellites were equipped with the multispectral scanner (MSS) having four visible and near infrared bands covering 0.5 to 1.1 μm and one thermal infrared band (Landsat-3 only). Resolution was about 80 meters. The Landsat-D program which launched Landsat-4 (1982) and Landsat-5 (1984) added the thematic mapper instrument (TM) which contains near infrared wavelengths at 1.6 μm and 2.2 μm with a resolution of about 30 meters. The TM also has four bands from 0.48 to 0.85 μm and a thermal infrared band.

Based on spectral coverage, the TM is an ideal candidate for cloud remote sensing. With a daytime crossing of about 9:30 am local time, and no afternoon pass possible, Landsat has less flexibility than the NOAA polar orbiters in achieving suitable

solar angles. And because of the Landsat's 180 kilometer surface scan (compared with over 2000 km for the AVHRR) there is no overlap between successive orbits and repeat coverage of the same scene is limited to once every sixteen days. This makes it difficult to study specific clouds features or developments observed through meteorological data or other satellite systems. For instance GOES images, such as of marine stratus off the coast of California, are readily available over various networks within minutes after the picture is completed. Monitoring these GOES pictures allows the location and date of useful cloud conditions to be noted. The chances of Landsat viewing the desired location on a given day is slim. However, the AVHRR is guaranteed at least one daytime view for each satellite. The French SPOT satellite is comparable to Landsat but includes visible channels only.

Use of Landsat-5 for cloud remote sensing has been reported by Nakajima et al. (1991). They mention the saturation of TM channel 2 ($0.57 \mu\text{m}$) for optically thick clouds which forced them to use in situ estimates for cloud optical thickness. This indicates a potential problem with cloud studies that use a sensor designed for the relatively small albedos encountered in land remote sensing. Bradley (1988, private communication) used Landsat-5 data from White Sands, N.M. to infer drop phase based on the remote sensing technique of Pilewskie and Twomey (1987).

3.1.3 GOES

The first geosynchronous satellite for weather observation was NASA's Application Technology Satellite (ATS-1) launched in December 1966. It was originally

conceived as a research satellite for communication systems but included a spin-scan cloud cover camera (SSCC) for obtaining meteorological pictures. Resolution was at about 3.2 kilometers at nadir with peak sensitivity in the green. With the ATS-3 launched in 1967, "color" pictures of the earth were available with the multicolor SSCC (or MSSCC). The geosynchronous very high resolution radiometer (GVHRR) was included on the ATS-6 in 1974. With two channels in the visible and thermal infrared and resolution of 5.5 and 11 kilometers respectively, it provided the necessary detail of cloud height and motion. The next generation of geosynchronous weather satellites included the two prototype synchronous meteorological satellites (SMS-1 and -2) and the geostationary operational environmental satellites (GOES-1, -2, and -3). These were launched between 1974 and 1978. The GOES systems included the VISSR instrument which had a higher visible spatial resolution (0.9 kilometers at nadir) than the GVHRR. And with the use of two satellites, GOES-east and GOES-west, more complete coverage of the western hemisphere was possible. The current generation (GOES-4 through -7) includes the vertical atmospheric sounder (VAS) system consisting of twelve spectral channels throughout the infrared and one visible channel. The VISSR mode mimics the original VISSR instrument. Channel 12 of the VAS is at $3.9\ \mu\text{m}$, useful for cloud microphysics studies, but with a best possible resolution of 6.9 kilometers. In addition, this channel is limited to the dwell sounding (DS) mode which can only cover about ten degrees of latitude and possibly the multispectral imaging (MSI) mode which covers the north limb of the earth disk and the southern latitudes to about 42 degrees. Details of the GOES development is given in the chapter by Greaves and Shenk in Schnapf (1985).

Further details of the current GOES system can be found in the GOES Data User's Guide (1984). Geosynchronous systems such as Meteosat and GMS lack the $3.9\ \mu\text{m}$ channel.

3.2 AVHRR Data

The AVHRR scans in an across-track direction through a rotating mirror. The radiation is then resolved into the various spectral channels and imaged onto the appropriate detector (see Table 3.2). Table 3.1 summarizes the basic AVHRR instrument and satellite parameters. NOAA-11 (and other odd numbered satellites in the series) has a northbound daytime equator crossing and scans from the east to the west. So when viewing the daytime image, successive scan lines are displaced to the north, and pixels in the left half of the image lie towards the east of the sub-satellite point. NOAA-10 (and other even numbered satellites in the series) have southbound daytime equator crossings with the daytime image having north at the top and west towards the left.

3.2.1 AVHRR Channel Spectral Response

The AVHRR instrument (specifically the AVHRR/2 system on NOAA-7 and subsequent satellites) has either four or five channels as specified in Table 3.2 based on pre-launch measurements. The fifth channel is available on the afternoon daytime pass (odd numbered) satellites only. The satellite primarily used in this susceptibility

Table 3.1 Characteristics of the NOAA TIROS-N series satellite and the AVHRR instrument.

Parameter	Value
IFOV	1.4 milliradians
Ground Resolution	1.1 km at nadir
Scanning Rate	360 scans/sec
Nominal Altitude	833 km
Orbital Period	102 min
Inclination (w.r.t equator)	99°
Cross Track Scan	$\pm 55.4^\circ$
Swath Width	2700 km
Calibration:	
Channel 3,4,5	On board blackbody
Channel 1,2	Space view only

study is NOAA-11 with some NOAA-10 and NOAA-9 images having also been acquired. Graphs of the detailed spectral response of each AVHRR channel on NOAA-11 can be found in Figures 3.1a through 3.1c. The NOAA Polar Orbiter Data User's Guide (1988) gives similar graphs for the AVHRR for each satellite in the TIROS-N series. Though there is some variation in the channel responses of NOAA-9 through

NOAA-11, all spectral analysis (such as weightings for albedo or bidirectional reflectance) used the NOAA-11 curves as a standard. Superimposed on the first two graphs of Figure 3.1 is the solar flux density at the top of the atmosphere. Solar data from Neckel and Labs (1984) is used for channels 1 and 2 and that of Thekaekara et al. (1969) for channel 3.

3.2.2 AVHRR Data formats

The data acquired by the AVHRR is available in four formats. The real-time direct broadcast of a complete high resolution scan (representing about 2048 pixels and a ground swath of 2700 kilometers) to ground stations within view of the satellite is known as the high resolution picture transmission (HRPT) mode. The satellite is typically in view for about ten minutes and during that time can transmit over 2000 scans which would represent a ground path flight of about 3600 kilometers. The spatial resolution at nadir is 1.1 kilometers. HRPT data acquired by ground stations in Redwood City, CA, Wallops Island, VA, and Gilmore Creed, AK are sometime archived, depending on user requests, at the National Environmental Satellite, Data, and Information Service (NESDIS) center operated by NOAA. If this occurs, the raw HRPT data is assembled and archived into a level "1b" format where earth location data, solar zenith angle, and channel calibration information has been appended. This data format is described in the NOAA Polar Orbiter Data User's Guide (1988). Available scenes may be ordered on computer compatible tape from NESDIS. The local area coverage (LAC) mode is identical to the HRPT except the satellite stores up to 10

minutes of high resolution data over some part of the world as directed from

Table 3.2 The approximate spectral half-power bandwidth of the AVHRR instrument channels for the TIROS-N series. Based on data given in the NOAA Polar Orbiter Data User's Guide (1988).

AVHRR Channel Bandwidth (μm)				
Channel	Detector	TIROS-N	NOAA-6,-8,-10	NOAA-7,-9,-11
1	Silicon	.55-.90	.57-.68	.57-.69
2	Silicon	.715-1.02	.72-.98	.72-.98
3	InSb	3.58-3.96	3.56-3.95	3.53-3.93
4	HgCdTe	10.52-11.45	10.55-11.45	10.30-11.30
5	HgCdTe	---	---	11.50-12.50

ground. The recorded data is then broadcast to a NOAA ground station at the next pass. The recording of LAC data must be scheduled through NESDIS. The third data mode is a reduced resolution version referred to as global area coverage (GAC) which provides for complete earth coverage. This is recorded on the satellite and transmitted to a NOAA ground station. GAC data is obtained from the full resolution data by

averaging four of every five points along a scan and skipping the fifth. Further reduction is obtained by processing only one out of every three scan lines. This gives an effective resolution of about 4 km though it is actually about 4 km along the scan line and 1.1 km along track with a 3 km gap between the next along track pixel. A spatial resolution of 4 km is still useful in cloud remote sensing but care must be taken to insure that the spatial uniformity of the cloud being analyzed exceeds this value. The advantage of GAC data is that it offers complete global coverage and is continually archived into a level 1b format by NESDIS. There also exists an automatic picture transmission (APT) mode which broadcasts an analog signal at 4 km resolution to ground stations in range of the satellite.

3.2.3 AVHRR Channel Counts and Physical Quantities

The signal received at the channel detector is digitized into a 10 bit binary and is referred to as the channel *count*. This count is then an integer from 0 to 1023. The physical significance of the count depends on the channel. For the first two channels the count is linearly related to the integrated albedo (in percent) and in the notation of the NOAA Polar Orbiter Data User's Guide is described by the formula

$$A_j(\%) = s_j C_j + i_j \quad (3.1)$$

The subscript j is an index for channel 1 or 2, C is the count, s is the slope in units of percent albedo per count, and i is the intercept in percent albedo. The slope is often referred to as simply the *calibration coefficient* while some designate it with the symbol G (Abel 1990a) or γ (Holben and Kaufman 1991). Slope and intercepts are appended to the NOAA level 1b data. The albedo used in this context is actually the bidirectional reflectance and is explicitly given by

$$\frac{\int_{\lambda_j} \pi I_{ref}(\lambda) \phi_j(\lambda) d\lambda}{\int_{\lambda_j} \pi I_o(\lambda) \phi_j(\lambda) d\lambda} \bigg/ \frac{\int_{\lambda_j} \phi_j(\lambda) d\lambda}{\int_{\lambda_j} \phi_j(\lambda) d\lambda} \quad (3.2)$$

where I_o is the intensity of solar radiation reflected from a perfectly reflecting Lambertian diffuser and $\phi_j(\lambda)$ is the spectral response of channel j . In the denominator, $\pi I_o(\lambda)$ can be replaced by $\mu_o F_o(\lambda)$. Equation 3.1 is defined for $\mu_o = 1.0$ and so must be divided by the actual μ_o to correct for solar zenith angle. The solar zenith angle is appended to the level 1b data or may be calculated from solar ephemeris data. An alternate form of Equation 3.1, often used by investigators attempting to make in-flight calibrations of these two channels, is

$$A_j(\%) = s_j (C_j - C_{oj}) \quad (3.3)$$

where C_{oj} is the count detected when the detector views deep space (about 40 for NOAA-11). For purposes of in-flight calibration it is most convenient to have the output of channels 1 and 2 in terms of the actual intensity detected. This gives rise to

the equation

$$I_j = \alpha_j (C_j - C_{oj}) \quad (3.4)$$

where α_j is usually in units of $W/(m^2\text{-sr-}\mu\text{m-count})$. Some refer to α as the gain coefficient (Teillet et al. 1990, Whitlock et al. 1988) which is a confusing designation since α is inversely related to the actual gain of the sensor system. The reciprocal $1/\alpha$ (or $1/s$) is more indicative of the system gain and is used by Nianzeng et al. (1991). This definition of gain will be used throughout the remainder of the chapter. The connection between s and α in the last two equations is dependent on the incident solar flux density and is given by $s_j = 100\pi\alpha_j / \langle F_o \rangle_{\lambda_j}$, where the denominator denotes an average of solar radiation over the channel band, i.e.,

$$\langle F_o \rangle_{\lambda_j} = \frac{\int_{\lambda_j} F_o(\lambda) \phi_j(\lambda) d\lambda}{\int_{\lambda_j} \phi_j(\lambda) d\lambda} \quad (3.5)$$

Solar flux given by Neckel and Labs (1984) and averaged over $0.1 \mu\text{m}$ bands (see Rao 1987) gives $\langle F_o \rangle_{\lambda_j} = 1629.2$ and $1052.1 \text{ W}/(m^2\text{-}\mu\text{m})$ for channel 1 and 2 respectively. Equations 3.4 and 3.5 also have to be divided by μ_o to give the actual bidirectional reflectance.

The counts for channels 3, 4, and 5 of the AVHRR give the received signal in terms of intensity using a formula similar to Equation 3.1. Now, however, the slope value given by NOAA for the level 1b data format has units of $mW/(m^2\text{-sr-cm}^{-1}\text{-count})$ and the intercept units of $mW/(m^2\text{-sr-cm}^{-1})$. Channels 4 and 5, being in the thermal

infrared, are more usefully expressed in terms of the temperature inferred if the intensity is assumed to be emitted by a blackbody. This temperature is sometimes referred to as the *brightness temperature* (eg., Brown et al. 1985, Chamberlain and Hunten 1987).

The inverse Planck function in the appropriate form is

$$T_B = \frac{c_2 \nu}{\ln \left(1 + \frac{c_1 \nu^3}{I(\nu)} \right)} \quad (3.6)$$

with ν as the wavenumber (cm^{-1}), I the monochromatic detected intensity for a given wavenumber, T_B the brightness temperature, $c_1 = 1.1910659 \times 10^{-5} \text{ mW}/(\text{m}^2\text{-sr-cm}^{-4})$ and $c_2 = 1.438833 \text{ cm-K}$. Or, if given the brightness temperature,

$$I(\nu) = \frac{c_1 \nu^3}{e^{T_B/c_2 \nu} - 1} \quad (3.7)$$

But, the signal seen by a detector is the weighted average of intensity over the channel spectral response. That is the intensity at channel j 's detector is

$$I_j = \frac{\int_{\nu_j} I(\nu) \phi_j(\nu) d\nu}{\int_{\nu_j} \phi_j(\nu) d\nu} \quad (3.8)$$

Making the substitution of Equation 3.7 for $I(\nu)$ gives I_j as a spectral integration involving T_B . The determination of brightness temperature from a measurement of I_j is a simple type of inversion problem that can be solved by iteration or with a lookup table. (This is much easier than the classic inversion problem since both the kernel, Equation 3.7, and $\phi(\nu)$ are known; only the parameter T_B is unknown). NOAA has

avoided the need for this approach by determining a *central wavenumber*, ν_j^* , for each channel (3, 4 and 5) such that Equation 3.6 can be used directly without spectral integration. The central wavenumbers are determined for various temperature ranges to minimize the retrieved temperature error and generally give a temperature within 0.2 K (Walton 1987, appendix B) of that obtained with both Equation 3.6 and 3.7. That is, ν_j^* is chosen such that $I(\phi_j(\nu), T_B) - I(\nu_j^*, T_B)$ is a minimum over some specified temperature range. The NOAA Polar Orbiter Data User's Guide provides central wavenumbers for each satellite in the TIROS-N series.

3.2.4 Digitization Error

The signal resolution of each channel is ultimately limited by the 10-bit accuracy of the analog to digital conversion. For instance, referring to Equation 3.3, the actual bidirectional reflectance can be written as $s_n(C_n + \Delta C_n - C_0)$ where C_n is the n-bit count and ΔC_n is a necessary correction. Then the difference between the actual value and the n-bit value is $s_n \Delta C_n$. For channel 1, with $s_{10} \approx 0.11$ %/count and ΔC_{10} assumed to be a maximum of $\pm 1/2$ a bit, the absolute error in bidirectional reflectance is about ± 0.055 %. Once AVHRR channel data have been extracted from a tape it is useful to save the channel values as an 8-bit binary. This allows the data to be stored as a compact binary file and automatically puts it into a raster form which can later be used for graphically displaying the image. The original 10-bit count can be reduced to 8 bits by stripping off the two least significant bits (say, bits 9 and 10) and adding one to bit 8 if the resulting two bit number is 2 or greater. This is equivalent to dividing the 10-

bit integer by four and rounding to the nearest integer. To convert the 8-bit count to bidirectional reflectance, the original 10-bit s must be multiplied by four, ie., $s_{10} = 4s_8$. The $\pm \frac{1}{2}$ bit error in the 10-bit count becomes $\pm \frac{1}{8}$ of a bit, but an additional $\pm \frac{1}{2}$ bit error occurs from rounding the 8-bit count. So in the same notation as the previous example, $\Delta C_8 = \pm(\frac{1}{8} + \frac{1}{2})$ and $s_8 = 0.44$ implying that the absolute error in bidirectional reflectance is now $\pm 0.275\%$, five times the error for the 10-bit count. As will be shown in the next section, this digitization error is still much less than the calibration error, that is, error in the determination of s . It can be shown that for the stated assumptions, the total digitization error in going from an n -bit count to an $(n-i)$ -bit count is $s_n(2^i \Delta C_n + \frac{1}{2})$, which is $2.5s_{10}$ for the example above. To summarize, data for channels 1 and 2 is reduced to an 8-bit count, as described above, and then stored as a binary, or raster, image file. Then calculations of reflectance require a calibration coefficient s (or slope α , see Equation 3.4) four times greater than the original. It is worth noting that because of the linear relation between reflectance and count, converting the 10-bit count to reflectance first and then performing the 8-bit digitization results in the same error.

Channel 3 emission can be an important, and sometimes dominating, component of the total signal. To extract solar reflection this emission must be removed based on an estimate of cloud temperature made with a thermal infrared channel (4 and/or 5). It is for this reason that the channel 3 data values are best left in units of intensity ($\text{mW}/(\text{m}^2\text{-sr-cm}^{-1})$) as opposed to reflectance or temperature. The 10-bit count is reduced to 8 bits in the same manner as above. For a slope of about -0.0015, the 8-bit

digitization error becomes $\pm 0.00375 \text{ mW}/(\text{m}^2\text{-sr-cm}^{-1})$. As a comparison, the solar reflectance from an optically thick cloud with a drop radius of $20 \mu\text{m}$ gives an intensity of about $0.15 \text{ mW}/(\text{m}^2\text{-sr-cm}^{-1})$, or forty times the digitization error (specific examples will be given in chapter 4).

Data for channels 4 and 5 is best saved in terms of the temperature of a blackbody that would account for the received intensity. Because of the highly nonlinear relation between intensity (or count) and Planck temperature, it was found in this study that the smallest digitization error occurs when the 10-bit count is converted to temperature first and then this temperature value is stored as an 8-bit binary. The temperature was assumed bounded between 200 K and 320 K, giving the stored 8-bit integer as $(T-200)*255/120$ where T is the temperature in degrees Kelvin. The temperature digitization error for this range is $120*(\pm 1/2)/255 \approx 0.235 \text{ K}$. But an additional error is introduced in the initial conversion from intensity to temperature which translates into a separate temperature error having the dependence

$$\Delta T(\Delta I_{10}, T) = \frac{c_2 v^*}{\ln \left[1 + \frac{c_1 v^{*2}}{\frac{c_1 v^{*2}}{\exp\left(\frac{c_2 v^*}{T}\right) - 1} + \Delta I_{10}} \right]} \quad (3.9)$$

The digitization error in intensity is again 2.5 times the slope (approximately -0.175 for channels 4 and 5) giving $\Delta I_{10} = \pm 0.4375 \text{ mW}/(\text{m}^2\text{-sr-cm}^{-1})$. A plot of $\Delta T(\Delta I_{10}, T)$ versus temperature using this error is shown in Figure 3.2 and is seen to range from about 0.24 K at 320 K to 1.08 K at 200 K. A central wavenumber of 927.8 cm^{-1} (which is applicable for channel 4) has been used. Adding 0.235 K to the graph gives the total digitization error. So for warm clouds the total error might be expected to be about $\pm 0.53 \text{ K}$.

3.3 Channel Calibration

Channels 1 and 2 have no on board calibration and must, without in-flight calibration techniques, rely on calibrations typically performed several years before launch by the instrument manufacturer, ITT. For example, the NOAA-10 AVHRR was calibrated by ITT in July 1980 but not launched until September 1986. A detailed description of the pre-launch calibration for these two channels is given by Rao (1987). The goal is to determine a simple linear regression between the channel counts and bidirectional reflectance. This is achieved by illuminating the AVHRR with an

integrating sphere, coated with barium sulphate, which is assumed to approximate an isotropic source. Changing the number of quartz lamps used to illuminate the inside of the sphere provides varying intensity levels which establish the calibration curve. The NASA 30-inch Integrating Sphere Source was used by ITT. The connection between sphere intensity and bidirectional reflectance is made with some assumption of solar exoatmospheric spectral flux. NOAA has apparently used the data of Thekaekara (1974). Table 10 in Rao (1987) indicates, for example, that the use of Neckel and Labs (1984) solar spectral data for calculation of the channel integrated solar flux differs from the Thekaekara result by about 5.5% in channel 1 of the NOAA-10 AVHRR and 2.55% in channel 2. This discrepancy translates directly to the calibration slope values. Calibration of the sphere's spectral output and linearity is also of obvious importance. AVHRR NOAA-11 calibrations, included in level 1b data tapes, were based on a 1987 calibration of the sphere. However a 1989 calibration of the sphere showed some inconsistencies (Abel 1990a) and NOAA recommended a new set of slope and intercept values.

The accuracy of the pre-launch calibration is a moot point if the sensor response is modified, either while the instrument awaits launch or during flight. So in-flight calibrations seem to be the most reasonable way of obtaining credible sensor calibrations. The in-flight calibration involves satellite observations of scenes of known reflectance. Applying atmospheric corrections and adopting a data set for the exoatmospheric solar flux, AVHRR channel count can be compared with the calculated intensity and a relation of the form of Equation 3.4 can be established. The α found

in this manner can then be used to calculate calibration slope (s) through Equation 3.5. Slater et al. (1987) and Teillet et al. (1990) report on various approaches to this type of calibration. Nianzeng et al. (1991) used on-site surface radiance measurements at White Sands, New Mexico to calibrate the French SPOT satellite and then compared concurrent overpasses of SPOT and NOAA polar orbiters to infer AVHRR channel 1 and 2 calibration. Kaufman and Holben (1991) report on AVHRR calibrations over the ocean. They find the channel 1 calibration directly from atmospheric scattering. Molecular scattering is weaker in channel 2 so this calibration is inferred by assuming some knowledge of spectrally independent ocean glint. This method is based on the work by Fraser and Kaufman (1986) for the GOES VISSR instrument. Sensor degradation was found by monitoring reflectance from the Libyan desert. A NOAA workshop report on radiometric calibration (Abel 1990b) gives a summary of various in-flight techniques and estimated gains. Besides the works just mentioned, this report includes the cloud technique of Justus (1989) and aircraft measurements over White Sands by Guenther et al. (results first presented in the report). The calibration coefficients reported by these investigators are shown in Figures 3.3a and 3.3b for NOAA-11 AVHRR channel 1 and 2 respectively. The uncertainty bars in the graphs are shown as given by the investigators. The uncertainty analysis by Nianzeng et al. includes errors in SPOT calibration, SPOT and AVHRR image registration, atmospheric correction, and temporal and view angle changes between the satellites. Uncertainties of about 5-7% are estimated, greatly dominating any digitization error.

AVHRR channel 1 and 2 in-flight sensor degradation (meaning a decrease in the

gain, or an increase in the coefficients α or s) has been reported for NOAA-7, -9, -10, and -11 (Whitlock et al. 1988; Teillet et al. 1990; Holben and Kaufman 1991; Nianzeng et al. 1991). Teillet et al. analysis of NOAA-9 sensor degradation relative to the pre-launch calibration estimates an increase in the calibration coefficient of between 25-30% for channel 1 and about 35% for channel 2. This is valid up to May 1987 (NOAA-9 was launched in December 1984 with an estimated pre-launch calibration in February 1980). With less in-flight data available, NOAA-10 degradation is estimated by Teillet et al. to be 21% in channel 1 and 35% in channel 2 as of March 1987 (NOAA-10 was launched in September 1986). The results of Nianzeng et al. indicate that the NOAA-11 in-flight calibration coefficient has increased by 4% and 13% in channel 1 and 2 respectively during the period from November 1988 to June 1990.

All NOAA-11 channel 1 and 2 AVHRR data analyzed in chapter 7 use Equation 3.3 with calibration coefficients s based on the gains ($1/\alpha$) reported by Nianzeng et al. Equation 3.5 is applied using Neckel and Labs (1984) solar flux data to relate s to α . A deep space count of 40 is used in all cases. The in-flight calibration made closest in time to the recording of the data being studied is chosen. NOAA-9 and -10 AVHRR data is calibrated in the same way with the Teillet et al. gain values. Table 3.3 summarizes the calibration values used.

On board calibrations are available for channels 3, 4, and 5. As the AVHRR scan mirror rotates, the sensors view deep space and an internal target in addition to the earth scene. The internal target temperature is at about 288 K and monitored with platinum resistance thermometers (PRT's). But it is the target's brightness temperature

Table 3.3 Calibration coefficients used for extracting channel 1 and 2 AVHRR data. Pre-launch coefficients are included for comparison, with NOAA-11 being the revised values. References given in the text reported values for α ; S was found with Equation 3.5 applied to the appropriate channel.

Satellite	Source of Calibration and Date	Coefficient α [W/(m ² -sr- μ m)]/count		Coefficient s (% Albedo) /count	
		CH1	CH2	CH1	CH2
NOAA-11	Nianzeng et al. (June 90)	.5917	.4608	.1141	.1376
	Nianzeng et al. (Nov. 1988)	.5554	.4032	.1071	.1204
	NOAA Pre-launch	.4927	.3014	.095	.090
NOAA-10	Teillet et al. (March 1987)	.621	.465	.1197	.1389
	NOAA Pre-launch	.5115	.3454	.0986	.1031
NOAA-9	Teillet et al. (May 1987)	.670	.441	.1292	.1328
	NOAA Pre-Launch	.5243	.3286	.1011	.0990

(or intensity) that is required. This is found using a pre-launch procedure. The relation between channel count and intensity is established before launch with a vacuum calibrated laboratory blackbody run through a sequence of temperatures at three instrument operating temperatures. The brightness temperature of the internal target versus PRT values is then derived from the channel count and laboratory blackbody data. Deep space is at about 3 K and provides the second point for a linear calibration. The channel 3 InSb detector has been shown to be highly linear over the two point temperature range. Channel 4 and 5 use a HgCdTe detector which shows some nonlinearity. A complete description of the calibration procedure is provided in Planet (1988). Appendix B of that report gives corrections for the channel 4 and 5 nonlinearity up through NOAA-10. Using channel 4 on NOAA-9 as an example, for a cloud at 285 K (typical of marine stratus), the two point linear calibration underestimates actual cloud temperature by a maximum of 0.5 K. For a cloud at 275 K this maximum increases to 0.9 K. At 295 K the linear calibration overestimates by a maximum of 0.7 K. The nonlinear errors in channel 5 are about half as great. The accuracy of channel 3 is not specified. Errors in atmospheric correction (chapter 5) will normally dominate these nonlinearities. Also, chapter 6 will show that the remote sensing of susceptibility is typically not sensitive to variations in cloud top temperatures of this amount. The recognition that the two point in-flight calibration itself depends on the pre-flight relation between the internal target brightness temperature and PRT count remaining fixed throughout the satellite's lifetime doesn't appear to be addressed by NOAA. Finally, after recalling that the total digitization error is about ± 0.53 K it was felt that the two

point calibration without nonlinear corrections was sufficient. An analysis of the NOAA calibration is given by Brown et al. (1985).

3.4 Geometric Considerations

In general, bidirectional reflectance is dependent on the cosine of both the solar (μ_0) and satellite (μ) zenith angles, as well as the relative azimuth angle (ϕ) between them, as seen from each pixel in an image. The typical definition of relative azimuth for this type of radiative transfer problem is shown in Figure 3.4 with ϕ being a minimum when the satellite is in a position to be looking towards the sun, and approaching 180° when the satellite is looking in a direction such that the sun is at its back. The determination of each of these geometric variables will be discussed in this section.

The solar zenith angle, latitude, and longitude at every 40 pixels along a scan line of HRPT data and every 8 pixels for a scan of GAC data is included in the NOAA level 1b format. A linear interpolation applied to the other pixels is adequate since the reflectance function is slowly varying with respect to illumination and viewing angles. The time at the start of each scan line is also available. The latitude and longitude determined by NOAA is needed for calculations of μ and ϕ , and for a check of μ_0 . Direct calculations of pixel latitude and longitude were made with the code of Baldwin (1991 personal communication) using satellite ephemeris data for a NOAA-11 orbit (20 March 1989, approximately 2200 GMT) over a marine region off southern California which was available as HRPT data. The latitude calculated with the Baldwin code is

the *geodetic* (angle between the equatorial plane and the normal to the surface point being considered) while NOAA presumably gives the *geocentric* (usual definition). For an ellipsoidal earth with major axis a and minor axis b , the two are related by the expression $\theta = \tan^{-1}\{(b/a)^2 \tan\theta_{\text{geodetic}}\}$ where $(b/a)^2$ is approximately 0.99328 (see Bate et al. 1971). For example, at 20 and 40 degrees geocentric latitude, the geodetic latitude is 20.142 and 40.190 degrees respectively. Making use of this relation, comparisons of the Baldwin code with NOAA calculations for nadir view pixels showed differences in geodetic latitude of 0.04° (at about 20° latitude) to 0.12° (at about 40° latitude). The differences in longitude were 0.03° and 0.05° respectively. This corresponds to spacial differences of about 8-10 km. At extreme scans (pixels 1 and 2048 for HRPT) on a given line the spacial error went from 10-15 km for pixel 1 to 20-25 km for pixel 2048. Though it isn't clear which values are closer to being correct, these discrepancies can be used later to check the sensitivities of μ and ϕ to pixel location errors. Solar ephemeris calculations made for the same HRPT tape were found to be well within a degree of those solar zenith angles given by NOAA. The worst case difference was about 0.8° for $\theta_o = 50^\circ$ and $\theta_{\text{sat}} = 40^\circ$ and improving to about 0.04° for $\theta_o = 20^\circ$ and $\theta_{\text{sat}} = 20^\circ$.

3.4.1 Satellite Zenith Angle

The calculation of the satellite zenith angle is made with reference to Figure 3.5 which shows a section through the center of the earth that includes the points imaged by the satellite along on a scan. Along its sub-orbital track the satellite flies above the

sub-satellite point (the satellite's earth point) of what eventually becomes the complete scan line path at location θ_{se}, ϕ_{se} (using earth latitude and longitude as coordinates) and is shown at the moment when it is viewing the pixel location θ_e, ϕ_e . The satellite zenith angle of interest is the one local to θ_e, ϕ_e . The limits of the satellite scan are given by the X. Two vectors, $\bar{A}$ and $\bar{B}$ are indicated in the figure. Then

$$\mu_{sat} = \frac{\bar{A} \cdot \bar{B}}{|\bar{A}| |\bar{B}|} \quad (3.10)$$

where the vectors can be defined in cartesian coordinates as a function of $\theta_{se}, \phi_{se}, \theta_e, \phi_e$, and the height of the satellite h . The algebra of Equation 3.10 is straightforward, though lengthy and not given here. The sensitivity of μ to these variables is examined after the satellite azimuth is derived.

3.4.2 Relative Azimuth Angle

Figure 3.6 shows the complete three dimensional view of the satellite and sun positions. The satellite is shown viewing a point located away from nadir and the sun is to the southwest of this point. The local view to the sun and satellite as seen from this point define the zenith angles. Again, the sub-satellite point is assumed to lie at the midpoint of the scan path. Before a relative azimuth (not shown in this diagram but looks to be roughly 90°) can be calculated, an absolute azimuth must be defined and is referenced for both the sun and the satellite as being positive from the south towards the west. Using the same notation as before, a spherical triangle with vertices at

$(\theta_{se}, \phi_{se}), (\theta_e, \phi_e)$, and the north pole is shown in Figure 3.7. The angles of the triangle are denoted by uppercase letters and the length of the sides in degrees by lowercase. Referring to the diagram, the angle C is $\phi_{se} - \phi_e$ and angular lengths a and b are $(90 - \theta_{se})$ and $(90 - \theta_e)$ respectively. After A and B are found using Napier's analogies or rules, c can be found using the law of sines for spherical triangles (see Kells et al. 1940). This is a general method for obtaining the great circle distance between any two points on a sphere. Refer to the second triangle which includes the sub-satellite point, the point on the scan path, and the point on the equator at the same longitude as the scan path point. The angular length of side a' can be found in the same way as was c , and b' is just θ_e . Then ϕ_{sat} can be found from the half-angle formula. Special care must be paid to the signs of the angles throughout the calculations. Finally the solar azimuth, ϕ_o can be found from solar ephemeris data for the particular location and time corresponding to the scan path point. Then the relative azimuth becomes $\phi_{sat} - \phi_o - 180^\circ$.

3.4.3 Sensitivities of μ , ϕ to the Geometry

The azimuth ϕ_{sat} depends on the latitude and longitude of the sub-satellite and scan path points, while μ also depends on the height, h , of the satellite. The difference between calculations of satellite azimuth and zenith when using the NOAA values of latitude and longitude versus those obtained with the Baldwin code are listed in Table 3.4 in parenthesis. The NOAA values are also shown. Results for two different heights and two separate scan lines, corresponding to about 20° and 40° latitudes, are given.

The table shows that the difference of 10-30 km in the location of a pixel noted earlier between NOAA and Baldwin does not translate to significant differences in the calculations of μ and ϕ_{sat} . The calculation of μ depends on knowledge of the satellite height which may vary by 20 to 30 km because of the slightly elliptic orbit. The last column in the table shows the difference in μ occurring when calculations at $h=833$ km and 850 km are made with NOAA latitude and longitude numbers. Again, no significant difference is found.

Because of the apparent lack of sensitivity, calculation of the satellite's μ and ϕ_{sat} for a given pixel were made using NOAA level 1b latitude and longitudes, and a nominal satellite height of 833 km. Solar ephemeris calculations of μ_0 and ϕ_0 at the pixel were made with latitude, longitude, and scan line start times taken from the NOAA 1b data.

3.4.4 Ground Resolution

The ground resolution at nadir ($\mu_{\text{sat}}=1.0$) was noted in section 3.2 as being about 1.1 kilometer. Also known as the linear instantaneous field of view (having units of distance), it is calculated from the instantaneous field of view (IFOV, in radians) by

$$x = \text{IFOV} * h \quad . \quad (3.11)$$

with h typically 833 km and the *IFOV* about 1.4 milliradians, x is about 1.1 km. The linear IFOV for pixels at non-nadir viewing angles is increased because of the larger

Table 3.4 Sensitivity of satellite zenith (θ , μ) and azimuth angle (ϕ_{sat}) to changes in satellite height (h in kilometers), and latitude and longitude of sub-satellite and scan path points. The number in each cell is the value obtained when calculations were made with pixel locations taken directly from the NOAA level 1b data tape. The parenthetic number is the difference (absolute value) between this NOAA derived number and the calculation made with pixel locations obtained from the Baldwin code.

Scan Path Point	ϕ_{sat} (degrees)	h=833 km		h=850 km		$\Delta\mu$ (μ_{850} - μ_{833})
		θ (degrees)	μ	θ (degrees)	μ	
Scan Line #1 for HRPT Mar. 20,1989 (sub-satellite point at 20.025°, -128.641°)						
26	82.250 (.354)	67.25 (.14)	.3867 (.0022)	66.79 (.13)	.3941 (.0021)	.0074
500	82.044 (1.156)	33.31 (.01)	.8357 (.0001)	32.83 (.01)	.8403 (.0001)	.0046
750	81.192 (2.341)	17.32 (.05)	.9547 (.0002)	17.03 (.05)	.9562 (.0002)	.0015
2024	-103.76 (.290)	67.32 (.34)	.3856 (.0056)	66.85 (.35)	.3931 (.0056)	.0075
Scan Line #2000 for HRPT Mar. 20,1989 (sub-satellite point at 39.509°, -134.171°)						
26	89.044 (.226)	67.22 (.28)	.3871 (.0054)	66.76 (.28)	.3946 (.0044)	.0075
2024	-111.02 (.150)	67.26 (.52)	.3866 (.0085)	66.79 (.53)	.3940 (.0085)	.0074

distance to the satellite relative to a nadir pixel, and because the earth's curvature tilts the imaged area away from the satellite view. For the distance from the satellite to the earth point being $L(\mu_{sat})$, the resolution becomes

$$x(\mu_{sat}) = \frac{IFOV * L(\mu_{sat})}{\mu_{sat}} \quad (3.12)$$

When $\mu_{sat}=1.0$, $L(\mu_{sat})=h$ which is identical with Equation 3.11. In this study it has been found that

$$L^2(\mu_{sat}) = 2rh + h^2 + 2r^2\mu_{sat}^2 - 2r\mu_{sat}\sqrt{2rh + h^2 + r^2\mu_{sat}^2} \quad (3.13)$$

where r is the radius of the earth. This last expression can be used in Equation 3.12 to obtain the resolution at any value of μ_{sat} . It is useful to graph this result in terms of the increase in the linear IFOV at some arbitrary image point compared with the linear IFOV for a nadir point. The increase is by the factor

$$\frac{1}{\mu_{sat}} \frac{L(\mu_{sat})}{h} \quad (3.14)$$

which is plotted in Figure 3.8. The correspondence between μ_{sat} and pixel position along a scan line is also shown and is helpful for inferring linear IFOV at various locations in an image. For example, if an image pixel is known to have a μ_{sat} of 0.50 then the linear IFOV will be a factor of 3.5 greater than a nadir pixel, or about 3.85 km. This corresponds to a pixel located at a position number of about 105 (or 1944)

on the scan line. For comparison, two thirds of the image has pixels with $\mu_{sat} > 0.62$ (linear IFOV less than 2.9 km) and one half of the image has pixels with $\mu_{sat} > 0.85$ (linear IFOV less than 1.45 km).

(Fig. 3.1a)

(Fig. 3.1b)

(c)

Figure 3.1 Relative spectral response of the NOAA-11 AVHRR (a) channels 1 and 2 along with exoatmospheric solar flux density from Neckel and Labs (1984), (b) channel 3 with solar flux from Thekaekara et al. (1969), and (c) channels 4 and 5.

Figure 3.2 The temperature error for AVHRR channel 4 due to a digitization uncertainty in detected intensity of $0.4375 \text{ mW}/(\text{m}^2\text{-sr-cm}^{-1})$, using Equation 3.9 in the text.

(a)

(b)

Figure 3.3 In-flight calibration coefficients, α , and uncertainties as reported by various investigators for NOAA-11 (a) channel 1 and (b) channel 2. Derived by Nianzeng et al. using White Sands and the SPOT satellite, Holben and Kaufman using ocean scenes, Guenther with aircraft measurements over White Sands, and Justus using a cloud technique. References given in the text.

Figure 3.4 Definition of the relative azimuth angle between the satellite and the sun as seen by the point being imaged.

Figure 3.6 Geometry used in defining the absolute solar and satellite azimuth angle as seen by the point being imaged. Azimuth is zero looking to the south and increases positive towards the west.

Figure 3.7 Geometry used in calculating the absolute satellite azimuth angle as seen by the point being imaged (located at θ_e, ϕ_e). The sub-satellite point is at θ_{se}, ϕ_{se} . Azimuth is zero looking to the south and increases positive towards the west.

Figure 3.8 The solid line represents the ratio of the linear instantaneous field of view (IFOV) for a pixel with μ_{sat} compared with that for a pixel at nadir ($\mu_{sat}=1.0$). The dotted line gives the pixel number in an HRPT scan line corresponding to various μ_{sat} 's. Pixel numbers 1 and 2048 are the beginning and end respectively of a scan line and farthest from nadir. Pixel 1024 is at nadir. The arrows indicate the appropriate ordinate.

CHAPTER 4

RADIATIVE TRANSFER

Satellite remote sensing of clouds begins with the detection of radiation in various channels for some determined geometry. A procedure must be devised to determine that set of cloud parameters (effective drop radius and optical depth) which best correspond to the satellite measurement. Regardless of the procedure used, a capability for determining the radiative properties of clouds for the necessary wavelengths and geometries is needed. This chapter discusses these properties for a few of the AVHRR channels examined in the previous chapter. Mie parameters for cloud drops are given for a distribution of drop sizes. Using these results, a doubling method for calculating cloud reflectance and emittance is discussed. With this capability a library of data for a variety of cloud drop radii, optical thicknesses, and geometries is developed. This library will eventually be used for comparison with satellite measurements as a means for inferring cloud parameters; the specific method is discussed in detail in chapter 7. The chapter concludes with some results taken from this library.

4.1 Mie Calculations for the AVHRR Channels

The three optical parameters needed to determine cloud reflectance are optical thickness, τ , single scattering albedo, ϖ_0 , and the phase function, $p(\Theta)$ (though in many cases the asymmetry parameter (g) alone is sufficient). The extinction efficiency, Q_{ext} ,

is implicit in the optical thickness. Each of these parameters is a function of wavelength and drop radius. Since the cloud does not contain a single monodisperse drop radius, an appropriately weighted average over the size distribution is required. The AVHRR channels have a substantial bandpass so an additional average with the channel wavelength response is needed. These two averages applied to each optical parameter are necessary to determine reflectance (and other radiative properties) for a particular channel.

4.1.1 The Effect of Drop Size Distribution

The proper averages over drop size only (Hansen and Travis 1974, Twomey 1977a) are:

$$\begin{aligned}
 \tau &= l\pi \int Q_{ext}(r)r^2n(r)dr \\
 &= l\pi \frac{\int Q_{ext}(r)r^2n(r)dr}{\int r^2n(r)dr} \frac{\int r^2n(r)dr}{\int n(r)dr} \int n(r)dr \\
 &= \overline{Q_{ext}} \pi \overline{r^2} Nl
 \end{aligned} \tag{4.1}$$

$$\begin{aligned}
 \omega_o &= \frac{\int \omega_o(r) Q_{ext}r^2n(r)dr}{\int Q_{ext}r^2n(r)dr} \\
 &= \frac{\overline{Q_{sca}}}{\overline{Q_{ext}}}
 \end{aligned} \tag{4.2}$$

$$g = \frac{\int g(r) Q_{sca} r^2 n(r) dr}{\int Q_{sca} r^2 n(r) dr} \quad (4.3)$$

Equation 4.1 comes directly from the definition of optical thickness presented in chapter 2 and implies that a useful average for Q_{ext} (and for Q_{sca} and Q_{abs}) should have an r^2 weighting. The weighting for single scattering albedo should be according to the probability of encountering a given drop size and is the extinction coefficient, $k_{ext}(r)$, which is the optical path per unit physical length. k_{ext} is equivalent to the reciprocal of the photon mean free path and so represents the average number of drops, within a given radius interval, encountered per unit length by a photon. Another way of justifying this weighting is to note that the probability that a scattered photon encountered a drop with radius between r and $r+dr$ while traveling through a size distribution with total optical path τ given by equation 4.1 is $\tau(r)dr/\tau$ or $k_{ext}(r)dr/k_{ext}$. Equation 4.2 is sometimes called the *bulk* or *volume* single scattering albedo. The asymmetry parameter has a weighting similar to single scattering albedo but now only those photons that are scattered (i.e., not absorbed) should contribute to the average.

The parameters are seen to be explicitly dependent on the nature of the distribution, however in practice it has been found that a couple of moments of the distribution are usually sufficient to specify each. That is, different distributions with the same set of moments will result in similar optical parameters and therefore similar reflectances. For example, optical thickness is proportional only to the second moment of radius when Q_{ext} has little variation over the radius interval, while liquid water

content (equation 2.1) goes with the third moment. Twomey and Cocks (1989) showed that when absorption can be approximated for a single drop radius by $1-\omega_0 \propto r$ (valid for weak absorption, see Twomey and Bohren 1980), then for a distribution of sizes, the approximation becomes

$$1-\omega_0 \propto \frac{\int Q_{ext} r^3 n(r) dr}{\int Q_{ext} r^2 n(r) dr} \approx \frac{\overline{r^3}}{\overline{r^2}} \quad (4.4)$$

implying that ω_0 is primarily dependent on the ratio of the third moment to the second. This ratio has been termed the *effective radius*, r_{eff} . Hansen and Travis (1974) showed with numerical examples that the phase function (and therefore the asymmetry parameter) has its major dependence on r_{eff} and not on the details of the distribution, while use of a mean radius (first moment) for several distributions gave significant spread. Since reflectance depends on τ , ω_0 , and g it appears that either r_{rms} or r_{eff} are two candidates which, if common to different distributions, would result in almost similar reflectances (or other radiative quantities). For narrow size distributions there is little difference between the two radii. The discussion by Twomey and Cocks and the results of Hansen and Travis suggest that r_{eff} be used. Another moment indicating the spread of the distribution is still needed according to Hansen and Travis. They defined the spread in terms of an *effective variance*

$$v_{eff} = \frac{1}{r_{eff}^2} \frac{\int (r-r_{eff})^2 r^2 n(r) dr}{\int r^2 n(r) dr} \quad (4.5)$$

where the r_{eff}^2 in the denominator is used to normalize v_{eff} to a dimensionless

quantity. Distributions with equivalent r_{eff} and v_{eff} were found to have almost identical phase functions. Presumably, the same is true of ω_0 though this isn't explicitly shown in their paper. The significance of this conclusion is critical to remote sensing. It implies that only two parameters of the distribution need to be known to build up a library of reflectance data. So if a nominal v_{eff} can be assumed, then r_{eff} can be inferred from measurements. However, the Hansen and Travis data, and apparently their conclusion, is based only on cases where $\omega_0 = 1$ and so not readily extendable to the AVHRR channel 3 wavelengths. The discussion by Twomey and Cocks for ω_0 is valid only for weak absorption, also not strictly true for channel 3.

The dependence of the optical parameters on r_{eff} and v_{eff} for channel 3 (and channel 1 for comparison) is presented in Table 4.1a and 4.1b for three different r_{eff} and two v_{eff} . Parameters were averaged over wavelength according to equation 4.10. Three types of size distributions were used - normal, log-normal, and a modified gamma. It was first necessary to relate the specific parameters of each distribution to r_{eff} and v_{eff} (see Table 4.2) so that common *effective* values could be specified for each calculation of ω_0 , g , and Q_{ext} . With the normal distribution, $\langle r \rangle$ is the mean radius; $\sigma/\langle r \rangle$ (known as the dispersion) was solved iteratively to give the correct v_{eff} , then $\langle r \rangle$ was determined from r_{eff} . Referring to Table 4.1, for $v_{\text{eff}} = 0.09$, $\sigma/\langle r \rangle = 0.462$ and $r_{\text{eff}}/\langle r \rangle = 1.35$, and for $v_{\text{eff}} = 0.032$, $\sigma/\langle r \rangle = 0.20$ and $r_{\text{eff}}/\langle r \rangle = 1.08$. The differences in ω_0 and g among the three distributions for AVHRR channel 3 appear to be too small to significantly alter the reflectance that would be calculated by each. The extinction efficiency is used in scaling the optical thickness by the factor

$2/Q_{\text{ext}}$. Variations in this factor can be determined from the table and appear to be less than 1% except for $r_{\text{eff}}=4.0$, $v_{\text{eff}}=0.09$. It looks as if the effective radius and spread of the distribution is sufficient for specifying the optical properties of a cloud drop size distribution, even in the relatively large absorption region of the AVHRR channel 3. The actual cloud size distribution is unknown, but calculations can be made with any reasonable distribution. In this study, all Mie calculations were averaged over a normal distribution with a dispersion of 0.20. ($r_{\text{eff}}=1.08\langle r \rangle$, $v_{\text{eff}}=0.032$) at various mean radii. When satellite data are analyzed, if a cloud is inferred as having a mean radius of $10\text{ }\mu\text{m}$ for a normal distribution, then the actual distribution will be assumed to have an $r_{\text{eff}}=10.8\text{ }\mu\text{m}$. The choice of dispersion (or v_{eff}) must be based on measurements and was chosen as 0.20 to represent a typical narrow distribution found in non-precipitating clouds. However, from Table 4.1, the dominating influence on the Mie parameters appears to be r_{eff} , with v_{eff} having a much lesser effect.

4.1.2 Monochromatic Examples

The dispersion in the size distribution serves to smear out much of the details of the Mie parameters when shown versus size parameter (or radius). Plots of g , ω_0 , and Q_{ext} versus mean radius for a normal distribution with a dispersion of 0.20 are given in Figures 4.1, 4.2, and 4.3 respectively for AVHRR channels 1, 3, and 4. Calculations were made for five wavelengths in each channel (see legends). Major differences with wavelength occur in all parameters and significant higher order detail remains especially in the asymmetry parameter.

Table 4.1a Optical parameters averaged over wavelength for AVHRR channel 3, calculated for three drop size distributions, each with the same r_{eff} and v_{eff} . Numbers 1, 2, and 3 correspond to normal, gamma, and log-normal distributions respectively (see Table 4.2). Calculations for the log-normal distribution were not made for $v_{\text{eff}}=0.032$.

AVHRR Channel 3: Distribution 1=Normal 2=Gamma 3=LogNormal									
$r_{\text{eff}}=4.0$			$r_{\text{eff}}=11.41$			$r_{\text{eff}}=20.0$			
$v_{\text{eff}}=0.09$			$v_{\text{eff}}=0.09$			$v_{\text{eff}}=0.09$			
	1	2	3	1	2	3	1	2	3
ω_o	.9569	.9580	.9588	.8639	.8643	.8644	.8009	.8013	.8013
g	.7841	.7891	.7923	.8217	.8232	.8244	.8678	.8693	.8705
Q_{ext}	3.104	3.175	3.229	2.305	2.314	2.317	2.206	2.211	2.209
$v_{\text{eff}}=0.032$			$v_{\text{eff}}=0.032$			$v_{\text{eff}}=0.032$			
	1	2	3	1	2	3	1	2	3
ω_o	.9595	.9598	---	.8617	.8614	---	.7977	.7989	---
g	.7925	.7983	---	.8332	.8328	---	.8711	.8728	---
Q_{ext}	3.355	3.360	---	2.289	2.302	---	2.196	2.200	---

Table 4.1b Optical parameters averaged over wavelength for AVHRR channel 1, calculated for three drop size distributions, each with the same r_{eff} and v_{eff} . Numbers 1, 2, and 3 correspond to normal, gamma, and log-normal distributions respectively (see Table 4.2). Calculations for the log-normal distribution were not made for $v_{\text{eff}}=0.032$.

AVHRR Channel 1: Distribution 1=Normal 2=Gamma 3=LogNormal									
$r_{\text{eff}}=4.0$			$r_{\text{eff}}=11.41$			$r_{\text{eff}}=20.0$			
$v_{\text{eff}}=0.09$			$v_{\text{eff}}=0.09$			$v_{\text{eff}}=0.09$			
1	2	3	1	2	3	1	2	3	
g	.8364	.8323	.8324	.8659	.8659	.8661	.8712	.8722	.8721
Q_{ext}	2.197	2.201	2.201	2.108	2.098	2.098	2.063	2.068	2.064
$v_{\text{eff}}=0.032$			$v_{\text{eff}}=0.032$			$v_{\text{eff}}=0.032$			
1	2	3	1	2	3	1	2	3	
g	.8432	.8309	---	.8661	.8673	---	.8725	.8725	---
Q_{ext}	2.175	2.198	---	2.085	2.084	---	2.056	2.068	---

Table 4.2 Distributions used for the analysis of Table 4.1.

Gamma Distribution:

$$n(r) = r^{\frac{1-3b}{b}} e^{-\frac{r}{ab}}$$

$$a = r_{eff}$$

$$b = v_{eff}$$

Log-Normal Distribution:

$$n(r) = \frac{1}{\sqrt{2\pi}\sigma_g r} e^{-\frac{(\ln r - \ln r_g)^2}{2\sigma_g^2}}$$

$$r_g = \frac{r_{eff}}{(1+v_{eff})^{5/2}}$$

$$\sigma_g = \sqrt{\ln(1+v_{eff})}$$

Normal Distribution:

$$n(r) = \frac{1}{\sqrt{2\pi}\sigma} e^{-\frac{(r-\langle r \rangle)^2}{2\sigma^2}}$$

$$r_{eff} = \langle r \rangle \frac{3\left(\frac{\sigma}{\langle r \rangle}\right)^2 + 1}{\left(\frac{\sigma}{\langle r \rangle}\right)^2 + 1}$$

$$v_{eff} = \frac{3\left(\frac{\sigma}{\langle r \rangle}\right)^6 + \left(\frac{\sigma}{\langle r \rangle}\right)}{\left[3\left(\frac{\sigma}{\langle r \rangle}\right)^2 + 1\right]^2}$$

Single scattering albedo is shown in terms of drop absorption, $1 - \omega_0$. The dotted line gives the approximation $1 - \omega_0 = 0.85k_B r$ (Twomey and Bohren 1980) using the bulk absorption coefficient value at $3.75 \mu\text{m}$. For a radius greater than about $5 \mu\text{m}$, absorption in channel 3 has an approximately $r^{1/2}$ dependence. The approximation is better for less absorbing, shorter wavelengths. The square root of absorption is proportional to reflectance from an optically thick layer (Twomey and Bohren 1980). So at larger radii, channel 3 cloud reflectance will have less sensitivity to drop size than shorter wavelengths used in remote sensing studies.

The single scattering albedo given for channel 3 is based on liquid water absorption only. Water vapor continuum absorption is present in this window and can absorb photons traveling the interstitial cloud space. Water vapor concentrations in this region would be approximately equal to the saturation vapor pressure at the local cloud temperature. The total optical thickness seen by a photon over some length l is

$$\tau = [\pi r^2 N (Q_{sca} + Q_{abs}) + k_v] l \quad (4.6)$$

where k_v is the vapor volume absorption coefficient, and extinction efficiency has been expressed as the sum of scattering and absorption efficiencies. With $\omega_0 = Q_{sca}/Q_{ext}$, equation 4.6 becomes

$$\begin{aligned}
\tau &= \pi r^2 N Q_{sca} \left[\frac{1}{\varpi_o} + \frac{k_v}{\pi r^2 N Q_{sca}} \right] \\
&= \pi r^2 N Q_{sca} \left[\frac{1}{\varpi_o} + \frac{k_v}{\frac{3W}{4\rho_l r} Q_{sca}} \right] \\
&= \pi r^2 N Q_{sca} \frac{1}{\varpi'_o} .
\end{aligned} \tag{4.7}$$

W is the liquid water content and replaces drop concentration in terms of radius. The symbol ϖ'_o can be interpreted as the effective single scattering albedo for drops in the absence of vapor that would account for the true situation of both liquid and vapor absorption. This effective albedo is

$$\varpi'_o = \frac{\varpi_o}{1 + \varpi_o \frac{4k_v \rho_l r}{3W Q_{sca}}} . \tag{4.8}$$

Across the wavelength interval covered by channel 3, k_v ranges over several orders of magnitude, so a single wavelength substitution into equation 4.8 can be misleading. The distribution of vapor absorption within the channel 3 band, as described by the k -distribution (Arking and Grossman 1972), was calculated from the AFCRL 1978 line parameter compilation using code written by Mergenthaler (private communication) and is displayed in Figure 4.4. The solid line indicates the cumulative distribution. The broken line gives the probability density function calculated from the cumulative curve. The effective single scattering albedo over the channel 3 band is an average with the

probability density function as a weighting, i.e.,

$$\omega'_0 = \frac{\int p(k_v) \omega'_0(k_v) dk_v}{\int p(k_v) dk_v} \quad (4.9)$$

Calculations of equation 4.9 for four radii are given in Table 4.3 below and the small differences between ω_0 and ω'_0 indicate that vapor absorption can be ignored at these wavelengths. The AVHRR channel response was not used in these calculations.

Table 4.3 The effective single scattering albedo (ω'_0) using equation 4.9 for a k-distribution derived for wavelengths between 3.6 and 3.9 μm at a pressure of 950 mb, $T=285$ K, and a vapor concentration equivalent to 100% RH.

$r(\mu\text{m})$	Q_{sca}	ω_0	ω'_0
1	.7838	.96282	.96251
4	2.879	.95083	.95050
20	1.718	.78677	.78495
30	1.558	.72821	.72634

4.1.3 Averages Over Wavelength

Averages weighted with the channel spectral response were approximated using

five equally spaced wavelengths within the band according to

$$\frac{\sum_{i=1}^5 f_i \phi(\lambda_i)}{\sum_{i=1}^5 \phi(\lambda_i)} \quad (4.10)$$

where f is the quantity being spectrally averaged. Table 4.4 gives the chosen wavelengths and corresponding channel response. Figures 4.5 and 4.6 show results for the asymmetry parameter and single scattering albedo respectively versus the mean radius of a normal distribution. Three different dispersions were used. If the abscissa was converted to effective radius, the curves would be shifted to the right with the larger dispersions being displaced the most. All three curves would then approximately overlap, indicating again that it is primarily effective radius, not effective variance (or dispersion), that is needed in specifying a size distribution. The spectrally averaged optical parameters must be used with care. In general the reflectance (or other radiative quantity) calculated from the average parameters is not equivalent to the average of the reflectances calculated for each wavelength, i.e.,

$$R(\overline{g}, \overline{\omega}, \overline{\tau}) \neq \frac{\sum_{i=1}^5 R(g_{\lambda_i}, \omega_{\lambda_i}, \tau_{\lambda_i}) \phi(\lambda_i)}{\sum_{i=1}^5 \phi(\lambda_i)} \quad (4.11)$$

The average bars indicate parameters averaged according to equation 4.10. As discussed in section 4.3, the spectrally averaged phase function for each radius and channel is used in determining reflectance at each wavelength through scaling (see

section 2.1.3), then equation 4.10 is applied for the spectrally averaged reflectance.

4.2 The Doubling Code

Suppose the reflectance and transmittance is known completely for some layer.

Using the layer's scattering properties to determine the properties of an identical layer with twice the optical thickness as the original is known as a *doubling* method. The

Table 4.4 The choice of λ and $\phi(\lambda)$ used in the band average calculations of equation 4.10. The channel spectral response is given for NOAA-11 using ITT data for channel 2, and NOAA Polar Orbiter Data User's Guide plots for channels 1, 3, and 4.

Wavelengths and Channel Response used in Calculating Band Averages							
CH 1		CH 2		CH 3		CH 4	
$\lambda(\mu\text{m})$	$\phi(\lambda)$	$\lambda(\mu\text{m})$	$\phi(\lambda)$	$\lambda(\mu\text{m})$	$\phi(\lambda)$	$\lambda(\mu\text{m})$	$\phi(\lambda)$
.56	.1628	.72	.6350	3.55	.7117	10.35	.5579
.60	.8000	.78	.9100	3.65	.9761	10.55	1.000
.64	.8838	.85	.8250	3.75	.9622	10.75	.9595
.68	1.000	.92	.7740	3.85	.9904	10.95	.9053
.72	.0838	.98	.6220	3.95	.2511	11.15	.7605

general method of superimposing two different layers is termed *adding* or *superposition*. The technique was separately developed by Twomey, Jacobowitz, and Howell (1966) using a matrix technique and by van de Hulst (1963). Application to cloud layers was given by Twomey et al. (1967). Extensions for polarization were made by Howell and Jacobowitz (1970, 1971), for absorption by Twomey (1975), and for emission by Twomey (1979). The matrix form of this method is couched in a notation convenient for numerical computation and is used in this work.

4.2.1 Azimuthally Independent Formulation

Consider the intensity at some level in a cloud layer to be spatially distributed as a function of μ (cosine of the layer's zenith) and ϕ (the azimuthal angle), and given by an N-dimensional vector with elements representing the intensity averaged into N latitude belts. In this study each hemisphere is divided into $N=10$ uniform bins with a width of $\Delta\mu=0.10$. The bins are designated as numbers 1,2,...,10 with the 10th bin containing the hemisphere's zenith ($0.90 \leq \mu \leq 1.0$) and the 1st bin including the horizon ($0 \leq \mu \leq 0.10$). The absolute value of μ is used so that any reference to up and down depends on the hemisphere being considered. The azimuthal dependence of the intensity can be expanded into a Fourier series as shown in the next section. For the moment only the azimuthally independent terms are developed. This is equivalent to an integration over ϕ and is related to the net flux in a latitude bin. Though simpler to construct than the full azimuthal solution, it is nevertheless a practical form since the net flux is often all that is required (e.g., a calculation of albedo). And in many

multiple scattering problems azimuthal variations are expected to be small near zenith.

If a downward propagating intensity vector, v_o , is incident at the top of a cloud then the reflected intensity, u (an *upward* traveling vector), and transmitted intensity, v (a *downward* vector), can be expressed in terms of the matrix equations

$$\begin{aligned} u &= S v_o \\ v &= (T + E) v_o \end{aligned} \quad (4.12)$$

where the matrices (capitalized variables) have dimension $N \times N$ (10×10). S and T represent the scattering properties of the layer for diffusely reflected and transmitted intensities respectively. The matrix E is diagonal and gives the directly transmitted intensity that passes through the layer without being scattered. For a layer with thickness τ , the diagonal elements of E are given by Beer's law, $\exp(-\tau/\mu_j)$, $j=1,2,\dots,10$. Diffuse refers to radiation scattered one or more times. As an example, the element $S_{i,j}v_{oj}$ gives the intensity reflected back (towards the opposite hemisphere) into the i th μ -bin, μ_i , due to an incident intensity from within the j th μ -bin, μ_j , having magnitude v_{oj} . $T_{i,j}v_{oj}$ has an identical interpretation but intensity is scattered into the same hemisphere as the incident beam. The reflection and transmission matrices will depend on the layer's optical thickness, single scattering albedo, and phase function.

Suppose that the three matrices for two different layers are known and indicated by S_1, T_1, E_1 and S_2, T_2, E_2 . Then the matrices S, T, E for the combined layer are calculated as a superposition of the intensities shown schematically in Figure 4.7.

Multiple reflections at the interface of the two layers give

$$S = S_1 + (T_1 + E_1)S_2(I + S_1S_2 + [S_1S_2]^2 + \dots) \times (T_1 + E_1) \quad (4.13)$$

$$E + T = (T_2 + E_2)(I + S_1S_2 + [S_1S_2]^2 + \dots)(T_1 + E_1)$$

where I is the identity matrix and $E = E_1E_2$. The reflected intensity of the combined layer is now Sv_0 . The infinite series converges for finite layers or non-conservative scattering and can be approximated by a finite number of terms. As n increases, the term $[S_1S_2]^{n+1}$ can eventually be approximated by $\lambda [S_1S_2]^n$ where λ is the largest eigenvalue of the matrix in brackets (Lanczos 1988). Then the infinite series becomes

$$\sum_{k=0}^{\infty} [S_1S_2]^k = \sum_{k=0}^{n-1} [S_1S_2]^k + [S_1S_2]^n \sum_{l=0}^{\infty} \lambda^l. \quad (4.14)$$

The infinite summation of λ^l is $(1-\lambda)^{-1}$. If the two layers are identical then the application of equation 4.13 is considered *doubling*. When they are different the term *adding* or *superposition* is used. A non-zero surface reflectance can be characterized in equation 4.13 by a second layer with T_2 and E_2 equal to zero. If the surface is considered Lambertian (a uniform reflected intensity pattern) with albedo α , then the elements of $S_{2i,j}$ are $\alpha\mu_j/5$ for all i , valid for $\Delta\mu=0.10$.

The reflectance for a homogeneous layer of any thickness can be built up by consecutive application of doubling and/or superposition formula. But the scattering matrices for a starting layer must be known to begin the process. An infinitesimal layer, with single scattering only, can be used as a *seed* since its scattering matrices

have a simple analytic form. In this study, a layer with some specified optical parameters and an optical thickness $\tau=2^{-12}$ serves as the seed. Then 12 doublings are required to build up to $\tau=1$. Successive superposition of the $\tau=1$ layer gives reflectance (or transmission or emission) for $\tau=1,2,3,\dots$. The scattering from an infinitesimal layer is given by

$$\begin{aligned} S(\Delta\tau) &= M^{-1} B \Delta\tau \\ T(\Delta\tau) &= M^{-1} P \Delta\tau \end{aligned} \quad (4.15)$$

The matrix M^{-1} is diagonal with elements μ_k^{-1} . The B matrix is derived from the phase function such that the i,j element gives the fraction of incident energy scattered out of the μ_j bin into the μ_i bin of the opposite hemisphere by a *single* drop. P is identical except it gives fractional energy scattered into the μ_i bin of the same hemisphere. Note that the angle μ is still defined with respect to the plane parallel layer and is not the drop's scattering angle which depends on the incident direction. Elements of the B and P matrices are found from

$$\begin{aligned} B_{ij} &= \frac{\int_{\Delta\mu_j} \int_{\Delta\mu_i} p^{(0)}(\mu, \mu') d\mu d\mu'}{\Delta\mu_j} \\ P_{ij} &= \frac{\int_{\Delta\mu_j} \int_{\Delta\mu_i} p^{(0)}(-\mu, \mu') d\mu d\mu'}{\Delta\mu_j} \end{aligned} \quad (4.16)$$

The term $p^{(0)}(\pm\mu, \mu')$ is the zero order Fourier expansion of the phase function and represents an average over azimuth, i.e.,

$$p^{(0)}(\pm\mu, \mu') = \frac{\int_0^{2\pi} p(\pm\mu\mu' + \sqrt{1-\mu^2}\sqrt{1-\mu'^2}\cos\phi) d\phi}{2\pi} \quad (4.17)$$

4.2.2 Azimuthally Dependent Formulation

Azimuthal variations in the reflected intensity are determined by expanding the phase function into the Fourier cosine series

$$p(\mu, \mu', \phi) = \sum_{m=0}^{\infty} p^{(m)}(\mu, \mu') \cos(m\phi) \quad (4.18)$$

The angle ϕ is defined as in Figure 3.4. Likewise, the B (and P) elements can be expanded as

$$B_{ij}(\phi) = \sum_{m=0}^{\infty} B_{ij}^{(m)} \cos(m\phi) \quad (4.19)$$

where $B^{(m)}$ can be found from equation 4.16 with $p^{(m)}(\mu, \mu')$ used in place of $p^{(0)}(\mu, \mu')$. However, the approach adopted in this work was to apply equation 4.19 directly to

$$B_{ij}(\Delta\phi) = \frac{\int_{\Delta\phi} \int_{\Delta\mu_j} \int_{\Delta\mu_i} p(\mu, \mu', \phi - \phi') d\mu d\mu' d\phi}{\Delta\mu_j} \quad (4.20)$$

where $\Delta\phi$ is an azimuthal angle bin. Then by definition, the scattering elements for an infinitesimal layer are

$$S_{ij}(\Delta\tau, \phi) = \sum_{m=0}^M S^{(m)}_{ij}(\Delta\tau) \cos(m\phi) \quad (4.21)$$

$$S^{(m)}_{ij}(\Delta\tau) = \frac{\Delta\tau}{\mu_i} B^{(m)}_{ij}$$

and similarly for $T_{\Delta\tau}$. All orders of directly transmitted radiation are represented by the same E matrix, that is, Beer's law is valid for each harmonic. The superposition formulae of equation 4.13 apply to all harmonics of the scattering matrices; finite layers are built up in the same manner as described in the last section.

The intensity vectors must also be expanded in a Fourier series. Then, for example, the reflection vector u can be expressed as

$$u = \sum_{m=0}^{\infty} u^{(m)} \cos(m\phi) \quad (4.22)$$

$$= \left[\sum_{m=0}^{\infty} S^{(m)} \cos(m\phi) \right] v_o .$$

The incident vector can be expanded in $(\phi - \phi')$ and substituted into the last equation. Integrating over ϕ' and summing over all μ_j gives the following for the harmonic reflection vector:

$$u^{(m)} = f_m S^{(m)} v^{(m)} \quad (4.23)$$

$$f_m = \begin{cases} T & m=0 \\ T/2 & m>0 \end{cases} .$$

T is the period and equal to 2π if the Fourier calculations are made with respect to angle. This equation can be simplified when the incident intensity is confined to a

single bin, say μ_j , and to a single ϕ value such as is the case for a plane-parallel source like the sun. If the flux into bin μ_j is to be F_0 , then the element v_j is represented by a delta function with a Fourier expansion given by $v^{(m)}_j = 2v^{(0)}_j$, where $v^{(0)}_j = F_0/2\pi\Delta\mu_j = 5F_0/\pi$ for $\Delta\mu_j = 0.10$. The product of the S matrix with this incident vector sifts out the j th row of the matrix. Finally, for $\Delta\mu_j = 0.10$, the bidirectional reflectance becomes

$$R_{i,j}(\phi) = \left(\frac{5}{\mu_j}\right) f_o S_{ij}^{(0)} + \sum_{m=1}^{\infty} \left(\frac{10}{\mu_j}\right) f_m S_{ij}^{(m)} \cos(m\phi) . \quad (4.24)$$

where $R_{i,j}$ is the inferred albedo for reflected intensity measured in bin μ_i and at azimuth ϕ due to an incident flux in bin μ_j . The actual albedo, with $\Delta\mu_i = 0.10$ as well, found from summing over all μ_i is

$$A = \frac{\sum_{i=1}^N \mu_i f_o S_{ij}^{(0)}}{\mu_j} . \quad (4.25)$$

4.2.3 Computational Checks

Several constraints act on the single scattering matrices B and P , and on S , T , and a . From reciprocity, the i,j value of B and P should be equal to the j,i value for all harmonics. Energy conservation requires the combined sum of any row or column from both zero order matrices to be unity (or ϖ_0 if the phase function is not normalize to unity). To maintain symmetry, only the diagonal elements are modified for energy conservation. Calculations of B and P , subject to the above constraints, were carried

out with equation 4.19 and 4.20, and independently with a Monte Carlo code. Elements of the zero order matrices generally agreed to better than 1%. Back directions, e.g., $B^{(0)}_{10,10}$, often differed by as much as 5%. Bidirectional reflectance comparisons for 8 μm and 25 μm drop radii for the AVHRR channel 3 were made. For both radii, reflectance calculations derived from the Monte Carlo B and P showed higher order structure than the same calculation using the analytically derived matrices. Excellent agreement occurred when a small number of harmonics were used in the Fourier series, but worsened with increasing harmonics. The Monte Carlo technique is inherently noisier because of the large number of photons needed to smooth out the statistics. This noise most likely affected the higher order matrices. Another check on the numerical integration of equation 4.20, and subsequent Fourier calculation of 4.19, can be made with the Henyey-Greenstein phase function (see van de Hulst 1980). Analytic, with a functional dependence on the asymmetry parameter, it has a relatively simple expansion in Legendre functions (see Whitaker and Watson 1962, p. 332). Applying the addition theorem puts the phase function in the form of a Fourier series with coefficients defined by associated Legendre functions. Equation 4.16 can be applied directly. Comparison with the direct calculation of B and P for the Henyey-Greenstein phase function showed better than 1% agreement for the zero order matrices, and typically a few percent at $m=2$. Higher order comparisons weren't made.

Several constraints are made on the scattering matrices. A symmetry condition is $[MS]^T = MS$ and $[MT]^T = MT$ (Twomey et al. 1966) providing a check at each level of the superpositioning sequence. Also, absorption can be calculated directly (Twomey

1975) and compared with an inference based on S and T .

4.2.4 Emission

Emission in the AVHRR channel 3 must be removed to determine the solar reflection component of the radiation. This can be determined with the matrix technique discussed above. Three components to the emission are shown in Figure 4.8. The cloud is assumed to be homogeneous and of uniform temperature. The direct upward emission is

$$u_1 = a B_{\lambda,c} \quad (4.26)$$

where a is the absorption vector (Twomey 1975) giving the net fractional absorption of the cloud layer for incident radiation in the various μ -bins. This can be calculated from an adding method or determined as one minus the sum of the net reflected flux (albedo) and the transmitted flux. By virtue of Kirchoff's law, this is equivalent to the layer's emissivity. Both vectors are functions of optical thickness, radius, and wavelength (emission is azimuthally symmetric so only zero order matrices are needed). $B_{\lambda,c}$ is the Planck function at cloud temperature. The downward component of cloud emission undergoes multiple reflections with the surface, contributing at cloud top

$$u_2 = (T_c + E_c) S_s (I + [S_c S_s] + [S_c S_s]^2 + \dots) a B_{\lambda,c} . \quad (4.27)$$

The subscripts s and c refer to surface and cloud respectively. Finally the direct surface component, for lambertian emission, is

$$u_3 = (T_c + E_c) (I + [S_s S_c] + [S_s S_c]^2 + \dots) e_s e B_{\lambda, s} \quad (4.28)$$

where e is a vector with unity for all elements and ϵ_s is the surface emissivity. All three terms can be combined in the form

$$\begin{aligned} u_{emit} &= u_1 + u_2 + u_3 \\ &= e_c B_{\lambda, c} + e_s B_{\lambda, s} \end{aligned} \quad (4.29)$$

The cloud and surface are seen to have effective emissivity vectors e_c and e_s respectively. Emissivities for AVHRR thermal channels can be determined in the same way.

4.3 The Library Files

The calculation of reflectance or emission, given cloud optical thickness and radius, represents the *forward* problem and was described in the last section. Inferring optical thickness and drop radius from satellite reflection and emission data characterizes an *inverse* problem. The inverse problem is solved in this study by determining the best fit between the satellite measurement and various entries in a library file. The library contains bidirectional reflectance for AVHRR channels 1 and 2, and with a slight modification for channel 3. Effective cloud and surface emissivities are kept for channels 3, 4, and 5. The contents of the reflection libraries include calculations for all combinations of the following: $r=1, 4, 6, 8, 10, 12.5, 15, 17.5, 20, 25, 30, 35, 45 \mu\text{m}$; $\tau=1, 2, \dots, 60$ for channels 3, 4, 5 and $\tau=1, 2, \dots, 50, 60, \dots, 150$ for channels 1 and 2; μ_i bins (satellite zenith angle) = 0.55, 0.65, ..., 0.95;

μ_j bins (solar zenith angle) = 0.25, 0.35,..., 0.95; harmonics from zero order up to twenty. The surface (assumed to be the ocean for this study) is Lambertian with albedos of 0.06, 0.03, 0.01, 0.0, and 0.0 for AVHRR channel 1 through 5 respectively. Mie parameters were calculated using a normal distribution with a dispersion of 0.20 ($r_{\text{eff}} = 1.08 \langle r \rangle$). The emission libraries don't contain harmonics greater than the zero order or solar zenith angles. In a sense, the inverse problem is solved by doing the forward problem many times.

For each radius and channel, the B and P matrices are determined from the spectrally averaged phase function according to equation 4.10. This does not incorporate other related weightings such as $Q_s(\lambda)$ (see equation 4.3), or the solar spectral incident flux for channels 1, 2, and 3. Such inclusions would not likely change the detailed structure of the averaged phase function. The reflectance or emittance at each of the five wavelengths listed in Table 4.3 is then calculated from this B and P by scaling (equation 2.5) with the appropriate asymmetry parameters. The shape of the azimuthal variation in reflection follows closely the shape of the phase function used to calculate the B and P . This is because single scattered photons show the greatest azimuthal variation; photons scattered more than once will soon forget their original incident azimuth. And the reflection of single scattered photons can be calculated directly from the phase function and the appropriate scattering angles. So while scaling can be appropriate for azimuth independent reflection, care must be taken otherwise.

For channel 1, with large size parameters, there is little change in the phase function over much of the wavelength and radius range, and scaling might successfully

be applied to all radii and wavelengths with a B and P derived from a single radius. Channel 1 reflectance calculations for $\mu_i = \mu_j = 0.75$ scaled with matrices derived from the band averaged phase function for a radius of 1, 4, 8, and 30 μm showed little discrepancy in the azimuthal solution. Over most azimuth angles the difference was less than 0.02 (absolute reflectance). But near an angle of 180° (back scattering) the differences rapidly increased to as much as 0.10. A similar analysis for channel 3, having smaller size parameters and substantial absorption, showed more significant differences. Now differences of about 0.05 are typical, representing a greater deviation relative to the small azimuthally averaged reflectance occurring in this channel. The azimuthal variation for single scattered photons was shown to account for over half the variation seen in the complete solution. But for a given radius, scaling from the spectrally averaged phase function gave almost indistinguishable results from the calculation using a separate B and P for each of the five wavelengths in the band.

Each combination in the library is an average over the channel's spectral response. For bidirectional reflectance (channels 1 and 2) this includes a spectral weighting for the incident solar flux approximated by

$$R(r, \tau, \phi) = \frac{\sum_{k=1}^5 \Phi(\lambda_k) \frac{F_{o, \lambda_k}}{\langle F_o \rangle} R(r, \tau, \phi, \lambda_k)}{\sum_{k=1}^5 \Phi(\lambda_k) \frac{F_{o, \lambda_k}}{\langle F_o \rangle}} \quad (4.30)$$

where $R_{i,j}(r, \tau, \phi, \lambda)$ is given in equation 4.24 and $\Phi(\lambda)$ is the appropriate channel response. The term $\langle F_o \rangle$ is any convenient normalization such as the average solar

flux in the band. Note that Equation 4.30 is not the spectrally averaged reflectance, which the satellite can't measure, but the net reflected flux, which the satellite does measure, divided by the net incoming solar flux. Channel 3 gives the average band intensity ($\text{mW}/(\text{m}^2\text{-sr-cm}^{-1})$) and for the reflected term is

$$I_{ij}^R(r, \tau, \phi) = \frac{\sum_{k=1}^5 I_{ij}^R(r, \tau, \phi, \lambda_k) \Phi(\lambda_k)}{\sum_{k=1}^5 \Phi(\lambda_k)} \quad (4.31)$$

$$= \frac{\mu_j \langle F_o \rangle}{\pi} \left[\frac{\sum_{k=1}^5 \frac{F_{o, \lambda_k}}{\langle F_o \rangle} \Phi(\lambda_k) R_{ij}(r, \tau, \phi, \lambda_k)}{\sum_{k=1}^5 \Phi(\lambda_k)} \right]$$

where reflected spectral intensity, with superscript R , given by $\mu_j F_o(\lambda) R_{ij}(r, \tau, \phi, \lambda) / \pi$ has been used. The value in brackets, which is nearly equivalent to bidirectional reflectance, is kept in the library. Figures 4.9a and 4.9b give the azimuthally averaged bidirectional reflectance, at six different radii, versus optical thickness for channel 1 and 3 respectively (the bracketed quantity of equation 4.31 is used for channel 3). Figures 4.10a and 4.10b give the azimuthal solution (average reflectance has been subtracted) for channel 1 and 3 for several thicknesses and radii. Up to 20 harmonics are stored in the library but typically only 10 are needed for either channel. The magnitude of successive harmonics fall off more slowly with channel 1 compared with channel 3, though the zero order is much larger for all but the smallest thicknesses. Accounting for the azimuthal variation becomes more critical in channel 3 with its small zero order

values.

With reflection, the incident solar spectral response is known and the integration of over wavelength can be calculated and placed in the library before any actual satellite data are processed. For the effective emissivities, equation 4.29, the spectral response over channel 3 of the Planck function is unknown until the cloud and surface temperatures of the scene are determined. For channels 3, 4 and 5, the satellite measures in units of average intensity; this is the first part of equation 4.31 with intensity replaced by the $B(T, \lambda)\epsilon(\lambda)$. The temperature is unknown apriori and so the emission cannot be taken out of the summation. However, for thermal channel 4 and 5, the Planck function across the band varies by less than 3% from the mid-band value for a temperature of 300 K, and by less than 0.5% at 273 K. So for these channels the mid-band value of the Planck function can be used for all wavelengths in the summation. But, the Planck function has a strong wavelength dependence across the channel 3 band, varying by as much as 100% from mid-band values. However, the ratio of Planck intensity at any wavelength in the band to the mid-band value is almost identical for all temperatures from 273 K to 300 K. The ratio curve for 283 K was applied to all five wavelengths used in the summation to give the following average emitted intensity:

$$B(T, \lambda_{k=3}) \left[\frac{\sum_{k=1}^5 \left(\frac{B(283K, \lambda_k)}{B(283K, \lambda_3)} \right) \epsilon(\lambda_k) \Phi(\lambda_k)}{\sum_{k=1}^5 \Phi(\lambda_k)} \right] .$$

Emission has been removed from the summation and the term in brackets is stored in the library. The relative contributions of solar reflectance and thermal emission to the total cloud top intensity for channel 3 is shown in Figure 4.11. For radii less than about $10\text{ }\mu\text{m}$, solar reflection dominates the signal. At radii greater than this, emission dominates. The ratio displayed in the figure is analogous to a signal to noise ratio, emission being akin to noise, which decreases with increasing drop size. The ability to extract the channel 3 reflected signal is necessary for remote sensing drop radius, and for that channel, accurate knowledge of emission becomes increasingly important at larger radii. To determine this emission in channel 3, one or both of the AVHRR thermal channels is needed to infer temperatures. As shown in Figure 4.12, the effective cloud and surface emissivities in channel 4 must be considered; for most radii, clouds with $(2/Q_{\text{ext}})\tau < 5$ are obviously not *black*. To solve for cloud temperature the surface temperature must be somehow determined (e.g., from neighboring clear sky pixels) and the optical thickness must be known.

(a)

(b)

Figure 4.1 Asymmetry parameter versus mean radius for a normal drop size distribution with a dispersion of .20, for (a) AVHRR channel 1 and (b) channel 3.

(a)

(b)

Figure 4.2 Absorption versus mean radius for a normal drop size distribution with a dispersion of .20, for (a) AVHRR channel 3 and (b) channel 4.

Figure 4.3 Extinction efficiency versus mean radius for a normal drop size distribution with a dispersion of .20, for (a) AVHRR channel 1, (b) channel 3, and (c) channel 4.

Figure 4.4 Cumulative distribution for vapor volume absorption coefficient, k_v , for wavelengths between 3.6 and 3.9 μm . Valid for $p=950 \text{ mb}$, $T=285 \text{ K}$, and a vapor concentration equivalent to 100% RH. The probability density function is calculated from the cumulative distribution.

Figure 4.5 Spectral average of asymmetry parameter versus mean radius for a normal drop size distribution using three dispersions, for (a) AVHRR channel 1, (b) channel 3, and (c) channel 4.

(a)

(b)

Figure 4.6 Spectrally averaged single scattering albedo versus mean radius for a normal drop size distribution using three dispersions, for (a) AVHRR channel 3 and (b) channel 4.

Figure 4.7 Schematic diagram for the individual components making up the net reflected and transmitted fields. Capitalized variables indicate matrices, small ones indicate vectors (from Mergenthaler 1981).

Emission components for cloud at uniform temp.

Figure 4.8 Three components to the net emission from both the cloud and the surface. The cloud is considered to be at a uniform temperature T_c .

(a)

(b)

Figure 4.9 The azimuthally averaged bidirectional reflectance for various drop radii for (a) AVHRR channel 1 and (b) channel 3. The solar and satellite zenith angles are indicated.

(a)

(b)

Figure 4.10 The azimuthally dependant bidirectional reflectance for various drop radii for (a) AVHRR channel 1 and (b) channel 3. The solar and satellite zenith angles, and optical thicknesses are indicated.

Figure 4.11 The ratio of reflected solar intensity to thermal emission for AVHRR channel 3. The calculation was made for both surface and cloud temperatures of 290 K.

Figure 4.12 The scene temperature inferred with AVHRR channel 4 when the surface is at 290 K and the cloud at 283 K. Asymptotic temperatures are slightly less than 283 K because the cloud emissivity has a limit of about .99.

CHAPTER 5

ATMOSPHERIC EFFECTS ON REMOTE SENSING

As discussed in previous chapters, the remote sensing of cloud drop radius and optical thickness, and the subsequent calculation of susceptibility, involves reflectance measurements in the non-absorbing and absorbing part of the spectrum. With the AVHRR instrument, the first two channels can be considered non-absorbing. The absorbing channel, channel 3, is centered at $3.75 \mu\text{m}$ where emitted radiation is comparable with, and can often exceed, reflected solar radiation from a cloud. This emission which is determined by surface and cloud temperatures (and emissivities) must be removed to resolve the reflected signal. So use of the AVHRR presupposes the use of one or both of the thermal IR channels (4 and 5) for inferring temperature. Consequently, all five channels of the AVHRR contain useful information.

This chapter examines the effect of the atmosphere on the signal received by each channel. The possible atmospheric effects are shown schematically in Figure 5.1, with the heavy lines indicating the primary effects: transmission of emitted and of direct and reflected solar radiation (important in all five AVHRR channels), solar radiation reflected from the atmosphere without ever encountering the surface (channels 1,2, and 3), and direct emission from the atmosphere (channels 3,4 and 5). Channel 3 is unique in that each of these influences must be considered. Multiple reflections between the surface or cloud and the atmosphere are ignored. The consequence of these dominant effects on the net signal received by the satellite can be expressed in the

following equations with the understanding that the atmospheric terms depend on the height of the object being viewed (i.e., the amount of intervening atmosphere). It is assumed that this object is opaque to radiation and so the portion of the atmosphere between, say, the cloud and the ocean does not contribute to the signal.

For scattering only (channels 1 and 2) with R_{sat} as the bidirectional reflectance seen by the satellite, R_s the actual reflectance of the scene in question (e.g., ocean or cloud), R_{atm} the atmospheric reflectance, and t the transmittance:

$$R_{sat}(\mu_o, \mu, \phi) = R_s(\mu_o, \mu, \phi)t(\mu_o)t(\mu) + R_{atm}(\mu_o, \mu, \phi) \quad (5.1)$$

For emission only (channels 4 and 5) with I_{sat} as the intensity seen by the satellite, I_s the actual emission from the scene in question, I_{atm} the atmospheric emission:

$$I_{sat}(\mu) = I_s(\mu)t(\mu) + I_{atm}(\mu) \quad (5.2)$$

For channel 3 where the atmosphere and the object both scatter solar radiation and emit radiation:

$$\begin{aligned} I_{sat}(\mu_o, \mu, \phi) = & I_{s_{scat}}(\mu_o, \mu, \phi)t(\mu_o)t(\mu) + I_{s_{emit}}(\mu)t(\mu) \\ & + I_{atm_{scat}}(\mu_o, \mu, \phi) + I_{atm_{emit}}(\mu) \end{aligned} \quad (5.3)$$

$I_{s,scat}$ is solar scattering off the scene in the absence of an atmosphere and can be expressed in terms of bidirectional reflectance as

$$I_{s_{\text{can}}}(\mu_o, \mu, \phi) = \frac{\mu_o}{\pi} \frac{\int_{\lambda} R_s(\mu_o, \mu, \phi) F_o(\lambda) \phi(\lambda) d\lambda}{\int_{\lambda} \phi(\lambda) d\lambda} .$$

F_o is the exo-atmospheric solar radiation. For channel 3 this calculates to approximately $(16.271 R_s \mu_o)/\pi$.

Each of these primary effects has been modeled using the LOWTRAN7 radiation code (Kneizys et al., 1988; for a review see Fenn et al., chapter 18 in Jursa 1985) for various standard atmospheres. Results are discussed for a sea level surface where the entire atmosphere is involved and for clouds whose tops are assumed to be at the top of the boundary layer. For channels 1 and 2, the reflection of diffuse transmission and multiple scattering in the atmosphere can be significant for optical thicknesses typically expected to be encountered. This is examined using the complete multiple scattering radiative transfer code of Herman and Browning (1965). The chapter ends with quantitative relations between uncertainty in the retrieved channel signal resulting from uncertainty in the knowledge of these atmospheric effects.

5.1 Lowtran Radiative Transfer Code

The Lowtran code was originally developed at the Air Force Cambridge Research Lab (now Air Force Geophysics Lab) as a "low" resolution, fast predictor for atmospheric transmittance (Selby and McClatchey 1972). This prediction scheme was developed by McClatchey et al. (1972) as an extension of the graphical technique developed by Altshuler (1961). A brief explanation of the method follows.

Given that scattering and absorption coefficients of the various atmospheric species are known at any arbitrary wavelength, monochromatic transmission is a straightforward application of Beer's exponential law. Over a finite bandwidth however, the average transmittance becomes a sum (or integration) of exponentials and generally can no longer be described by a single exponential. An exception would be when attenuation is slowly varying with wavelength such as with molecular scattering, aerosol extinction, and molecular continuum absorption. Molecular line absorption on the other hand is highly varying and depends on the location of the spectral lines. A complete calculation of average transmittance for this latter case would involve "line-by-line" calculations including all relevant spectral line locations and parameters (i.e., line intensity, half-widths, temperature, pressure, etc.). To speed up the computation, the technique of McClatchey et al. calculated averages based on an empirical transmittance function of the form

$$\overline{\tau(\nu)}_{\Delta\nu} = f(C(\nu) \omega^*) \quad (5.4)$$

where $C(\nu)$ is the absorption coefficient at wavenumber ν , $\Delta\nu$ is the bandwidth over which the average is valid, and ω^* is the equivalent absorber amount scaled with standard temperature and pressure as

$$\omega^* = \Delta L \left(\frac{P(z)}{P_o} \sqrt{\frac{T_o}{T(z)}} \right)^n. \quad (5.5)$$

ΔL is the equivalent length of pure absorber in the path (i.e., cm-atm). This form of the transmittance function is similar to that obtained from band models, such as those

of Elsasser and Goody, when reduced to the strong line approximation. For $n=0$ the transmittance function gives the weak line (exponential) approximation and the strong line for $n=1$. Using laboratory measurements of transmittance and knowledge of spectral lines, the function f and the parameter n are empirically determined separately for the uniformly mixed gases (i.e., constant mixing ratio with height - O_2 , CO_2 , N_2O , CH_4 , and CO), water vapor, and ozone. Coefficients for molecular scattering, aerosol attenuation, and continuum absorption are slowly varying and described empirically; the average transmittance functions are simple exponentials. The total average transmission is the product of each of these individual average transmittances. In all cases, the averaging bandwidth used, $\Delta\nu$, is 20 cm^{-1} , with steps in ν of 5 cm^{-1} .

The notion of a single empirical transmittance function (f) valid for all equivalent absorber amounts seems unlikely. In fact, the determination of f represents a mean that best fits the relation between $\tau_{\Delta\nu}(\Delta\nu)$ and $C(\nu)\omega^*$ for each ν and a range of path lengths. This concept of a mean, which is essentially an average over frequency, obviously depends on the weighting given to the various frequencies (or $C(\nu)$'s) and the path length over which the mean is to be valid. The transmittance functions developed by McClatchey et al. and later used in LOWTRAN were weighted for "moderate" path lengths (the precise meaning of moderate is not specified in the literature). Longer path lengths will be given a higher transmittance than they should and shorter paths less transmittance. No explicit mention of the $C(\nu)$ weighting is given in the literature. However, McClatchey et al. state that the greatest error might be in the distant wings

of strong absorption lines, presumably a weak absorption region. The overall accuracy of this method is said to be better than 10%, with an implicit assumption being that all data and measurements concerning spectral lines, continuum absorption, etc., are known and accurate.

In LOWTRAN the atmosphere (assumed horizontally homogeneous) is stratified into 33 layers, with species concentrations based on either user input or standard models. Aerosol size distribution and complex indices of refraction are specified with various models in four height layers. The integrated species amount along a given path includes atmospheric refraction. Later editions of LOWTRAN added emission (Selby et al. 1978), single scattering (Kneizys et al. 1983), and multiple scattering (Kneizys et al. 1988). The multiple scattering algorithm was not used in this study. Comparisons of LOWTRAN predictions with measurements are given in each of the Air Force Geophysics Lab publications. A summary is presented by Jursa (1985).

LOWTRAN results for atmospheric transmission, emission, and scattering are presented for each of the AVHRR channels. A built in, user specified, filter program was used to calculate band integrated values. Two standard atmospheres were used - the mid latitude summer (MLS) and winter (MLW) models (see Table 5.1). Temperature, pressure, and water vapor (to 10 km) are based on the supplemental standard atmospheres given by Cole et al. in the Handbook of Geophysics and Space Environments (Valley 1965). Ozone distributions are taken from the same source. Above 10 km, water vapor profiles are given by Sissenwine et al. (1968). The MLS atmosphere is probably most appropriate to the marine environment being studied. The

Table 5.1a The first 11 levels of the midlatitude summer (MLS) model atmosphere used by LOWTRAN, from McClatchey (1972).

MIDLATITUDE SUMMER				
Height (km)	P (mb)	T (K)	Mixing Ratio	RH(%)
0	1013	294	11.86	76
1	902	290	8.69	64
2	802	285	6.08	55
3	710	279	3.74	45
4	628	273	2.38	31
5	554	267	1.39	31
6	487	261	.94	31
7	426	255	.63	31
8	372	248	.40	31
9	324	242	.26	31
10	281	235	.15	30

Table 5.1b The first 11 levels of the midlatitude winter (MLW) model atmosphere used by LOWTRAN, from McClatchey (1972).

MIDLATITUDE WINTER				
Height (km)	P (mb)	T (K)	Mixing Ratio	RH(%)
0	1018	272	2.20	76
1	897	269	2.16	70
2	790	265	1.74	65
3	694	262	1.30	55
4	608	256	.80	50
5	531	250	.51	47
6	463	244	.32	45
7	402	238	.14	31
8	347	232	.07	23
9	299	226	.035	20
10	257	220	.018	10

Navy Maritime Aerosol Model (Gathman 1983, Kneizys et al. 1983) based on measurements is included in LOWTRAN and was used for these studies. The model defines three components to the marine boundary layer aerosol - the fresh marine component (related to current surface wind speed), stationary marine component (consisting of smaller particles and related to past wind speeds), and the continental component (gas to particle). The model size distributions and optical properties are modified by weightings of these components and humidity and visibility. Default values, based on the model atmosphere, were used. This model is valid only in the boundary layer which is placed at 2 km in LOWTRAN. Between 2 and 10 km a tropospheric model is used, and above this level a stratospheric model.

5.2 Atmospheric Transmission

Transmission predictions were made for each of the five AVHRR channels for solar zenith angles at 0° , 30° , and 60° ($\mu_0 = 1, .87$, and $.50$), mid latitude summer and winter atmospheres with the navy maritime aerosol model, and for paths from sea level and 2 km to the top of the atmosphere. Examples of transmission for MLS, vertical path from sea level to space are shown in Figures 5.2a-5.2e indicating the spectral transmission for major attenuators superimposed on the AVHRR channel response. Unfortunately, as a general rule, the largest contributions to attenuation come from the most variable species. For channel 1, aerosol scattering is the largest component to attenuation and because of its variable nature is also a cause of large uncertainty. For example, band integrated transmission varies from 0.795 (Fig. 5.2a) to 0.662 when

switching to a LOWTRAN 23 km visibility maritime aerosol model. In channel 2 (Fig. 5.2b), aerosol scattering is still strong but attenuation may be dominated by water vapor absorption, also highly variable with results to be discussed later (refer ahead to Figure 5.3). Transmission varies from 0.761 to 0.644 when changing aerosol models as mentioned above. Strong molecular oxygen absorption is present at about $0.76\ \mu\text{m}$. In channel 3 (Fig. 5.2c), water vapor line absorption is the greater contributor with some water vapor continuum absorption and a small amount of N_2 continuum absorption. Aerosol scattering is significant as well, with transmission going from 0.735 to 0.687 when changing aerosol models. Both channels 4 and 5 (Fig. 5.2d,e) are dominated by water vapor continuum absorption with some line absorption. A theoretical description for water vapor continuum absorption is incomplete with possible sources being the extreme wings of absorption lines, dimers (Bignell 1970, Grassl 1973), and charged H_2O clusters (Carlson and Harden 1980). Calculations from known spectral line parameters and models are not sufficient to explain the absorption. LOWTRAN (also line-by-line compilations, Rothman et al. 1987) use empirical formulations based on measurements by Burch et al. (1971) which include self and foreign broadening terms. A review is given by Burch (1982). The thermal channels suffer from this uncertainty associated with the continuum in addition to variability in water vapor amount. A line-by-line study of transmission in the AVHRR channels was done by Saunders and Edwards (1989) and provides a useful check of LOWTRAN. They report results for all major absorbing gas species and aerosols (modeled with a profile recommended by the World Climate Research Programme (WCRP), 1986) with

0.05 cm^{-1} resolution. A comparison of LOWTRAN with their results for the U. S. Standard Atmosphere is given in Table 5.2. Agreement is within a percent for most of the gaseous absorption but with a major disagreement for channel 2 and 3 water vapor lines. In both cases LOWTRAN shows greater absorption. The discrepancy in aerosol attenuation is most likely due to the different maritime models used. The table also shows results from Saunders and Edwards using an urban aerosol model, pointing out extreme differences in transmission compared with the maritime aerosol. A customary view found in the literature is that atmospheric moisture uncertainty precludes the use of channel 2 for accurate reflection measurements, implying an advantage to the use of channel 1. However, this conventional wisdom seems to ignore the significant uncertainty that aerosol variability produces. Looking ahead to data presented throughout the rest of the chapter, LOWTRAN comparisons of variability in the two channels do not give a meaningful edge to either.

Figures 5.3a-5.3e summarize LOWTRAN transmission results for each channel as a function of zenith angle (solar or satellite). Symbols are defined in the legend. The solid lines give a quadratic fit to the average of the MLS and MLW data for each path (sea level to space and 2 km to space). The modeled profiles of moisture and aerosol show less seasonal variability above the boundary layer and this is reflected in the plots. The seasonal variability in channel 1 is due to the season dependent defaults used by the navy maritime aerosol model. Above the 2 km boundary layer the aerosol model shows no seasonal dependence and the MLS and MLW transmissions are practically identical. The same applies to channel 2 though now moisture variability is

Table 5.2 Transmittance for the AVHRR channels using U.S. Standard atmosphere profiles for vertical path, sea level to space. LT is Lowtran7 using the navy maritime aerosol model, S&E is Saunders and Edwards (1989) line-by-line calculation with the WCRP standard maritime aerosol profile.

AVHRR Channel	1		2		3		4		5	
Contributor	S&E	LT	S&E	LT	S&E	LT	S&E	LT	S&E	LT
H ₂ O Lines	1.0	.98	.96	.92	.92	.88	.97	.96	.94	.93
H ₂ O continuum	1.0	1.0	1.0	.99	.98	.98	.93	.93	.90	.90
Mixed Gases	1.0	.99	.98	.99	.95	.94	.98	.98	.99	.98
O ₃	.97	.98	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0
Rayleigh Scattering	.94	.95	.98	.98	1.0	1.0	1.0	1.0	1.0	1.0
Aerosol-Maritime	.93	.89	.94	.91	.97	.97	.99	.99	.99	.99
Aerosol-Urban	.42		.56		.93		.97		.97	
Total- Maritime	.85	.80	.86	.79	.83	.78	.87	.86	.82	.80
Total- Urban	.38		.52		.80		.85		.80	

a factor. Note that the overall variability is less than that of channel 1. In channels 4 and 5 the variability is exclusively from the seasonal moisture profiles and is quite large for the sea level paths. Channel 3 primarily shows moisture variability with some aerosol sensitivity. As expected the functional dependence on zenith angle for all channels is smooth and generally follows the form $\exp(-\tau/\mu)$. For the 2 km to space path, the effective optical thickness is small and seasonal variability dominates the variation with μ . In this case a single value of transmittance could be used in atmospheric correction. With the assumption that marine stratus tops lie near the top of the boundary layer, this is the path which is of primary interest in this study. While the aerosol model defines the boundary layer to be at 2 km, this is not true of the MLS and MLW profiles which have little similarity to typical marine boundary layer inversion profiles. So equating the 2 km path with the top of a marine boundary layer is approximate at best. A LOWTRAN emission calculation using a user supplied marine sounding is presented later in this chapter.

5.3 Atmospheric Scattering

For single scattering and no emission, the radiative transfer equation for a given wavelength reduces to

$$\mu \frac{dI(\tau, \mu, \phi)}{d\tau} = \omega_o F_o(\tau, \mu_o, \phi_o) p(\cos\Theta) \quad (5.6)$$

where F_o is the incoming solar flux at a level τ within the atmosphere, I is the intensity of scattered radiation in the direction μ, ϕ coming from the differential layer at τ , and

Θ is the scattering angle. The monochromatic intensity at the top of the atmosphere ($\tau=0$) is given by an integration over τ and is

$$I(0, \mu, \phi) = \int_0^\tau F_o(0, \mu_o, \phi_o) p(\cos \Theta) e^{-\tau'(\frac{1}{\mu} + \frac{1}{\mu_o})} \frac{d\tau'}{\mu} . \quad (5.7)$$

This is the equation solved in LOWTRAN using previously discussed transmittance functions for the exponential term. Phase functions for aerosol and molecules are calculated separately and then combined with a weighting corresponding to the scattering coefficients of each. Standard phase functions corresponding to various aerosol models are stored in LOWTRAN and suitably interpolated to provide an average aerosol phase function. The incoming solar flux at the top of the atmosphere is given by Thekaekara (1974) (data by Neckel and Labs (1984) is used for channels 1 and 2 elsewhere in this study). Comparisons of the LOWTRAN implementation with complete multiple scattering code is given by Kneizys et al. (1983).

The LOWTRAN single scattering code was run for channels 1, 2, and 3 of the AVHRR for zero surface albedo, MLS and MLW atmospheres, and the navy maritime aerosol model as a function of satellite zenith angle. The solar zenith angle was 45° with a few runs done for 0° and 60° . Two azimuth angles were considered, 0° and 180° . In keeping with the format of the AVHRR data, results for the first two channels are displayed as bidirectional reflectance, i.e.,

$$\frac{\int_{\lambda} \pi I_{reg}(\lambda) \phi(\lambda) d\lambda}{\mu_o \int_{\lambda} F_o(\lambda) \phi(\lambda) d\lambda} \quad (5.8)$$

The wavelength integration is over the appropriate AVHRR channel and $\phi(\lambda)$ is the channel response function (In the LOWTRAN filter code, output the numerator of eq. 5.5, excluding π , is referred to as the 'sensor weighted integrated radiance'. The denominator must be calculated separately and was found to give, using solar spectral data of Neckel and Labs (1984) for channel 1 and 2 and Thekaekara (1974) for channel 3, to be 184.0, 241.1, and 4.52 W/m²). The geometric dependencies are not shown. For channel 3 the results are in both intensity units and reflectance. A graphical summary is presented in Figure 5.4a-5.4c. Again, the solid lines represent a quadratic fit to the average of the MLS and MLW results at each azimuth. For the 2 km to space path, there was little difference (typically less than a few percent) between the MLS and MLW atmospheres; only the MLS is shown. The larger reflectance for $\phi=180^\circ$ (antisolar direction) is caused by enhanced scattering in the backwards direction. Exact backscattering for this geometry occurs for $\phi=180^\circ$ and a satellite zenith angle of 45° . These extra points are shown in the channel 1 plot and indicate only a slight peaking, suggesting that the averaging of the aerosol phase functions have reduced the extremes and so the functional dependence on satellite zenith angle will be relatively smooth (excluding the forward scattering peak at large solar zenith angles). The reflectance in channel 1 is about twice that of channel 2 and the maximum variability including azimuth and season is roughly 3.2% reflectance for channel 1 and 1.2% for channel 2.

Calculations of albedo versus satellite zenith angle by Masuda et al. (1988) for a clear atmosphere over the ocean compare well with these plots if an underlying ocean reflectance of 1.5% is added to the atmospheric reflectance (ocean reflectance will be discussed in more detail later).

5.3.1 Gauss-Seidel Calculations

As a check to the accuracy of the single scattering solution to atmospheric scattering, comparisons were made for channel 1 with the full multiple scattering code of Herman and Browning (1964) utilizing a Gauss-Seidel iteration scheme. The code used was implemented for monochromatic calculations by Ben-David and Thome (1990) for user provided optical thicknesses of aerosol, molecular scattering, water vapor, the uniformly mixed gases, and ozone. Aerosol optical properties and profiles of each of these components are also needed. It is difficult to fit these inputs to LOWTRAN where optical thicknesses are never explicitly calculated. Only an effective optical thickness for the given path and band can be inferred from LOWTRAN transmittance data, which will not necessarily behave as a monochromatic optical thickness for scattering calculations. Rather than comparing these results absolutely with those of LOWTRAN, emphasis will be on the ratio of single scattering to total scattering as a measure of LOWTRAN's underestimation of scattering. The Gauss-Seidel output also allows a calculation of the diffuse to direct ratio of transmitted solar radiation, which if significant would have to be accounted for in the atmospheric correction scheme (see Figure 5.1).

The channel 1 calculations were made at $\lambda=0.65 \mu\text{m}$. Two different choices of parameters are given in Table 5.3. The Rayleigh optical thickness was from Elterman and Toolin (Chapter 7, Valley 1965); a depth four fifths of the total was used for the 2 km path, assumed at about an 800 mb level. Aerosol size distributions and optical properties were from Jursa (Chapter 18, 1985). Case I aerosol parameters are for a pure marine environment with sea salt solution particles; case II is for a rural aerosol mixture. The navy maritime aerosol model that is used in LOWTRAN would presumably lie somewhere in between. Aerosol optical thicknesses were chosen to give comparable transmissions with LOWTRAN. Other optical thicknesses were from MLW values given by Selby et al. (1978).

A summary of the Gauss-Seidel calculations for both cases of aerosol parameters is presented in Tables 5.4a and 5.4b for the sea level and 2 km path respectively. Angles used for the satellite-sun geometry are the same as in the LOWTRAN runs (with the exception of $\theta_{\text{sat}}=5^\circ$, which is as close to a zero zenith angle as was allowed by the code). The column labeled M.S. shows the complete multiple scattering solution. The main intention of the analysis was to estimate the contribution of higher order scattering to reflection. This is given in the table as the ratio of single scattering reflection to the complete multiple scattering solution (see column S.S./M.S.). This factor multiplied by the complete solution gives the single scattering contribution only, which can be compared with the LOWTRAN values shown in the last column. LOWTRAN overestimates the single scattering by 20 to 40 percent compared with the Gauss-Seidel for the sea level to space path, and underestimates by up to 20 percent for

Table 5.3 Inputs used in Gauss-Seidel calculations, including two different cases of aerosol parameters.

Optical Depths:	Case I		Case II
H ₂ O	.0003		Same as for Case I
Uniformly mixed gases	.0002		
O ₃	.03		
Atmospheric Path (height to space)	0 km	2 km	
Rayleigh	.05	.039	
Aerosol	.115	.050	
Aerosol Parameters:			
n _{real}	1.377		1.53
n _{imaginary}	3.33x10 ⁻⁸		5.0x10 ⁻³
Mode radius (μm)	.32		.03
σ (for log normal distribution)	.40		.35

the 2 km to space path. These differences are most likely linked to the choice of aerosol optical thicknesses and should not be taken as a failure in the single scattering computation algorithm of LOWTRAN. An average of about 0.85 for the single to multiple scattering ratio is obtained for the sea level path if θ_{sat} is limited to 5° and 30° , and both cases are given equal weighting. Satellite zenith angles of 60° are subject to larger atmospheric corrections and it appears from this data, also from LOWTRAN transmission and reflection data, that it is judicious to disregard that part of the image beyond a θ_{sat} of about 45° . This is further justified in the next section when atmospheric emission is analyzed. This restriction still leaves two thirds of the image, centered evenly about the sub-satellite point, for analysis. For the 2 km path, the single to multiple scattering ratio has an average of about 0.90. In other words, considering both paths, reflectance is underestimated by 10 to 15 percent. As an approximation, this can be used to modify the LOWTRAN single scattering results. The ratio of diffuse to direct solar radiation at the sea level and 2 km level is also given in Table 5.4. It is a function of solar zenith angle only. These values might be significant for reflectance calculations and can be approximately incorporated into the atmospheric correction scheme described at the beginning of this chapter. For instance the direct transmittance $t(\mu_0)$ in equation 5.1 can be increased by the factor $(1 + \text{diffuse/direct})$ to allow for the extra solar radiation present at the object due to diffuse transmission. No attempt has been made to quantify the angular distribution of the diffuse radiation and it will be assumed that it can be considered as coming from the same direction as the direct beam for purposes of approximating atmospheric correction.

Table 5.4a Summary of Gauss-Seidel calculations for path from sea level to space for the two cases given in table 5.2; solar zenith angle is 45° . M.S. is the complete multiple scattering solution, S.S./M.S. is the ratio of single scattering reflection to the complete solution, and diffuse/direct is the ratio of diffusely transmitted solar radiation to the direct solar beam at sea level.

Geometry		Bidirectional Reflectance (%)						
		Case I			Case II			LOWTRAN (avg. of MLS and MLW)
θ_{sat}	ϕ	M.S.	S.S. /M.S.	diffuse /direct	M.S.	S.S. /M.S.	diffuse /direct	
5	-	2.33	.84	.20	4.67	.84	.114	2.29
30	180	3.43	.86		5.75	.97		3.90
	0	2.15	.78		4.20	.81		2.12
60	180	5.90	.81		10.6	.84		6.31
	0	5.0	.70		7.63	.79		5.00

Table 5.4b Summary of Gauss-Seidel calculations for path from 2 km to space for the two cases given in table 5.2; solar zenith angle is 45° . M.S. is the complete multiple scattering solution, S.S./M.S. is the ratio of single scattering reflection to the complete solution, and diffuse/direct is the ratio of diffusely transmitted solar radiation to the direct solar beam at a 2 km height.

Geometry		Bidirectional Reflectance (%)						
		Case I			Case II			LOWTRAN (avg. of MLS and MLW)
θ_{sat}	ϕ	M.S.	S.S. /M.S.	diffuse /direct	M.S.	S.S. /M.S.	diffuse /direct	
5	-	1.65	.91	.102	2.70	.90	.066	1.23
30	180	2.44	.91		3.84	.91		1.81
	0	1.46	.87		2.42	.86		1.12
60	180	4.19	.89		6.41	.90		3.06
	0	3.01	.83		4.36	.85		2.47

All ratios in the table are for a single solar zenith angle. Single to multiple scattering ratios will generally increase at smaller solar zenith angles (smaller optical paths) and diffuse transmission of radiation will decrease. However in light of other uncertainties (LOWTRAN accuracy and validity of seasonal standard atmospheres for actual images) the values in the table are considered sufficient for current purposes.

Because of computation time, only two azimuth angles (0° and 180°) were considered in the LOWTRAN runs. An implicit assumption was that these two choices represent extremes in the azimuthal variation of reflectance. The Gauss-Seidel code outputs bidirectional reflectance for azimuthal angles $0^\circ, 30^\circ, 60^\circ, \dots, 180^\circ$. This variation for $\theta_0 = 45^\circ$ is shown for several satellite viewing angles in Figure 5.5. Up through viewing angles of about 45° , the 0° and 180° azimuths do approximate the extremum.

5.3.2 Comparisons with HRPT clear sky data

It is tempting to use the quadratic approximations of the previous plots as a fit to the atmospheric reflectance dependance on satellite zenith angle. But actual clear sky data show this to be an optimistic use of the LOWTRAN results for atmospheric correction. In fact, no such zenith angle dependance was apparent in twenty clear sky areas analyzed in an HRPT image from March 20, 1989 located off the west coast of California and Mexico (see Figures 5.6a and 5.6b). Channels 1 and 2 were analyzed; no attempt was made to do the same for channel 3 because of the need to correct for emission. Calibration coefficients were those of Holben and Kaufman (1991). The data

is grouped into two azimuthal angle bins; similar solar zenith angles occur for each group. For azimuth angles in the 50° bin (open circles), reflectance actually increases as satellite zenith decreases. At this azimuth, specular reflectance off the sea surface, called sun glint (magnitude would depend on surface roughness), could exist and would peak at a satellite zenith about equal to the solar zenith. This explanation seems plausible for that data group. For the points where the satellite is looking away from the sun (dark circles) there is little suggestion of a satellite zenith dependence. The clear sky regions examined in the image were almost always surrounded by clouds and it is possible that variations in humidity or aerosol are responsible for the lack of variation with satellite zenith angle. Differences in ocean reflectance away from the sun glint direction can also cause a spread in the data. A constant value for atmospheric reflectance is used in the correction scheme (see Table 5.6).

Nevertheless, these last two plots still provide an opportunity to compare the general results of LOWTRAN atmospheric reflectance with actual data. Even when ignoring pixels contaminated by sun glint, this comparison requires some estimate of ocean surface bidirectional reflectance. One component to this reflection arrives from underneath the sea surface by light which is refracted through the surface and then scattered off suspended particulates such as plankton, sediments, and other material. This component would be present even for a perfectly smooth water surface. Reflectance spectra in the visible for various particulate combinations are summarized by Robinson (1985). For phytoplankton-dominated water, this subsurface reflectance is less than 1 % of the incident flux just below the surface. This value is quite sensitive

to phytoplankton concentrations in the visible but seems to tend to zero in the channel 2 band. Of course some of this scattered light undergoes Fresnel reflection from the sea-air interface back down into the water and the process repeats. Preisendorfer and Mobley (1986) give lambertian distributed, water-incident radiance reflections of 50 to 70% for moderate wind speeds from monte carlo simulations. So, ignoring multiple reflections between the surface and the water particles, the net subsurface reflectance seen above the water is likely to be less than 0.5 to 0.7 %.

The second and more obvious component to ocean reflectance is Fresnel reflectance from those sea wave facets directing light towards the satellite sensor. It is dependent on wind speed but unlike subsurface scattering is virtually independent of the radiation wavelength. The Preisendorfer and Mobley study also presents albedos for several point source geometries, varying from 2% to 3% as solar zenith angles go from zero to 45° and for moderate wind speeds. This albedo is heavily weighted by scattering in the sun glint direction and is not necessarily indicative of bidirectional reflectance in other directions. Rao et al. (1989) remark, based on model computations, that away from the sun glint direction, the bidirectional reflectance can be approximated as about 0.2 to 0.4 of the albedo. This would imply a net Fresnel reflectance range of 0.4% to 1.2% for both channels 1 and 2, though no consideration can be allowed for the actual satellite view in this simple approximation. Combining with subsurface reflectance gives a net reflectance in channel 1 of about 0.9% to 1.9%, most applicable to solar zenith angles less than 45° .

Now, using a nominal ocean reflectance in channel 1 of 1.5%, multiplying by

channel 1 transmittance (using 0.75) squared, and then subtracting this product from the off glint data of Figure 5.6a (approximately 3.5% to 5%) leaves an atmospheric reflectance in the range 2.7% to 4.2%. This fits well within the extent of the LOWTRAN results of Figure 5.4a for a sea level to space path. With a nominal Fresnel reflectance of .8% for channel 2, again with a transmittance of 0.75, and going through the same calculations as for channel 1 give an atmospheric reflectance of 0.8% to 2.3%. This range appears to be an excellent match with the LOWTRAN results of Figure 5.4b.

The HRPT data can also be compared with simulations of the collective ocean-atmosphere system. Monte Carlo simulations at the top of the atmosphere for such a case with ocean surface waves and subsurface scattering was done by Plass et al. (1975). Calculations included bidirectional reflectance. Using a doubling type code, Ahmad and Fraser (1982) used a similar ocean-atmosphere system as Plass et al. though without subsurface scattering. These groups reported bidirectional reflectances from 5% (Ahmad and Fraser for $\phi=180^\circ$) to 10% (Plass et al. for $\phi=150^\circ - 180^\circ$) for a solar angle interpolated to 45° and a satellite zenith angle of 30° . This range is a couple of percent larger than the HRPT data. Both groups used an aerosol optical thickness of 0.214 (supposedly corresponding to $\lambda=0.7 \mu\text{m}$) which seems high for a clean maritime atmosphere and could explain the larger reflectances. Using the model of Ahmad and Fraser with a more reasonable aerosol optical thickness of 0.1 and 0.08 for channel 1 and 2 respectively, Holben and Kaufman (1991) show reflectances of 4% for channel 1 and 2% for channel 2 using an azimuth of 180° . These values are at the mean of the

HRPT data.

Both the net reflectance and the inferred atmospheric component analyzed from clear sky regions of an HRPT image are within the limits calculated by theoretical models. This provides some confidence in the LOWTRAN atmospheric reflectance results. It also gives an indication that the calibration coefficients used for channels 1 and 2 are not in gross error.

5.4 Atmospheric Emission

Ignoring scattered photons, the radiative transfer equation becomes

$$\mu \frac{dI(\tau, \mu)}{d\tau} = -I(\tau, \mu) + (1 - \varpi_0) B(\tau, \mu, T) \quad (5.9)$$

where I is the intensity at some level τ , B is the Planck blackbody function and is dependent on the temperature T at that level τ , and the wavelength dependence of each optical term is to be understood. Integrating over the atmosphere, noting that $(1 - \varpi_0)d\tau = d\tau_a$, and allowing for a surface emission determined by temperature T_s gives

$$I_v(\mu) = B_v(T_s)t_T + \int_{t_a}^1 B_v(T(t_a))t_s dt_a \quad (5.10)$$

The subscripts on the transmittances designate total, scattering, and absorption. When this equation is implemented by LOWTRAN, the transmittances become average transmittances in a 20 cm^{-1} band. For wavelengths where $\varpi_0 = 0$, no scattering occurs (only absorption) and the scattering transmittance goes to unity and $d\tau_a = d\tau$. This is the case for the thermal channels (4 and 5) of the AVHRR. For channel 3 at about 3.75

μm scattering can be significant. Though equation 5.7 accounts for scattering through ω_0 (or τ_a), this only serves to modify the emissivity of a layer. Photons which have been scattered one or more times after being emitted are neglected. This omission will modify the emission results for channel 3. Accounting for the channel spectral response gives an integrated emission

$$\frac{\int_{\nu} I_{emit}(\nu) \phi(\nu) d\nu}{\int_{\nu} \phi(\nu) d\nu} . \quad (5.11)$$

As with Equation 5.5, the numerator is obtained as output from the filter subroutine in LOWTRAN, however the denominator must be calculated separately (with values of 277.79, 77.986, and 60.27 cm^{-1} for channel 3,4, and 5 respectively).

Calculated emission is sensitive to the temperature and moisture profiles of the model atmospheres used. The MLS and MLW profiles have a generally asymptotic behaviour whereas temperature and moisture inversions are common in maritime atmospheres. It seems prudent then to examine the sensitivity of emission to the atmospheric profile. A maritime profile was developed separately for the region within and above the boundary layer. Several studies of marine stratus in various parts of the world included sounding data for the boundary layer with emphasis usually on the energy balance of the layer. Some data is also available for clear sky boundary layer regions of the tropical Pacific and Atlantic. Above the boundary layer little is published and data from radiosonde soundings at sites on the coast of California during February

of 1991 were used for this purpose.

Combining boundary layer profiles from the literature with sounding data for the rest of the atmosphere results in the atmosphere described in Table 5.5. The specific humidity (mixing ratio) is a constant 8 g/kg from the ocean surface up to the boundary layer top where the atmosphere is allowed to gradually dry out. The temperature profile shows about a 6°K inversion. Several sources were used to assemble this profile and a review of these reports follows. Data from the California coast includes the experiments of Brost et al. (1982) performed during June of 1976. They reported mixing ratios of 5-6 g/kg throughout a boundary layer containing thin stratus to 7-8 g/kg measured for thick stratus observed in the layer. Sea surface temperatures ranged from 282 K to 287 K and cloud tops from 905 mb to 940 mb. Additional data from the same experiment was given by Albrecht et al. (1985). Moisture data collected in the same region during FIRE (First ISCCP Regional Experiment) is reported by Albrecht et al. (1988) who give mixing ratios of 7-8 g/kg from the surface up to 1200 meters obtained by aircraft measurement in the summer of 1987. FIRE tethered balloon measurement at San Nicholas Island at the same time is summarized by Duda et al. (1991) and includes cloud tops ranging from 900 mb to 950 mb. Satellite-derived cloud climatologies of the California status inferred cloud heights of 750 m near the coast to 1250 m several hundred miles to the west (Heck, et al. 1990). In other oceans, Slingo et al. (1982) present data from JASIN (the Joint Air-Sea Interaction Experiment, August of 1978) for a small region to the northwest of the British Isles. Averaged data for the mixing ratio was 5-6 g/kg, cloud top at about 850 mb, and surface temperatures of 279

Table 5.5 The first 14 levels of the maritime atmosphere used by LOWTRAN in calculating emission for AVHRR channels 3, 4 and 5. Values of height for each layer are associated with the pressure through the standard U.S. atmosphere. Explanation for the profile is given in the text.

level	h(km)	p(mb)	T(K)	q (g/Kg)
1	.004	1017	284.5	8.0
2	.143	1000	284.5	8.0
3	.300	983	284.0	8.0
4	.420	962	288.0	8.0
5	.750	922	290.8	6.0
6	.900	908	290.8	2.0
7	1.52	850	286.0	1.17
8	3.11	700	272.5	.42
9	4.00	614	267.6	.30
10	5.72	500	256.0	.11
11	7.35	400	242.9	.03
12	8.30	339	234.3	.01
13	9.33	300	228.1	.0
14	10.53	250	221.1	.0

to 285 K. A mean temperature inversion of about 5 K was found above cloud top. Data from the Beaufort Sea was compiled during the Arctic Stratus Experiment in June 1980 (see Tsay and Jayaweera 1984 and Curry 1986). Mixing ratios were lower than the previous reports, from 3-5 g/kg. Cloud tops ranged from 300 to 1100 meters, and the boundary layer topped temperature inversions appear to be as great as 10 K. In stratacumulus near Coff's Harbor Australia, Paltridge (1974) shows average mixing ratios as high as 11.5 g/kg for a particular cloud with a top at 1500 meters. Ocean temperatures for this example were about 291 K. The emphasis of this chapter is on the effects of a cloud-free atmosphere and each of these sited measurements were in the presence of at least some cloud. It could be argued that the existence of cloud indicates a process at work different than that for a cloud-free region and that mixing ratio and temperature structures could differ from the data mentioned above. Some measurements, which apparently include clear sky situations are summarized by Riehl (chapter 5, 1979) for a region between California and Hawaii and also in the whole of the tropical Atlantic (data from the Atlantic Trade Expedition, 1969). Soundings for the Pacific region were taken by the U.S. Navy during summer and fall of 1945. Mixing ratios at the sea surface were as much as 10 g/kg 800 miles off the coast of California and increasing to 14 g/kg near Hawaii. Along the same section the inversion heights increased from about 860 mb ($\approx$ 1.3 km) to 800 mb ($\approx$ 2 km). Riehl's review shows the tropical Atlantic inversion to be at typical heights of 1.0 km to 2.0 km. Temperature inversions ranged from 1 K to 8 K depending on the region of the Atlantic.

For the atmosphere above the boundary layer, radiosonde soundings along the California coast were collected during the latter part of February 1991. The maritime quality of the coastal atmosphere was judged by the generally offshore flow at upper levels in the sounding, and GOES satellite images showing the formation of extensive marine stratus offshore as an indication of widespread existence of an unbroken inversion. A sounding from Vandenburg Airforce Base on February, 25, 1991 was chosen for the above boundary layer profile. On the following day a profile from Oakland, CA showed upper level moisture at the 400mb level with approximately 60% RH indicating that there is no one atmosphere valid for a clear sky region. Placing this upper level moisture into the Vandenburg profile showed little change in LOWTRAN calculated emission which is no surprise because of the associated low temperatures in this instance. A few soundings taken along the Atlantic and Gulf coast showed inversions at typically greater heights with less drastic gradients.

While this assembled atmosphere cannot be considered climatologically or seasonally (and of course not geographically) correct, it does offer the essence of the maritime profile and might be considered typical for the eastern Pacific near California. Other atmospheric species, including aerosols, kept the same profiles as discussed earlier in this chapter.

A user-defined atmosphere featuring this inversion structure was supplied to LOWTRAN, with nadir-view results being displayed in Figures 5.7a-c. Model MLS and MLW results are also shown versus viewing angle. For channels 4 and 5, the maritime atmosphere gives emissions close to that of the MLS-MLW mean value for the

sea level to space path. Emission from the top of the boundary layer to space is nearer the MLS emission value. Recall that for the model atmospheres, the boundary layer top is somewhat arbitrarily defined (based on the aerosol profile) at a height of 2 km (nominally 800 mb) but is about 920 mb for the maritime atmosphere. Nevertheless, for the thermal channels, each class of atmosphere gives very comparable emissions for both paths and so the results of the MLS-MLW atmospheres will be considered representative of the maritime conditions. However, the channel 3 data of Figure 5.7c implies that the maritime emission is more than twice that of the MLS-MLW average.

5.5 Summary of AVHRR Atmospheric Corrections

The data presented in sections 5.3 - 5.5 shows large variability for atmospheric transmission, reflectance and emission for the two model atmospheres (MLS and MLW) and the two paths, sea level and top of the boundary layer to space. All effects depend on the satellite viewing angle, and for reflectance, solar geometry as well. Given the uncertainties in having quantified these behaviors and the validity in the model atmospheres, it was concluded that a single value of correction for each path and all viewing angles would be as far as the above analysis should be carried. Satellite zenith angles will be limited to 45° if possible to avoid the large variations associated with the increasing optical depths. With this restriction, the rms satellite viewing angle turns out to be about 25° ($\langle \theta_{\text{sat}}^2 \rangle = \int \theta_{\text{sat}}^2(x) dx / \int dx$ where x represents the number of pixels away from nadir and the limits of the integration extend to $\pm x_{\text{max}}$, x_{max} being the pixel location where the viewing angle is 45°). The correction value was chosen

Table 5.6 Atmospheric quantities obtained from LOWTRAN. The average of MLS and MLW results for a satellite zenith angle of 25° . For reflectance (2nd data column) the average includes azimuths of 0° and 180° at a solar zenith angle of 45° , and limits for channel 1 and 2 include nadir view calculations at a solar zenith angle of 0° and 180° . The paranthetic values include the multiple scattering approximations.

AVHRR Channel/Path		Transmission	Reflectance (Scattering)	Emission ($\text{mW/m}^2\text{-Sr-cm}^{-1}$)
C H 1	0km	.74(.78) +.06,-.07	(% Albedo)	---
			2.76(3.25) +3.15,-.96	
	2km	.85(.88) +.02,-.05	1.91(2.12) +1.48,-.51	---
C H 2	0km	.73 +.03,-.06	1.47 +1.88,-.62	---
	2km	.87 +.02,-.03	.83 +.62,-.23	---
C H 3	0km	.75 +.03,-.07	($\text{mW/m}^2\text{-Sr-cm}^{-1}$)	.047 +.039,-.025
			.010 +.016,-.007	
	2km	.88 +.02,-.04	.005 +.003,-.004	.014 +.011,-.008
C H 4	0km	.79 +.11,-.16	---	17.50 +16.54,-11.36
	2km	.93 +.03,.05	---	4.56 +4.44,-2.56
C H 5	0km	.70 +.16,-.22	---	27.88 +26.15,-17.95
	2km	.89 +.05,-.09	---	7.61 +6.38,-4.42

as the average of the MLS and MLW values at 25° ; the 25° value was found from the quadratic approximations shown in previous figures. For reflectance an additional average of the 0° and 180° azimuth was done. A summary is given in Table 5.6.

5.7 Uncertainty Analysis

Equations 5.1-5.3 describe the net signal seen by the satellite. With knowledge of the atmospheric effects the object's reflection or emission can be inferred. Of course, values for the atmospheric transmission, scattering, and emission terms can never be known precisely. As seen throughout this chapter, without apriori information concerning the state of the atmosphere, the use of nominal values (Table 5.6) is about as well as can be done. For such a case, it is difficult to determine an uncertainty in the retrieved quantity since the conclusion would only be bounded by the number of model atmospheres used in the analysis and their deviations from the chosen nominal value. A preferable approach, independent of model atmospheres, is to consider the maximum uncertainty allowable in the knowledge of the atmospheric terms, with respect to some assumed nominal values, in order for the uncertainty in the retrieved quantity to be less than some prescribed value. For example, in order to be able to determine, say, cloud reflectance to within 5%; how close to the chosen nominal values would the transmission and reflectance of the overlying atmosphere have to be known? The answer provides a quantifiable means for understanding signal sensitivity to the atmosphere and is quite independent of any actual atmosphere though a nominal value must sometimes be chosen. The results are general and can be applied to any satellite

sensor, of any wavelength, providing that it can be lumped into one of the categories associated with equations 5.1-5.3 (approximations for atmospheric scattering only, emission only, and both scattering and emission).

Consider a channel where atmospheric emission can be ignored and the appropriate quantity is reflectance (AVHRR channel 1 and 2) and where the correction scheme uses nominal values that were somehow determined to be representative. The unknown deviation in atmospheric transmission and reflectance from the nominal values causes an uncertainty in the retrieved reflectance of the scene under study. Rearranging equation 5.1 for scene reflectance R_s gives

$$R_s = \frac{R_{sat} - R_{atm,o}}{t_o^2} \quad (5.12)$$

where the geometric dependencies have been dropped and are to be understood. $R_{atm,o}$ and t_o^2 represent the nominal correction. This is the equation used for channel 1 and 2 atmospheric correction in the next chapter. But the satellite signal is affected by the real atmosphere whose properties are unknown, not the assumed atmosphere. This uncertainty which then influences the retrieved reflectance can be shown explicitly in the form of equation 5.1 by:

$$R_{sat} \Big|_{-o+}^+ = R_s t(\mu_o) \Big|_{-o+}^+ t(\mu) \Big|_{-o+}^+ + R_{atm} \Big|_{-o+}^+ . \quad (5.13)$$

The symbols -, o, and + are to represent minimum, nominal, and maximum limits to the unknown respectively. Let the nominal values be the same as those used in equation

5.12. Substituting this expression for R_{sat} into equation 5.12 gives

$$R_s^* = \frac{R_s t(\mu_o) \left(\frac{+}{-} \right) t(\mu) \left(\frac{+}{-} \right) + R_{atm} \left(\frac{+}{-} \right) - R_{atm}}{t_o^2} \quad (5.14)$$

R_s^* is the retrieved or inferred scene reflectance, not necessarily the actual one unless the chosen nominal atmospheric transmission and reflectance turn out to be valid for the current atmosphere. Defining $\Delta R_s = R_s^* - R_s$, an error for the retrieved reflectance can be expressed as

$$\frac{\Delta R_s}{R_s} = \left[\frac{t}{t_o} \right]^2 - 1 + \frac{R_{atm} \left(\frac{+}{-} \right) - R_{atm}}{R_s t_o^2} \quad (5.15)$$

Note that R_s in this equation is the actual scene reflectance, not the retrieved value. The effect of each of the atmospheric terms can be analyzed separately.

Consider first the uncertainty in transmission. If the atmospheric scattering back towards the satellite is considered to be known exactly, and the transmission error is defined as the difference between the actual transmission and the nominal transmission, then Equation 5.15 can be written as

$$\frac{\Delta R_s}{R_s} = \left[1 + \frac{\Delta t}{t_o} \right]^2 - 1 \quad (5.16)$$

No assumed nominal transmission is involved in this expression. Referring to Figure 5.8, if the error in retrieved scene reflectance is to be less than 5% than the uncertainty

in transmission must be less about 2.5%! Calculations (direct or remote) of optical thickness are more likely than actual transmission. So considering the monochromatic relation

$$\frac{\Delta \tau}{\tau_o} = \frac{\ln \left(1 + \frac{\Delta t}{t_o} \right)}{\ln(t_o)} \quad (5.17)$$

to be valid, the last graph can be couched in terms of τ as shown in Figure 5.9. Now, better than 5% uncertainty in reflectance implies optical depths should be known to within 8% to 15% (depending on actual optical thickness; $t_o = .85$ corresponds to the top of the boundary layer to space path for both channel 1 and 2 of the AVHRR, $t_o = .75$ to the sea level path).

Figure 5.10 shows a similar analysis as above but with respect to atmospheric reflectance. When atmospheric transmission is considered to be known exactly, the uncertainty is

$$\frac{\Delta R_s}{R_s} \Bigg|_o^+ = \frac{\Delta R_{atm}}{R_{atm}} \Bigg|_o^+ \frac{R_{atm}}{R_s t_o^2} \quad (5.18)$$

The results are not as simple to represent as before because of the dependence on R_{atm} , t_o , and R_s . Two different cases, again for the sea level and boundary layer paths, are shown. For the case of 5% maximum error in reflectance, $\Delta R_{atm}/R_{atm}$ must be known within a factor of 1-2 (depending on the path) of R_s ; for a cloud with 20% reflectance

the uncertainty in R_{atm} must be less than 20% to 40%. These graphs can also be used with the ordinate as the independent variable. For example, given the percent transmission uncertainty, the resulting uncertainty in retrieved reflectance can be read from Figure 5.9.

Some combination of uncertainty will exist for both transmission and scattering. In fact, there will most likely be a correlation between the two. For example, if transmission is known exactly there should be a relatively small uncertainty in scattering. Equation 5.17 and 5.18 sum to give the total reflectance uncertainty so any combination of transmittance and scattering uncertainty can be used with the graphs to give the total.

Next consider a channel where atmospheric scattering and emission are important (AVHRR channel 3) and the pertinent quantity is intensity. Proceeding in a manner similar to the analysis for reflectance uncertainty, Equation 5.3 is solved for the solar intensity scattered from the scene, $I_{s,scat}$, using nominal correction values:

$$I_{s,scat} = \frac{I_{sat} - t_o I_{s,emk} - I_{atm,emk,o} - I_{atm,scat,o}}{t_o^2} \quad (5.19)$$

Using Equation 5.3 directly for the uncertainty in I_{sat} and substituting this into the last equation gives the retrieved scene scattered intensity $I_{s,scat}^*$. For transmission error only (atmospheric emission and scattering, and the temperature of the emitting body known exactly),

$$\left(\frac{\Delta R_s}{R_s} \right)_o^+ = \frac{\left(\frac{\Delta t}{t_o} \right)_o^+ + I_{s,emit} \left(\frac{\Delta t}{t_o} \right)_o^+ - t_o I_{s,emit}}{t_o^2} \quad (5.20)$$

Since $R_s = \pi I_{s,scat} / K \mu_o$, where K is a constant, Equation 5.20 can be placed in terms of fractional reflectance uncertainty as

$$\left(\frac{\Delta R_s}{R_s} \right)_o^+ = \left[1 + \frac{\left(\frac{\Delta t}{t_o} \right)_o^+}{t_o} \right] - 1 + I_{s,emit} \left[\frac{\left(\frac{\Delta t}{t_o} \right)_o^+}{t_o} \right] \frac{1}{t_o R_s K \mu_o / \pi} \quad (5.21)$$

Note that if there is no emission, the expression is identical to Equation 5.16. The relation between fractional reflectance and fractional transmission uncertainty includes a dependance on R_s , t_o , $I_{s,emit}$ which in turn is a function of temperature, and solar zenith angle μ_o . Figures 5.11a,b give a plot of Equation 5.21 for $\mu_o = .707$ and a temperature of 280 K and 290 K respectively. Four different combinations of R_s and t_o are shown which provide useful bounds on the error. The dependence of t_o is minor. Mapping transmission into optical thickness gives Figures 5.12a,b for the same conditions. Now the dependence on t_o (or τ_o) is as significant as for R_s .

For atmospheric emission error only (atmospheric transmission and scattering, and temperature of the emitting body known exactly), an analogous derivation has as the end result

$$\frac{\left(\frac{\Delta I_{atm,emit}}{I_{atm,emit,o}} \right) \frac{o}{-}}{=} = \frac{\left(\frac{\Delta R_s}{R_s} \right) \frac{o}{-}}{=} \frac{K \mu_o R_s t_o^2}{I_{atm,emit,o}} . \quad (5.22)$$

Likewise, for atmospheric scattering error only,

$$\frac{\left(\frac{\Delta I_{atm,scat}}{I_{atm,scat,o}} \right) \frac{o}{-}}{=} = \frac{\left(\frac{\Delta R_s}{R_s} \right) \frac{o}{-}}{=} \frac{K \mu_o R_s t_o^2}{I_{atm,scat,o}} . \quad (5.23)$$

These last two equations are plotted together in Figure 5.13 for $\mu_o = .707$, and two combinations of t_o and $I_{atm,scat}$ or $I_{atm,emit}$ corresponding to the sea level and 2 km path (see Table 5.6). R_s is incorporated into the ordinate. Emission uncertainty causes about four times the retrieved reflectance uncertainty as does scattering uncertainty.

Finally consider a channel where only atmospheric emission is important (AVHRR channels 4 and 5) and the quantity of interest is temperature. Referring to Equation 5.2, it can be shown that

$$\frac{\left(\frac{\Delta I_s}{I_s} \right) \frac{o}{-}}{=} = \frac{\left(\frac{\Delta t}{t_o} \right) \frac{o}{-}}{=} + \frac{\left(\frac{\Delta I_{atm}}{I_{atm,o}} \right) \frac{o}{-}}{=} \frac{I_{atm,o}}{I_s t_o} . \quad (5.24)$$

The trick is to express the uncertainty in intensity in terms of temperature. First, assume the atmospheric emission is known exactly. Using the Planck function, and a bit of algebra, the link between transmission uncertainty and the resulting uncertainty in retrieved temperature is

$$\frac{\Delta t}{t_o} = \frac{\exp(C_2\nu/T_s) - 1}{\exp\left(\frac{C_2\nu/T_{s_o}}{1 + \frac{\Delta T_s}{T_{s_o}}}\right) - 1} - 1 \quad (5.25)$$

T_s is the temperature of the scene (object) being viewed, $c_2 = 1.438833$ cm-K, and ν is the wavenumber. A plot of this equation for absolute scene temperature error and two scene temperatures is shown in Figure 5.14. If it is the transmission that is known exactly, the equivalent expression

$$\frac{\Delta I_{atm}}{I_{atm_o}} = \left(\frac{\exp(C_2\nu/T_s) - 1}{\exp\left(\frac{C_2\nu/T_{s_o}}{1 + \frac{\Delta T_s}{T_{s_o}}}\right) - 1} - 1 \right) \frac{I_{s_o}}{I_{atm_o}} \quad (5.26)$$

results and is displayed in Figure 5.15. It appears that t_o is of primary importance here.

As was argued at the beginning of this section, a look at the spread of retrieved quantities for different specific atmospheres is only limited by the variety of atmospheres that can be developed. Nevertheless, such information can provide a good summary of the LOWTRAN results and helps develop an appreciation for the need of accurate atmospheric corrections in cloud microphysics remote sensing. Figures 5.16a-e summarize the application of Equations 5.15, 5.21, 5.25, and 5.26 for channels 1-4 of the AVHRR. In each equation the absolute uncertainty in the scene reflectance,

intensity, or temperature is evaluated for the range of atmospheric corrections of Table 5.6. The nominal values give the zero error line on the graphs. The spread away from this zero is then plotted for the two atmospheric paths repeatedly discussed in this chapter.

Figure 5.1 The effect of the atmosphere on the received signal at the satellite. Heavy lines indicate the dominant effects analyzed in this chapter.

(5.2-a)

(5.2-b)

(5.2-c)

(5.2-d)

(e)

Figure 5.2 Spectral transmission for major attenuators superimposed on AVHRR channel response for (a) channel 1, (b) channel 2, (c) channel 3, (d) channel 4, and (e) channel 5; using LOWTRAN with the Navy Maritime Aerosol Model and the mid-latitude summer model atmosphere.

(5.3-a)

(5.3-b)

(5.3-c)

(5.3-d)

(e)

Figure 5.3 AVHRR atmospheric transmission in (a) channel 1, (b) channel 2, (c) channel 3, (d) channel 4, and (e) channel 5 for a path from space to the surface (0 km) or to the top of the boundary layer (2 km). Calculations made using LOWTRAN with the Navy Maritime Aerosol Model and the mid-latitude summer model atmosphere.

(5.4-a)

(5.4-b)

(c)

Figure 5.4 Intensity of atmospheric scattering in AVHRR (a) channel 1, (b) channel 2, and (c) channel 3 for a path from space to the surface (0 km) or to the top of the boundary layer (2 km), with solar zenith angle of 45°. Calculations made using LOWTRAN and the Navy Maritime Aerosol Model.

Figure 5.5 Azimuthal variation in bidirectional reflectance for a solar zenith angle of 45° and several satellite zenith angles. Calculated with Gauss-Seidel code.

(a)

(b)

Figure 5.6 Measured bidirectional reflectance of a clear sky ocean scene for AVHRR (a) channel 1 and (b) channel 2. Data from HRPT 20 March 1989, latitude = 20-35° N, longitude = 115-145° W. The range of solar zenith angles and azimuth angles is shown in the legend.

(c)

Figure 5.7 Intensity of atmospheric emission in AVHRR (a) channel 3, (b) channel 4, and (c) channel 5 for a path from space to the surface (0 km) or to the top of the boundary layer (2 km), with a solar zenith angle of 45° . Calculations made using LOWTRAN and the Navy Maritime Aerosol Model.

Figure 5.8 Allowable uncertainty in atmospheric transmission for AVHRR channel 1 for a specified maximum error in retrieved reflectance.

Figure 5.9 Allowable uncertainty in atmospheric optical thickness for AVHRR channel 1 for a specified maximum error in retrieved reflectance. Transmission is specified in the legend corresponds to LOWTRAN calculations for a path from space to the surface ($t_0=0.75$) or to the top of the boundary layer ($t_0=0.85$).

Figure 5.10 Allowable uncertainty in atmospheric scattering for AVHRR channel 1 for a specified maximum error in retrieved reflectance. The ordinate must be multiplied by the actual scene reflectance in % (i.e., 0-100). Transmission is specified in the legend corresponding approximately to LOWTRAN calculations for a path from space to the surface ($t_o = 0.75$) or to the top of the boundary layer ($t_o = 0.85$).

(a)

(b)

Figure 5.11 Allowable uncertainty in transmission for AVHRR channel 3 for a specified maximum error in retrieved reflectance, for a scene temperature of (a) 280 K and (b) 290 K. Atmospheric transmission and actual scene reflectance are specified in the legend. The solar zenith angle is 45°.

Figure 5.12 Same as Figure 5.11 but for allowable uncertainty in atmospheric optical thickness for AVHRR channel 3, for a scene temperature of (a) 280 K and (b) 290 K.

Figure 5.13 Maximum allowable uncertainty in atmospheric scattering and emission for a specified maximum error in retrieved reflectance for AVHRR channel 3. The ordinates must be multiplied by the actual scene reflectance in % (i.e., 0-100). Atmospheric transmission and emission (I_e) or scattering (I_s) are specified in the legend.

Figure 5.14 Maximum allowable uncertainty in transmission for a specified maximum error in retrieved temperature for AVHRR channel 4 at two scene temperatures.

Figure 5.15 Maximum allowable uncertainty in atmospheric emission for a specified maximum error in retrieved temperature for AVHRR channel 4 at two scene temperatures and two transmission values.

(5.16-a)

(5.16-b)

(5.16-c)

(5.16-d)

(e)

Figure 5.16 Range in error for retrieved quantities based on the following: (a) the spread in atmospheric transmission and scattering for AVHRR channel 1 and (b) channel 2; (c) the spread in atmospheric transmission, scattering, and emission for AVHRR channel 3 at scene temperatures of 290 K and (d) 280 K; (e) the spread in atmospheric transmission and emission for AVHRR channel 4. The spread in the atmospheric effects were calculated using LOWTRAN.

CHAPTER 6

THE REMOTE SENSING ALGORITHM AND SOME RESULTS FOR MARINE STRATUS CLOUDS

Analysis of several AVHRR satellite images is presented in this chapter. A library of reflectances and effective emittances for a variety of solar and satellite viewing angles, and cloud optical thickness and mean drop radius has been set up as described in section 4.3. Details of the remote sensing algorithm using the library are discussed first. A goal of this study is to define susceptibility, demonstrate the feasibility of its remote sensing, and obtain some feel for the expected values associated with relatively clean maritime clouds. Calculations of susceptibility require inference of optical thickness and microphysics (i.e., r or N), each of which is, of course, a fundamental quantity of use beyond the narrow confines of this susceptibility study. So an equally valuable, more fundamental goal, is the retrieval of these cloud parameters. Throughout this chapter, results will be presented for both susceptibility and the cloud parameters from which it is calculated. One of the obvious difficulties is the lack of actual cloud microphysical data for comparison with the inferred values. Several investigators have made in situ cloud drop radius measurements in marine stratus at about the same time as an overpass of a NOAA polar orbiter. Two such AVHRR images are examined and the results compared with the experimental data. The stratus surrounding ship tracks provides a good indication of relatively susceptible clouds; images containing these tracks are analyzed next. Ship tracks are an interesting phenomena in their own right, and microphysical data for these tracks can help clarify

their development. For instance, is it the microphysics or the meteorology of the stratus in which the tracks form that is of major importance in track development? Stratus in the absence of tracks is also analyzed. At the other end of the susceptibility scale is fog in California's central valley which can be expected to be quite polluted from both anthropogenic and natural sources. Also, being the more homogeneous and long-lived of low level continental clouds, the central valley fog provides a credible check for the remote sensing of small susceptibilities. The chapter concludes with a summary of these results.

Some further comments on susceptibility are useful before results are presented. Cloud climatologies, such as those by Heck et al. (1990) give cloud albedo only. This is sufficient for understanding the current shortwave energy balance, but *is not* sufficient for determining the change in albedo that would occur as a result of increasing cloud drop concentrations; cloud microphysics must also be known. Climate modeling of this effect typically assumes that drop concentrations in clean maritime clouds increase by some factor (e.g., Ghan et al. 1990 use a factor of 4 in their study). While changes in other climate modifiers, such as CO_2 , may be usefully expressed in this way, the same is not true of cloud drop concentrations. For example, pollution is more likely to increase CCN concentrations by some absolute number, say 10 cm^{-3} , over some geographical region. The effect this would have on a cloud that would otherwise have drop concentrations of 10 cm^{-3} can be very different than for a cloud having the same albedo but with $N=100 \text{ cm}^{-3}$ (and a smaller geometrical thickness). Each will respond differently to the increase in drop concentration (see Figure 2.5). It is not at all clear

that every maritime cloud should have about the same drop concentration, which is the underlying assumption of using a factor increase in N to model albedo modification. The important point is that, in addition to albedo, cloud drop concentration must be known to adequately infer albedo modification by pollution, and simple assumptions about the microphysics can be misleading.

Once A , N , and ΔN are known, a calculation can be made for the change in albedo. Since ΔN is variable, it is useful to define some figure of merit for the sensitivity of albedo to changes in drop concentration. Cloud susceptibility, as defined in this study, is interpreted as giving the change in albedo resulting from the addition of one cloud drop per cm^3 ($\Delta N=1$) under conditions of constant liquid water content (this is approximate to the extent that the addition of one drop represents a discrete change in N and susceptibility is defined for a physically unrealizable differential change). This is not, by any means, the only figure of merit, or sensitivity parameter, that could be used. For instance, the change in albedo resulting from a doubling in drop concentrations (or some other factor χ , see Equation 2.12), or the change from adding a specified number of drops where $\Delta N \neq 1$. The first alternative suffers from the reality, as previously mentioned, that polluted clouds will have a harder time in experiencing the same factor increase in N as will a cleaner cloud. The choice for some other ΔN in the second alternative is not obvious. Regardless, any definition used as a figure of merit for the change in albedo will have a two dimensional functional dependence that will include albedo (or optical thickness) and the microphysics (e.g., A and N , A and r , τ and r , A and χ); the exact definition is largely irrelevant.

However, the retrieved values of susceptibility presented in this chapter can also be used to infer changes in albedo for some $\Delta N > 1$. Starting with retrieved initial cloud parameters, follow the trajectory of dA/dN about the two dimensional surface of A and N (assume the derivative has been expressed as a function of both A and N) for incremental changes in N . This requires a calculation for the new value of dA/dN at each step in N . So, in addition to susceptibility, A and N must also be known. All three quantities are known from the remote sensing retrieval although an assumption of liquid water content is needed for calculating susceptibility and for converting radius to drop concentration. A caveat is in order. A cloud with small susceptibility might end of having a larger change in albedo than a cloud with greater susceptibility if ΔN is substantial enough. The approximation of Equation 2.12 is easier and may be just as useful; in this case A and $\Delta N/N$ must be known, and a value for susceptibility is not required. A feel for the magnitudes of susceptibility can be had with one final approximation, made by replacing χ with $1 + \Delta N/N$ in Equation 2.12 and noting that $\Delta N=1$ gives the susceptibility. Then the change in albedo for $\Delta N \neq 1$ is

$$\Delta A \approx Susc * \frac{1 - \left(\frac{1}{1 + \frac{\Delta N}{N}} \right)^{\frac{1}{3}}}{1 - \left(\frac{1}{1 + \frac{1}{N}} \right)^{\frac{1}{3}}} .$$

Since radius is retrieved, the substitution $N=3W/(4\pi\rho_l r^3)$ can be made. As an example, if $N=100 \text{ cm}^{-3}$ ($r \approx 9 \text{ } \mu\text{m}$ for $W=0.3 \text{ g/m}^3$) and $\Delta N=10$, then ΔA is susceptibility

multiplied by the factor 9.4, and if susceptibility is 10^{-3} , the change in albedo is almost 0.01. As will be seen, it is useful to express susceptibility in units of 10^{-3} cm^3 (the dimension of cm^3 will be understood in the results that follow).

6.1 The Remote Sensing Algorithm

The creation of the library data constitutes part of the forward problem, i.e., cloud drop radius, optical thickness, and temperature as well as the surface albedo and temperature are specified. This is shown schematically in Figure 6.1. Variables inside the boxes represent specified quantities; all others are calculated. The intermediate quantities in the dotted box are independent of temperature and stored in the library for the geometries mentioned in section 4.3. With cloud and surface temperature given, the library data can be used to calculate the quantities available from AVHRR data: reflectance in channels 1 and 2, total channel 3 intensity including both the reflected and emitted component, and brightness temperature for channels 4 and 5. Channels 2 and 5 are not explicitly indicated in the figure.

In this study, the inverse problem is solved by finding the best match between the measured satellite quantity for the pixel being analyzed in the AVHRR image and an entry in the library. The radius and optical thickness corresponding to this entry becomes the solution. Throughout this chapter, optical thickness is understood to mean $(2/Q_{\text{ext}})\tau$ and radius is the mean value for a normal drop size distribution with a dispersion of 0.2. The effective radius (section 4.1.1) for this dispersion is the mean radius multiplied by 1.08. Susceptibility is calculated for a liquid water content of 0.3

g/m^3 unless otherwise stated (if actual in situ liquid water content is known then the susceptibility value should be multiplied by the ratio of 0.3 to the measured value). The procedure is as follows:

- (1) For a given solar/satellite geometry, a combination of radius and optical thickness (say, the i^{th} combination) is chosen corresponding to the library quantities R_{1i} , R_{3i} , ϵ_{c3i} , ϵ_{s3i} , ϵ_{c4i} , and ϵ_{s4i} .
- (2) The quantities R_1 , I_3 , T_4 for the pixel being analyzed are extracted from the satellite data. The surface temperature, T_s , for ocean scenes is determined using a split-window technique with channels 4 and 5 (McClain et al. 1985).
- (3) The cloud temperature is determined from channel 4 emission which, using the atmospheric corrections of chapter 5, is

$$I_{A_{T_{c1}}} = \frac{I_4 - \epsilon_{s4} I_{A_{T_s}} t_4 - I_{4_{\text{atm,emit}}}}{\epsilon_{c4} t_4} \quad (6.1)$$

Temperature is found by applying the Planck function to this intensity using the AVHRR central wavenumbers described in section 3.2.3.

- (4) Now that both cloud and surface temperatures are known, the emission in channel 3 at cloud top (not at the top of the atmosphere) is given by

$$I_{3_{\text{emit}_1}} = \epsilon_{c3i} I_{3_{T_{c1}}} + \epsilon_{s3i} I_{3_{T_s}} \quad (6.2)$$

- (5) The reflected solar component of channel 3 intensity at cloud top is then

$$I_{3_{refl}} = \frac{I_3 - t_3 I_{3_{atm}} - I_{3_{atm,refl}} - I_{3_{atm,refl}}}{t_3^2} . \quad (6.3)$$

(6) The reflectance for channel 3 at cloud top is determined from the solar flux in the channel (see equation 4.31, $\langle F_0 \rangle / \pi \approx 5.3028$) as

$$R_{3_i}' = \frac{\pi I_{3_{refl}}}{\mu_o \langle F_0 \rangle} . \quad (6.4)$$

(7) Finally the reflectance for channel 1 (or channel 2) at cloud top is

$$R_{1_i}' = \frac{R_{1_i} - R_{atm}}{t_1^2} . \quad (6.5)$$

This procedure is shown schematically in Figure 6.2. Lower case t represents atmospheric transmission in the specified channel. An error function can be defined for each i^{th} combination of radius and optical thickness as given in the figure. The combination with the smallest error becomes the solution. Channel 2 can be included in this error function as well. The weights, indicated by W_1 and W_3 , are adjusted to provide the tightest error contours about the desired solution. Typically this occurs when channel 3 is given a weight with a factor of three greater than that of channels 1 and 2. For the library data, resolution in μ_o and μ is limited to 5 or 8 bins with size

0.1, (e.g., μ is confined to 0.5-0.6, 0.6-0.7,..., 0.9-1.0). The library consists of radiative properties averaged across these bins. Reflectance and emission data at discrete zenith angles for the satellite data is determined from a linear interpolation made between appropriate bins in the libraries. For the interpolation, the library entry is considered valid at the midpoint of the bin.

6.1.1 Sensitivity to Algorithm Parameters

A NOAA-11 HRPT image of stratus containing ship tracks off the coast of southern California taken March 20, 1989 was used to test the sensitivities of the solution to surface temperatures, atmospheric corrections, and weightings (see Figure 6.9). Three regions of stratus (each consisting of a 3x3 pixel area) with small to large optical thicknesses were examined. One was located just to the south of a ship track in thin stratus (optical thickness of about 2 or 3) at location 30.36°N , 125.25°W . The second was in stratus with an optical thickness of about 10 located to the west of the tracks at a location of 30.43°N , 127.71°W and the third farther to the west at 31.09°N , 127.94°W with an optical thickness of about 35. In all cases the solar and satellite zenith angles ranged from 35° to 40° and ϕ was about 140° . The linear IFOV for such satellite viewing angles is about 1.7 km.

The thin stratus case presents several challenges. As indicated in Figure 4.12, the effect of surface temperature on channel 4 brightness temperature is important for an optical thickness below about 5 (excluding the $1\text{ }\mu\text{m}$ radius curve). The uncertainty in cloud temperature produces an uncertainty in channel 3 emission which in turn affects

the calculation of channel 3 reflectance. This emission error is exacerbated by thin clouds where the channel 3 reflectance is small for all radii. Inference of optical thickness is not changed by the temperature error. And the uncertainty in the inferred optical thickness is small since dR/dr is greatest in this region. However in this thin cloud, the influence of surface reflectance will make the inversion problem more difficult in both channel 1 and 3. In the limit to a clear sky situation, if surface and atmospheric effects are properly known, optical thickness should go to zero (while radius is undefined). Sensitivity of retrieved cloud parameters is summarized in Table 6.1a versus various combinations of atmospheric correction, channel 1 calibration, surface albedo, and surface temperature. Three atmospheric situations were considered (see Table 5.6): top of the boundary layer to space, surface to space, and no correction. The 1988 channel 1 calibration of Nianzeng et al. (1991) or the 1989 calibration of Holben and Kaufman (1991) are used (see Table 3.3). Surface reflectance is considered Lambertian and was assumed to be 0.06 or 0.03 in channel 1 and 0.01 in channel 3 (see section 5.3.2) while channel 2 reflectance data is not used. Split-window temperatures of apparently clear sky areas showed ocean temperatures from 289.9 K to 292.8 K, the former value being in a region closest to the above locations. In the sensitivity study, a surface temperature of 290 K was considered nominal. A spread from 287 K to 293 K was used in the table. The table shows the retrieved values for radius and optical thickness to be surprisingly robust with respect to all variables mentioned. A large radius is always indicated. Note that a change in radius from 20 to 17.5 or 25 μm represents one increment of radius in the library. The resulting

susceptibility varies by only a factor of two (calculated with a liquid water content of 0.3 g/m^3). Recall that a susceptibility of 10×10^{-3} implies that the addition of one drop per cm^3 , under conditions of constant liquid water content, gives an increase in albedo of 0.01. Inclusion of channel 2 data (with $\alpha_{\text{sfc}}=0.06$ and the Nianzeng et al. 1988 calibration) increased the radius from 20 to $23 \mu\text{m}$ for $T_s=290 \text{ K}$ but had no effect on the optical thickness.

The cloud in the second and third case to be examined are optically thick enough so that surface effects are insignificant and the inference of radius should be independent of optical thickness. Now, $dR/d\tau$ for channel 1 decreases so the uncertainty in optical thickness can become large and is very sensitive to channel 1 atmospheric correction and calibration values. Results are shown in Table 6.1b and 6.1c. Radius is normally within one library radius increment of the nominal value. Note in Table 6.1c, that retrieved optical thickness varies from 16 to 83. Susceptibility changes almost an order of magnitude over these conditions. Inclusion of channel 2 data in these last two scenes had no effect on the retrieved radius or the optical thickness.

In summary, the thin and moderately thick stratus show little sensitivity of susceptibility to the different conditions. The thick cloud is much more variable in optical thickness and susceptibility. For each cloud, radius never varies from the nominal case by more than one library increment. The conditions used throughout the rest of this chapter are the boundary layer to space correction, $\alpha_{\text{sfc}}=0.06$ in channel 1, and the Nianzeng et al. channel calibrations. Channel 2 is not included unless otherwise stated and channel 5 is used for the split-window temperature calculation only.

Table 6.1a The effect on retrieved parameters from varying surface temperature (T_s), surface albedo (α_s), atmospheric correction (see table 5.6), and channel 1 calibration values. Data from HRPT NOAA-11, 20 March 1989, at location 30.36°N, 125.25°W.

T_s (K)	Atmospheric Correction	α_{sfc}	CH1 Calibration	r (μm)	τ	susceptibility $\times 10^3$ (cm^3)
287	B.L.→Space	.06	Nianzeng et al. -1988	20	1.6	4.39
290				23	1.7	6.89
293				26	1.7	9.19
290	Sfc.→Space	.06		16.7	1.8	2.65
		.03		18.3	2.6	4.41
	No Correction	.06		30	1.6	13.20
		.03		31	2.0	16.19
	B.L.→Space	.06	Holben & Kaufman - 1989	24	1.8	7.45

Table 6.1b The effect on retrieved parameters from varying atmospheric corrections (see table 5.6) and channel 1 calibration values. Data from HRPT NOAA-11, 20 March 1989, at location 30.43°N, 127.71°W.

T_s (K)	Atmospheric Correction	α_{sfc}	CH1 Calibration	$r(\mu m)$	τ	susceptibility $\times 10^3$ (cm ³)
290	B.L.→Space	.06	Nianzeng et al. -1988	8	7.4	0.583
	Sfc.→Space			6	10.4	0.248
	No Correction			10	6.1	1.092
	B.L.→Space		Holben & Kaufman - 1989	8	8.4	0.588

Table 6.1c The effect on retrieved parameters from varying atmospheric corrections (see table 5.6) and channel 1 calibration values. Data from HRPT NOAA-11, 20 March 1989, at location 31.09°N, 127.94°W.

T_s (K)	Atmospheric Correction	α_{sfc}	CH1 Calibration	$r(\mu m)$	τ	susceptibility $\times 10^3$ (cm ³)
290	B.L.→Space	.06	Nianzeng et al. -1988	13.6	28.9	2.33
	Sfc.→Space			10.0	83.3	0.492
	No Correction			17.5	16.8	0.587
	B.L.→Space		Holben & Kaufman - 1989	13.6	38.3	2.02

Sensitivity to weightings was also considered. Figure 6.3 shows isolines of error with $W_1=1$ and $W_3=2$ for the thin and moderate thickness clouds. In Figure 6.4 $W_1=1$ and $W_3=4$. The solution is tighter about radius for the latter combination of weights but the inferred cloud parameters are unchanged. A value of $W_3=3$ was adopted for all subsequent analysis.

A troublesome feature of these contour plots is the secondary minimum at $r=1$ μm (which always had a greater residual error than the minimum occurring at larger radius for these clouds). The channel 3 bidirectional plots of Figure 4.9 indicate that multiple solutions are possible when both radius and optical thickness are small. Referring to that figure, when considering the azimuthally averaged reflectance, a radius of 1 μm cannot be distinguish from a radius of 6 μm at $\tau \approx 4$. This is due to a smaller Q_{ext} at 1 μm compared with larger radii. Figure 4.10 shows that for all optical thicknesses the azimuthally dependent reflectance for 1 μm is reduced relative to other radii for $\phi > 120^\circ$. This helps to further lessen the distinction between the 1 μm reflectance and that of larger radii at these azimuth angles. It is caused by the phase function at 1 μm having less backscattering relative to larger radii. As a test of uniqueness, the forward problem was calculated for $\tau=4$, $r=10$ μm , $\mu_0=.65$, $\mu=.85$, and $\phi=135^\circ$ giving $R_1=.27$ and $R_3=.1125$. Using these values of reflectance in the inversion algorithm showed the global minimum for the error function to be at $\tau=2$ and $r=1$ μm . A local minimum, but with larger residual error, was found at the *correct* solution. Cloud microphysical measurements indicate that a mean drop radius of 1 μm is highly unlikely. So, eliminating the 1 μm radius solution from consideration will

ensure the more credible solution. This is the case, unless otherwise stated, in all results that follow.

6.1.2 Error Calculations Based on Chapter 5 Results

Approximations for atmospheric influences in the AVHRR channels was presented in chapter 5. Section 5.7 presented results for expected worst case errors in retrieved reflectance (see figures 5.16a-5.16e) due to uncertainties in these atmospheric values. These errors can be applied to the three cloud examples from the last section. For the thin stratus, retrieved channel 1 and 3 reflectances were approximately 0.16 and 0.028 respectively. Figure 5.16a for channel 1 and 5.16c for channel 3 (see the graph including AVHRR channel 4 temperature retrieval) indicate that for these reflectances, worst case errors in reflectance for the boundary layer to space path would be about ± 0.025 and ± 0.020 respectively. For the moderate and thick clouds, approximate values and uncertainties for the channel 1 reflectances were chosen as 0.47 ± 0.05 and 0.80 ± 0.07 respectively; for channel 3 they were 0.13 ± 0.04 and 0.074 ± 0.04 respectively.

A Monte Carlo analysis of retrieved cloud parameters based on these reflectance limits was calculated with a modified inverse code (R_3 is known apriori and is not calculated from I_3 and cloud temperature). Uniform probabilities for the true reflectance being anywhere within the intervals defined by Figure 5.16 were used. A total of 500 calculations were made for each cloud. The spread in the solutions was much greater than the analysis of the last section. For the thin cloud, retrieved radii

ranged from the $17.5\ \mu\text{m}$ solution to $45\ \mu\text{m}$; optical thickness remained limited to 1 or 2. Because of the bounds on radius and thickness, the retrieved susceptibility was limited to nine values ranging from 2.0×10^{-3} to 50.0×10^{-3} . The moderately thick cloud was found to have radii from $10.0\ \mu\text{m}$ to $15.0\ \mu\text{m}$; optical thickness varied from 4 to 10 and susceptibility was confined to thirteen values from 0.58×10^{-3} to 3.88×10^{-3} . Retrieved parameters were spread widely for the thick cloud. Radius and optical thickness ranged from $12.5\ \mu\text{m}$ to $30\ \mu\text{m}$ and 25 to 50 respectively. Susceptibility varied from 1.31×10^{-3} to 25.6×10^{-3} . The spread in retrieved radius, for all cases, indicates that the incremental spacings for radius used in the library are probably sufficient.

6.2 Comparisons with In Situ Measurements

As a part of the First ISCCP (International Satellite Cloud Climatology Project) Regional Experiment (FIRE), extensive field observations were made of marine stratocumulus clouds off the coast of southern California in the summer of 1987. Two reports have been published of cloud microphysical measurements taken at times almost concurrent with the pass of a NOAA polar orbiter. The measurements by Rawlins and Foot (1990) were taken over the course of several hours during a NOAA-9 overpass on the afternoon of 30 June 1987. An encounter with ship tracks by Radke et al. (1989) occurred on 10 July 1987 within 20 minutes after a NOAA-10 pass. Cloud parameters retrieved from high resolution images of both of these scenes are compared with in situ values reported by the authors.

The exact location of the measurements reported by Rawlins and Foot (1990) were obtained from the Meteorological Research Flight of the United Kingdom Meteorological Office (Taylor 1991, private communication). Several runs were made along an "L" shaped path within and above a cloud with a thickness of about 300 meters. Effective radii of about $9.0\ \mu\text{m}$ were calculated from drop size distributions near cloud top (at about 900 m) with a PMS Forward Scattering Spectrometer Probe (FSSP). Optical thickness, estimated from above cloud reflectance data, varied from 10 to 25. NOAA-9 AVHRR LAC image data was analyzed for the north-south flight path made from approximately 32.36°N , 123.25°W to 33.41°N , 123.32°W . The in-cloud measurements along this path were taken at about 2100, 2137, and 2332-2355 UTC; the NOAA-9 pass occurred between these times at 2218 UTC. For the region being analyzed, $\mu_0 \approx .88$, $\mu \approx .51$ (corresponding to a resolution of about 3.8 km), and $\phi \approx 8^\circ$. Because of the large satellite viewing angle, it was felt that the atmospheric corrections of Table 5.6, based on $\mu \approx 0.9$, needed to be modified. From the results of Chapter 5, atmospheric transmission was chosen to be 0.75, 0.82, and 0.88 in channels 1, 3, and 4 respectively. Other atmospheric terms were as follows: atmospheric reflection in channel 1 equal to 0.041, atmospheric reflected intensity in channel 3 equal to $0.01\ \text{mW}/(\text{m}^2\text{-Sr-cm}^{-1})$, emission in channel 3 equal to $0.02\ \text{mW}/(\text{m}^2\text{-Sr-cm}^{-1})$, and emission in channel 4 equal to $7.0\ \text{mW}/(\text{m}^2\text{-Sr-cm}^{-1})$. The channel 1 image is shown in Figure 6.5 with the flight path indicated. Figure 6.6 summarizes retrieved radius and optical thickness along the flight path using the NOAA-9 data and the algorithm presented in section 6.1. Data along the south half of the path shows larger radii than

that reported by Rawlins and Foot. Along the north half of the path comparison with in situ data is better. Here, radius varies from $8\text{ }\mu\text{m}$ to $10\text{ }\mu\text{m}$ ($r_{\text{eff}}=8.64$ to $10.8\text{ }\mu\text{m}$) and optical thickness is seen to typically be 10 to 30. At the far north end where optical thickness is within the range reported by Rawlins and Foot, the averaged mean radius would be about $8\text{ }\mu\text{m}$ ($r_{\text{eff}}=8.64\text{ }\mu\text{m}$). The $9\text{ }\mu\text{m}$ effective radius measurements of Rawlins and Foot represent an average over the flight path; no information is provided with regard to measurements at specific locations. However, the cloud is seen to have a mottled appearance throughout. Towards the southern end a hole (i.e., a relatively smaller albedo) is seen in the cloud deck indicating inhomogeneities in that region of the cloud. Ignoring the southern end of the flight path, agreement between retrieved values of radius and in situ measurements are within the radius increments available from the library data (at a radius of $10\text{ }\mu\text{m}$, increments of $+2.5$, $-2.0\text{ }\mu\text{m}$ in radius are possible to resolve with the library). Susceptibilities (for a cloud liquid water content of 0.3 g/m^3) vary over an order of magnitude, from about 0.5×10^{-3} at the northern end of the path to 6.0×10^{-3} at about a quarter of the way from the southern end. Using the atmospheric corrections of Table 5.6 instead of the corrections given above lowered the retrieved optical thickness (as expected since atmospheric transmission and reflectance are now less) but had no effect on the retrieved radius.

Enroute to a FIRE mission (Radke et al. 1989, King et al. 1990) reported passage through two ship tracks within a stratocumulus cloud layer. Radiation and microphysical measurements were taken midway between the approximately 500 meter thick cloud as the aircraft crossed the tracks. Drop concentrations were seen to increase

from 30 to 50/cm³ outside the track to over 100/cm³ within the track, indicative of larger CCN numbers. Liquid water content, optical thickness, and effective radius are summarized in Table 6.2 based on data presented by Radke et al. and assuming a cloud thickness of 500 meters for the optical thickness calculation (the parenthesis values were based on radiation measurements reported by King et al. (1990)). The track referred to as being the *first* track in the paper is evidently the one farther to the east. Figure 6.7 shows the HRPT NOAA-10 channel 1 image, where for the region being analyzed, $\mu_0 \approx .50$, $\mu \approx .79$ (corresponding to a resolution of about 1.7 km), and $\phi \approx 165^\circ$. Cloud parameters were retrieved along two section lines, *A* and *B*, shown in the figure. Results for radius and optical thickness along the sections, graphed in Figures 6.8a and 6.8b, show larger values for both parameters compared with the measurements.

Microphysical studies in stratus typically show increasing drop sizes with height (e.g., Curry 1986; Rawlins and Foot 1990; Slingo, Nicholls, and Schmetz 1982). Because of the large absorption at 3.75 μm , it is likely that drop sizes near cloud top contribute a far greater weighting to the inferred sizes than drops further down in the cloud. The vertical location of scatterings for reflected photons in a cloud with a single scattering albedo of 0.890 (corresponding to a 10 μm radius at 3.7 μm) was found using a Monte Carlo simulation. An exact definition for the vertical weighting of cloud drop sizes needed to give retrieved radii is arguable. One possible weighting is as follows. Consider a single photon that is ultimately reflected from the cloud. The contribution of any layer in the cloud to the retrieved radius is proportional to the number of scatterings by that photon taking place in the layer and inversely proportional

to the strength of the emerging photon (which is in turn proportional to the total number of scatterings since a reflected photon with a large number of total scatterings will be too attenuated to contain any information regarding the layers with which it interacted). Monte Carlo calculations were made using this definition for the contribution, or weighting of the layers for a cloud with $\tau=8$ and $\omega_0=.890$. Reflectance has reached its asymptotic limit by this thickness and so thicker clouds will have the same weighting. More than 90 % of the integrated weight is confined to $\tau < 4$. This could account for some of the difference seen between the retrieved radii and those found at mid-cloud where measurements show radii might be expected to be a few microns smaller (e.g., Rawlins and Foot 1990). A few micron difference between mid-cloud and cloud top radii can begin to explain the 3 to 5 μm size differences seen in Table 6.2. For comparison, in the visible 90 % of the integrated weight is confined to $\tau < 7$ for a total thickness of 10 and to $\tau < 28$ for a thickness of 100.

It has already been shown that retrieval of optical thickness is very sensitive to large channel 1 reflectances (i.e., $dA/d\tau$ is small). For example, if no atmospheric correction is used (which acts to reduce the inferred reflectance) optical thickness in the tracks is reduced from over 100 to about 30. This doesn't necessarily suggest that the atmospheric correction is too strong (or even weak), but merely emphasizes the need for accurate atmospheric corrections when retrieving optical thickness from clouds with large reflectances. Susceptibilities give the expected results. Out of the track values average 1.2×10^{-3} just to the east of the tracks, 1.1×10^{-3} between the tracks, and 1.9×10^{-3} just to the west of the tracks. Average susceptibilities in the tracks are 0.5×10^{-3} , a

factor of 2 to 4 less than in the surrounding cloud. Average results from these sections are given in the Table 6.2 along with the calculated susceptibilities. The cloud parameter values from Radke et al. can also be used to calculate susceptibility, giving a similar range of values of about 0.2×10^{-3} to 0.64×10^{-3} in-track and 0.72×10^{-3} to 1.44×10^{-3} out of the track.

The liquid water path (g/m^2) for a uniform cloud is the product of the geometrical thickness h and the liquid water content. The liquid water content (in g/m^3) is $r\tau/1.5h$, with h in meters and radius in microns. Radke et al. measured higher liquid water contents within the ship tracks than in the surrounding cloud. The retrieved data of Table 6.2 shows the ship tracks with about twice the liquid water path compared with the out-of-track value. But liquid water contents, assuming a geometrical cloud thickness of 500 m, are about 1.3 g/m^3 out-of-track and 2.5 g/m^3 in-track - almost an order of magnitude larger than the measured values.

In summary, while the retrieved parameters agree well with the measurements of Rawlins and Foot (1990) over part of the flight path, comparisons with the measurements of Radke et al. (1989) are poor. Nevertheless, even in the latter case, relative retrievals were as measured; radii and susceptibilities were smaller and optical thickness and liquid water content larger in the ship tracks.

Table 6.2 Comparison of NOAA-10 retrieved parameters with the measurements of Radke et al. (1989) in a stratocumulus cloud containing ship tracks.

Parameter	Track #1	Track #2	Out of Track
<i>Measurements from Radke et al. (1989) (taken at mid-cloud)</i>			
r_{eff} (μm)	10.5	7.5	11.2
τ	40 (72)	33 (48)	13-20 (18)
LWC (g/m^3)	0.3-0.7	0.3-0.5	0.20-0.40
<i>Retrieved parameters from NOAA-10 using the algorithm of this study</i>			
r_{eff} (μm)	13.5	13.5	13.5-16.2
τ	140	140	30-90
Susceptibility ($\times 10^3$)	0.6	0.5	1.1-1.9

6.3 Marine Stratus and Ship Tracks

Marine stratus clouds that contain ship tracks would be expected to have relatively high values of susceptibility, while the tracks, having been contaminated by ship exhaust, should have smaller values. Such stratus provides a useful test for the remote sensing of susceptibility. With large variations in susceptibility over a small scale, the solar and satellite viewing angles, cloud temperatures, surface properties, and atmospheric influences are the same and so relative measurements of susceptibility are

more credible than comparisons between distant regions. As discussed in section 2.2.2, it is probable that high susceptibility is not necessarily sufficient for the formation of tracks. Usually, tracks are not seen in satellite images of marine stratus despite the regular presence of ship traffic. So it is likely that certain meteorological conditions are also required for the tracks to form. The remote sensing of susceptibility can provide some insight into this matter.

6.3.1 Marine Stratus Near the West Coast of the United States

The results of Table 6.2, for the ship tracks measured by Radke et al. (1989), demonstrate the reduced susceptibility expected for ship tracks. Using NOAA-10 retrieved parameters, out-of-track regions have susceptibilities 2 to 4 times greater than in-track. Similar relative differences are calculated using the Radke et al. data. A uniform region of stratus to the north west of the tracks (from approximately 35.67°N, 125.23°W to 36.51°N, 123.84°W) contained no apparent ship tracks. Analysis of three locations in this region showed radii from 15 μm to 17.5 μm , optical thickness from 25 to 80, and susceptibilities from 1.32×10^{-3} to 3.95×10^{-3} , which is in the same range as the stratus in which the tracks were found. Assuming the presence of ships in this area, the retrievals suggest a meteorological reason, instead of a microphysical one, for the inability of tracks to form.

The tracks of Radke et al. are hard to discern in the satellite imagery and, without in situ data, it might be difficult to conclude that they are indeed ship tracks.

The two well defined, intersecting tracks that formed in the thin stratus described in section 6.1.1 are obvious (see Figure 6.9). Three locations in the tracks were analyzed using six to fifteen pixels at each position. At the intersection of the two tracks, the average retrieved radius, optical thickness, and susceptibility for the pixels analyzed were $8.0 \mu\text{m}$, 13.0, and 0.57×10^{-3} respectively. In the center of the track that was farther to the north, averaged retrieved values were $8 \mu\text{m}$, 4.2, and 0.50×10^{-3} , while towards the end of the southern track, averaged values were $13.2 \mu\text{m}$, 6.1, and 2.51×10^{-3} . These susceptibilities are less than those reported in Table 6.1a by a factor of 2 to 30. Note that the average susceptibility is not necessarily equivalent to the susceptibility calculated from the average radius and optical thickness. The stratus just to the west of the tracks is the apparent origin of the ships but no tracks can be seen in either in channel 1 or 3. Table 6.1b shows susceptibilities in this region to be as small as the values found in the tracks. In this case, the small susceptibilities might be sufficient to explain the lack of observable tracks.

A dramatic instance of ship tracks was seen on 2 March 1990 in a large region centered near 52°N , 140°W . The HRPT NOAA-11 image is shown in Figure 6.10. Viewing angles are $\phi \approx 60^{\circ}$, $\mu_0 \approx 0.50$, and $\mu \approx 1.0$ to the far right (1.1 km resolution) and $\mu \approx 0.64$ (2.4 km resolution) to the far left. Tracks are seen to be forming in the uniform stratus to the west as well as in the nearly cloud-free regions in the center of the image. A number of locations in the image have been analyzed and are indicated in the figure. Average results for out-of-track locations, designated by letters, are summarized in Table 6.3a. Liquid water path is included in the table. Three to five

pixels were typically used to calculate the averages. In the western stratus, susceptibilities range from about 3×10^{-3} to 7×10^{-3} . In-track results are summarized in Table 6.3b and were calculated with a four pixel average when possible. Susceptibilities range from about 1×10^{-3} to 2×10^{-3} for the tracks in the western stratus. Tracks in the thin stratus show some smaller susceptibilities and overall smaller liquid water paths.

6.3.2 Some Observations on Ship Tracks

Though the emphasis of this study is on susceptibility, the results for the 2 March 1990 image allow for a few comments regarding ship tracks themselves. There is some suggestion of horizontal entrainment of the track into nearby regions. The first two tracks between locations *G* and *H* (going from east to west) appear to be younger tracks than the last two, and have smaller radii and susceptibilities. For locations 1, 2, and 3, optical thickness decreased slightly moving away from the head of the track. However, in some cases no significant or systematic change in the retrieved values can be noted along the length of a track. No difference is seen between locations 5 and 6, though further analysis showed retrieved radius and optical thickness of $8 \mu\text{m}$ and 3, respectively, at the head of the track. The track referenced by location 8 showed a radius of $6.0 \mu\text{m}$ farther to the south, then increasing again to $10.0 \mu\text{m}$ closer towards the head of the track; optical thickness increased to 14 and then decreased to 7. A general conclusion from Table 6.3b, but not a rule, is that the more diffuse or widespread the track, the larger the radii and the susceptibility (note however, that the

narrow track between locations *I* and *J* has a relatively large radius and susceptibility compared with other narrow tracks). There appears to be some limit to the horizontal spreading of the tracks, at least over relatively short periods of time. For example, the tracks running from north to south between locations *G* and *H* show little spread over a distance of about 100 km; similarly for the track between locations *I* and *J*. Radke et al. (1989) state that ship tracks are not brighter at their head and suggest that gases responsible for CCN production horizontally diffuse before particles develop. The results for locations 1 through 4, mentioned above, refute the generality of this notion. For that track, optical thickness increased (i.e., brighter track) towards the head of the track though radius remained constant. However, this was not necessarily true for other tracks. It may also be that CCN (not the precursor gases) can diffuse before nucleation occurs; the driving force for nucleation is not clear and should not be presumed to be uniform spatially or temporally. Still, the fine point seen at the head of many tracks, and the results from locations 1 through 4, does indicate that nucleation often occurs before CCN, or gases responsible for CCN production, are allowed to diffuse substantially.

The difference in radii between in-track and out-of-track cloud regions ranged from about 2 to 10 μm . The following relation exists between changes in radius and cloud drop concentration:

$$1 + \frac{\Delta r}{r} = \left(1 + \frac{\Delta N}{N}\right)^{-\frac{1}{3}}.$$

So for typical cases of $\Delta r/r = -0.15$, $\Delta N/N = 0.63$, and for $\Delta r/r = -0.3$, $\Delta N/N = 1.9$.

Radke et al. (1989) measure increases in liquid water content in ship tracks compared with the surrounding cloud. Liquid water paths given in Table 6.3 show similar results for the western stratus, assuming constant cloud geometrical thickness in and out of the track. Based on results from Albrecht (1989), Radke et al. proposed that the smaller drop sizes in ship tracks would suppress the formation of drizzle drops, resulting in the higher liquid water contents. It is also possible that some aspect of the ship (e.g., the warm plume) locally modify the boundary layer and enhance cloud formation. When tracks form in mature stratus, supersaturations would normally be too low to activate new CCN and some mechanism must allow for the scavenging of existing liquid water for new drop growth. The tracks seen in the center of the image have similar radii and optical thicknesses to those in the thicker stratus to the west, but yet have formed in a region that was evidently not as conducive to substantial stratus development. If the ship itself is partially responsible for allowing activation of new CCN, other aspects of the track, such as its liquid water content may also be modified. If so, implicating the observable increase in liquid water content with the suppression of drizzle may not always be valid. With the special circumstances involved in ship track development, it is not at all clear that the global susceptibility of marine stratus would be effected by this drizzle suppression. However, if it proved to be true this would serve to increase susceptibilities as discussed in section 2.2.4. Estimates in that

section indicated a factor of two increase using the measurements of Radke et al. as an example.

Table 6.3a Retrieved results for out-of-track regions for NOAA-11 HRPT, 2 March 1990. Locations are given in Figure 6.10.

Location	Radius (μm)	Optical Thickness	Susceptibility ($\times 10^{-9}$)	Liquid Water Path (g/m^2)
A	15.6	4.9	3.42	51
B	14.7	8.1	2.87	79
C	15.6	5.7	3.42	59
D	17.2	6.0	4.61	69
E	14.2	6.4	2.60	61
F	18.3	4.0	5.62	49
G	19.2	4.0	6.74	51
H	15.0	5.8	3.30	59
I	19.4	6.7	6.63	87
J	18.9	7.3	6.10	92

Table 6.3b Retrieved results for in-track regions for NOAA-11 HRPT, 2 March 1990. Locations are given in Figure 6.10. Tracks found between letter designated locations are given in order from east to west.

Location	Radius (μm)	Optical Thickness	Susceptibility ($\times 10^{-3}$)	Liquid WaterPath(g/m^2)
A-B	12.5	10.8	1.72	90
C-D	10.0	10.0	0.89	67
E-F	13.0	13.6	2.02	118
	12.5	18.3	1.59	153
G-H	10.0	10.0	0.98	67
	10.0	17.0	0.91	113
	13.2	15.9	2.21	140
	12.5	15.2	1.83	127
I-J	11.0	13.0	1.43	95
1	8.0	9.5	0.46	51
2	8.0	7.3	0.46	39
3	8.0	6.4	0.46	34
4	10.0	6.0	0.90	40
5	10.0	7.8	0.90	52
6	10.0	8.5	0.89	57
7	12.5	8.3	1.75	69
8	10.0	11.3	0.88	75
9	12.5	10.1	1.73	84

6.3.3 Marine Stratus Near Southern Africa

Marine stratus is quite common, especially near the western coasts of the continents in both hemispheres. Several GAC images were obtained for stratus in the vicinity of southern Africa that was initially found in photographs published in the *Meteosat Image Bulletin*. The susceptibility of these clouds is of interest. As it turned out, ship tracks appear in at least one of the images and perhaps in a second.

Linear features appearing to be ship tracks show up in a channel 3 NOAA-11 GAC image, from 4 January 1989, off the west coast of South Africa (see Figure 6.11). However, the features are not apparent in the channel 1 image which shows the stratus to be fairly uniform. Several retrievals across these cloud lines were made. For purposes of discussion, these lines will be referred to as tracks. Radii were found to be quite small, invariably being $8\text{ }\mu\text{m}$ throughout the out-of-track regions examined. In-track pixels had smaller radii, as expected for ship tracks, of $6\text{ }\mu\text{m}$. However, contrary to expectation, optical thickness was as high as 12 outside the track but decreased to about 3 to 7 in-track. The largest susceptibility found was about 0.7×10^{-3} , smaller than expected for the development of ship tracks (at least when using California stratus as a guide). The results are mixed; radius retrievals support the notion of increased CCN but optical thicknesses are actually smaller in the tracks and susceptibilities are small throughout the stratus.

Figure 6.12 shows channel 1 and 3 NOAA-11 GAC images of two apparent tracks on 22 January 1989, located south of Madagascar in the Indian Ocean at 33.30°S , 45.71°E . The channel 1 image shows a broken stratocumulus deck while the channel

3 image appears more uniform. With channel 3, the characteristic narrowing of the cloud near the head of the ship can be seen for both tracks. Satellite viewing angles are near nadir in this region giving about 4 km resolution. In channel 1, it appears that this resolution is just able to make out the cloud structure; it is difficult in such a situation to be sure that the cloud cover fraction for any pixel will be one or zero. Compared with a cloud cover fraction of one, retrievals for pixels with partial cloud cover will give smaller optical thicknesses and larger radii.

Four sections across the tracks were analyzed; two across the southern track and two across the northern track. In the southern track, retrieved optical thickness and radius was 8 to 15 and $8\text{ }\mu\text{m}$ to $10\text{ }\mu\text{m}$ respectively; susceptibilities were 0.6×10^{-3} to 1.2×10^{-3} . Out-of-track retrievals gave optical thicknesses of 3 to 12 and radii from $12.5\text{ }\mu\text{m}$ to $17.5\text{ }\mu\text{m}$; susceptibilities were always larger than for the in-track pixels and went as high as 6.5×10^{-3} . This analysis of clouds outside the track ignores pixels with retrieved optical thickness of 1, assuming that this implies a small cloud cover fraction and so is not indicative of the actual clouds in that area. A cloud cover fraction of one can most likely be associated with those places where adjacent pixels had the larger optical thicknesses. There seems then, to be an obvious reduction in drop size in the southern track. Susceptibilities are also consistent with the expectations for ship tracks, and are comparable to those found in California stratus.

For the northern track, retrieved optical thicknesses were 1 to 6 out of the track and 2 to 11 in the track. It can be concluded that either the clouds were quite sometimes quite thin or cloud cover fraction was less than one. Nevertheless, the

retrieved in-track radius was $10\ \mu\text{m}$ while out-of-track radii ranged from a low of $10\ \mu\text{m}$ to as high as $20\ \mu\text{m}$ (always associated with $\tau=1$). Again this suggests that radii in the track were, in fact, less than those out of the track. Susceptibilities were similar to values reported for the southern track.

Analysis of both tracks supports the view that they were caused by ships. To this author's knowledge, this would be the first report of ship tracks being found in the southern hemisphere.

The stratus of 4 January 1989 ranges from South Africa to Namibia along the west coast of Africa. The low susceptibilities are indicative of high numbers of CCN, yet this air would be expected to be relatively clean unless there had been contamination from land sources. Pictures in the Meteosat Bulletin show the extent of the stratus to have been unchanged since the previous day. It seems useful to look at other examples of stratus in this region to see if these small susceptibilities were an exception. A smaller, isolated patch of stratus on 19 April 1989 off the coast of Namibia (at about 22.0°S , 13.46°E) was analyzed and found to have somewhat larger drop sizes than the January image. Five 2 by 2 pixel areas were chosen and found to have radii ranging from $6.0\ \mu\text{m}$ to $9.0\ \mu\text{m}$ with an average of $7.75\ \mu\text{m}$, optical thicknesses from 3.50 to 9.25, and susceptibilities from 0.23×10^{-3} to 0.81×10^{-3} . It is impossible to make any climatological conclusion regarding the microphysics of this stratus, but for the two cases studied, radii and susceptibilities are less than those typically found in retrievals from uncontaminated California stratus. If this were generally true, it might explain the

lack of ship track sightings in this region even though stratus development is relatively common.

6.4 California Valley Fog

Wintertime fog throughout the central valleys of California provides a uniform continental stratus cloud for satellite remote sensing. Unlike the clean maritime clouds of the last section, these fogs can be expected to be formed in air containing large concentrations of CCN, both from agricultural, industrial, and natural sources. Measurements of fog microphysics generally show less liquid water content, smaller drop sizes, and smaller drop concentrations than for clouds (Pruppacher and Klett 1980). Using a cascade impactor to capture drops on slides, Garland (1971) made measurements in both advection and radiation fog. Effective radii can be calculated from his Table 1 which gives mass concentrations and visual range. For radiation fog, the effective radius varied from $3.34\ \mu\text{m}$ to $10.38\ \mu\text{m}$ for the fogs studied. Drop concentrations varied from $330\ \text{cm}^{-3}$ to over $3000\ \text{cm}^{-3}$. Roach et al. (1976) found liquid water contents in radiation fog at Cardington, England to range from $0.05\ \text{g/m}^3$ to over $0.20\ \text{g/m}^3$. Drop size distributions in both of the last two studies show a characteristic lower peak at less than $1\ \mu\text{m}$ and a smaller upper peak at $10\ \mu\text{m}$ or greater. Gerber (1980) shows a similar size distribution for radiation fog near Reston, Virginia.

Three NOAA-11 GAC images of California valley fog from the winter of 1989/1990 were analyzed. The inferred radius and optical thickness, along sections

down the center of the central valley, are given in Figure 6.13 for 20 December 1989 (with $\phi \approx 125^\circ$, $\mu_0 \approx 0.45$, and $\mu \approx 1.0$ corresponding to a resolution of about 4 km), and 11 January 1990 (with $\phi \approx 60^\circ$, $\mu_0 \approx 0.50$, and $\mu \approx 0.70$ giving a resolution of about 8 km). These sections represent distances of about 235 km and 120 km respectively. Because studies indicate the possibility of small mean drop sizes, entries for 1 μm radii in the libraries were included in the solutions space for the analysis. Retrieved radius is especially uniform throughout the length of the valleys, with typically 6 μm mean radii - smaller than for marine stratus as expected and in general agreement with the results of Garland (1971). Analysis of a central valley fog on 25 January 1990 gave average mean radii of 5.3 μm to 7.4 μm for three locations examined. The average optical thicknesses ranged from 38 to 96. Susceptibilities were relatively small as expected; 0.12×10^{-3} to 0.43×10^{-3} . Along the section from 20 December 1989, susceptibilities were a maximum of 0.16×10^{-3} at the south end of the valley to 0.05×10^{-3} at the north end. For the section from 11 January 1990, susceptibility ranged from 0.50×10^{-3} in the south to 0.07×10^{-3} midway along the section at $r=6 \mu\text{m}$ and $\tau=90$. These values are one to two orders of magnitude less than those given in Table 6.3.

As previously mentioned, the liquid water content (in g/m^3) for a uniform cloud with geometrical thickness h is $r\tau/1.5h$ where radius is in microns and h is in meters. Assuming h to be 500 m, the estimated liquid water content for the fogs range from 0.16 g/m^3 (with $r=6 \mu\text{m}$ and $\tau=20$) to 1.5 g/m^3 ($r=8 \mu\text{m}$, $\tau=140$). The latter value seems much higher than measurements suggest and could be due to a

geometrically thicker fog or the error in retrieval of τ that accompanies large channel 1 reflectances.

6.5 Conclusions

A definition of cloud *susceptibility* has been proposed as a measure of the extent of CCN influence on cloud albedo. The definition was originally couched in terms of τ , $dA/d\tau$, and cloud drop concentration N . However, N cannot be found directly by remote sensing clouds but must be inferred from the drop radius, which can be sensed, and an assumption regarding the value of cloud liquid water content. Susceptibilities were typically normalized to a liquid water content of 0.3 g/m^3 , a representative value for marine stratus.

Retrieval of cloud optical thickness and drop radius is, in itself, a justification for developing this remote sensing technique. The parameters are useful in most cloud studies and were presented in this chapter along with the calculation of susceptibility.

Retrieved radius and optical thickness using the method of this study was compared with in situ measurements made by two groups of investigators during FIRE in 1989. Retrievals agreed well with the measurements of Rawlins and Foot (1990) over the northern half of their flight path. Comparisons with the measurements of Radke et al. (1989) showed larger retrieved radii and optical thicknesses. The error in optical thickness can be considerable for clouds with large reflectances, which was the case here. Radke et al. sampled midway in the cloud which was likely to have had smaller drop sizes than at cloud top.

The retrieved range of susceptibilities (in units of cm^3) for the marine stratus studied varied by about two orders of magnitude; from as low as 0.23×10^{-3} in stratus off the west coast of southern Africa to about 20×10^{-3} in the thin stratus of Figure 6.9. Susceptibilities for California valley fog were as low as 0.05×10^{-3} , extending the measured range, for all clouds studied, to almost three orders of magnitude. Studies in ship tracks have shown that the tracks are indeed less susceptible than out-of-track regions after having been contaminated with CCN originating from the ship's effluent. The susceptibility, in and out of the track, may differ by only a factor of 2 to 4 (see Table 6.3) or as high as 30 for the thin stratus of Figure 6.9. The use of susceptibility extends beyond fog and marine stratus which have been highlighted in this study because of their relative homogeneity for remote sensing purposes.

Figure 6.1 The Forward Problem

Figure 6.2 A Method for Solving the Inverse Problem

Figure 6.3 Isolines of residual error using the error definition given in Figure 6.2 with $W_1=1$ and $W_3=2$. (a) For the thin stratus and (b) the moderately thick stratus analyzed in the text.

Figure 6.4 Isolines of residual error using the error definition given in Figure 6.2 with $W_1=1$ and $W_3=4$. (a) For the thin stratus and (b) the moderately thick stratus analyzed in the text.

Figure 6.5 NOAA-9 LAC, 30 June 1987, AVHRR channel 1. North is to the top of the page.

Figure 6.6 Retrieved mean radius and optical thickness along the flight path of Rawlins and Foot (1990) in stratocumulus off southern California. From NOAA-9 LAC, 30 June 1987.

Figure 6.7 NOAA-10 HRPT, 10 July 1987, AVHRR channel 1. North is to the top of the page.

(a)

(b)

Figure 6.8 Retrieved mean radius and optical thickness for two sections perpendicular to two ship tracks encountered by Radke et al. (1989). For NOAA-10 HRPT, 10 July 1987. Figure (a) is a more northerly section than is (b).

Figure 6.9 NOAA-11 HRPT, 20 March 1989, AVHRR channel 1. North is to the top of the page.

Figure 6.10 NOAA-11 HRPT, 2 March 1990, AVHRR channel 1. North is to the top of the page.

Figure 6.11 NOAA-11 GAC, 4 January 1989. AVHRR channel 1 is to the left, channel 3 to the right. North is to the top of the page. The land mass to the east is southern Africa.

Figure 6.12 NOAA-11 GAC, 22 January 1989. AVHRR channel 1 is to the left, channel 3 to the right. North is to the top of the page. Madagascar is the land mass to the far north.

(a)

(b)

Figure 6.13 Retrieved mean radius and optical thickness for California valley fog. (a) Section along the center of the central valley on 20 December 1989, and (b) on 11 January 1990, for NOAA-11 GAC.

LIST OF REFERENCES

- Abel, P., 1990a: Private Communication. *NOAA/NESDIS*.
- Abel, P., editor, 1990b: Report of the Workshop on Radiometric Calibration of Satellite Sensors of Reflected Solar Radiation. *NOAA Technical Report NESDIS 55*.
- Ahmad, Z., and Fraser, R. S., 1982: An Iterative Radiative Transfer Code For Ocean-Atmosphere Systems. *J. Atmos. Sci.*, 39, 656-665.
- Albrech, B. A., Penc, R. S., and Schubert, W. H., 1985: An Observational Study of Cloud-Topped Mixed Layers. *J. Atmos. Sci.*, 42, 800-822.
- Albrecht, B. A., Randall, D. A., and Nicholls, S., 1988: Observations of Marine Stratocumulus Clouds during FIRE. *Bull. Amer. Met. Soc.*, 69, 618-626.
- Albrecht, B. A., 1989: Aerosols, Cloud Microphysics, and Fractional Cloudiness. *Science*, 245, 1227-1230.
- Altshuler, T. L., 1961: *Infrared Transmission and Background Radiation by Clear Atmospheres*. GE Report 61 SD 199, AD-401923.
- Arking, A., 1991: The Radiative Effects of Clouds and their Impact on Climate. *Bull. Amer. Met. Soc.*, 71, 795-813.
- Arking, A., and Grossman, K., 1972: The Influence of Line Shape and Band Structure on Temperatures in Planetary Atmospheres. *J. Atmos. Sci.*, 29, 937-949.
- Arking, A., and Childs, J. D., 1985: Retrieval of Cloud Cover Parameters from Multispectral Satellite Images. *J. Climate and Appl. Meteo.*, 24, 322-333.
- Baldwin, D., 1991: Personal Communication. University of Colorado, Boulder.
- Bate, R., Mueller, D., and White, J., 1971: *Fundamentals of Astrodynamics*, Dover Publications, New York.
- Ben-David, A., and Thome, K., 1990: Private Communication. Institute of Atmospheric Physics, University of Arizona, Tucson.
- Bignell, K. J., 1970: The Water-Vapour Infra-Red Continuum. *Quart. J. R. Met. Soc.*, 96, 390-403.

- Bohren, C. F., 1980: Multiple Scattering of Light and Some of its Observable Consequences. *Am. J. Phys.*, 55, 524-533.
- Bradley, S. G., 1988: Private Communication, Department of Physics, University of Auckland, New Zealand.
- Brown, O. B., Brown, J. W., and Evans, R. H., 1985: Calibration of the Advanced Very High Resolution Radiometer Infrared Observations. *J. Geophys. Res.*, 90, 11,667-11,677.
- Burch, D. E., 1971: Investigation of the Absorption of Infrared Radiation by Atmospheric Gases: Water, Nitrogen, Nitrous Oxide. *AFCRL-71-0124*, AD 882876.
- Burch, D. E., 1982: Continuum Absorption by H₂O. *AFGL-TR-81-0300*, ADA 112264.
- Carlson, H. R., and Harden, C. S., 1980: Mass Spectrometry of Ion-Induced Water Clusters: An Explanation of the IR Continuum Absorption. *Appl. Opt.*, 19, 1776-1786.
- Cess, R. D., 1990: Climate Change: An Appraisal of Atmospheric Feedback Mechanisms Employing Zonal Climatology. *J. Atmos. Sci.*, 33, 1831-1843.
- Chamberlain, J. W., and Hunten, D. M., 1987: *Theory of Planetary Atmospheres, An Introduction to Their Physics and Chemistry*. Academic Press, New York.
- Charlson, R. J., Lovelock, J. E., Andreae, M. O., and Warren, S. G., 1987: Oceanic Phytoplankton, Atmospheric Sulphur, Cloud Albedo and Climate. *Nature*, 326, 655-661.
- Coakley, J. A., Bernstein, R. L., and Durkee, P. A., 1987: Effect of Ship-Stack Effluents on Cloud Reflectivity. *Science*, 237, 1020-1022.
- Conover, J. H., 1966: Anomalous Cloud Lines. *J. Atmos. Sci.*, 23, 778-785.
- Conover, J. H., 1969: New Observations of Anomalous Cloud Lines. *J. Atmos. Sci.*, 26, 1153,1154.
- Curry, J. A., 1986: Interaction Among Turbulence, Radiation and Microphysics in Arctic Stratus Clouds. *J. Atmos. Sci.*, 43, 90-106.

- Dinger, J., Howell, H. B., and Wojciechowski, T. H., 1970: On the Source and Composition of Cloud Nuclei in an Subsident Air Mass over the North Atlantic. *J. Atmos. Sci.*, 27, 791.
- Duda, D. P., Stephens, G. L., and Cox, S. K., 1991: Microphysical and Radiative Properties of Marine Stratocumulus from Tethered Balloon Measurements. *J. Appl. Met.*, 30, 170-186.
- Fairall, C. W., Hare, J. E., and Snider, J. B., 1990: An Eight-Month Sample of Marine Stratocumulus Cloud Fraction, Albedo, and Integrated Liquid Water. *J. Climate*, 3, 847-864.
- Feigelson, E. M., 1978: Preliminary Radiation Model of a Cloudy Atmosphere, 1, Structure of Clouds and Solar Radiation. *Beitr. Phys. Atmos.*, 51, 203-229.
- Foot, J. S., 1988: Some Observations of the Optical Properties of Clouds, I: Stratocumulus. *Quart. J. R. Met. Soc.*, 114, 129-144.
- Fouquart, Y., Buriez, J. C., and Herman, M., 1990: The Influence of Clouds on Radiation: A Climate-Modeling Perspective. *Reviews of Geophysics*, 28, 145-166.
- Garland, J. A., 1971: Some Fog Droplet Size Distributions Obtained by an Impaction Method. *Quart. J. R. Met. Soc.*, 97, 483-494.
- Gathman, S. G., 1983: Optical Properties of the Marine Aerosol as Predicted by the Navy Aerosol Model. *Optical Eng.*, 22, 57-62.
- Gerber, H. E., 1980: Microstructure of a Radiation Fog. *J. Atmos. Sci.*, 38, 454-458.
- Ghan, S. J., Taylor, K. E., Penner, J. E., and Erickson, D. J. III, 1990: Model Test of CCN-Cloud Albedo Climate Forcing. *Geophys. Res. Lett.*, 17, 607-610.
- GOES Data User's Guide 1984: edited by J. Gibson, NOAA/NESDIS.
- Gorbinet, G., and Serpoly, R., 1985: Influence of an Industrial Site on the Characteristics of CCN Supersaturation Spectra Associated with Air Masses of Different Origin. *J. Rech. Atmos.*, 19, 193-202.
- Grainger, R. G., 1990: *The Calculation of Cloud Parameters from AVHRR Data*. Ph.D. Dissertation, University of Auckland, New Zealand.
- Grassl, H., 1973: Separation of Atmospheric Absorbers in the 8-13 Micrometer Region. *Beiträge zur Physik der Atmosphäre*, 46, 75-88.

- Grassl, H., 1982: The Influence of Aerosol Particles on Radiation Parameters of Clouds. *Időjárás*, Journal of the Hungarian Met. Service, 86.
- Hale, G. M., and Querry, M. R., 1973: Optical Constants of Water in the 200-nm to 200- μ m Wavelength Region. *Appl. Opt.*, 12, 555-562.
- Hansen, J. E., and Travis, L. D., 1974: Light Scattering In Planetary Atmospheres. *Space Science Reviews*, 16, 527-610.
- Heck, P. W., Byars, B. J., and Young, D. F., 1990: Climatology of Satellite-Derived Cloud Properties over Marine Stratocumulus Regions. *AMS Seventh Annual Conference on Atmospheric Radiation*, San Francisco.
- Herman, B. M., and Browning, S. R., 1965: A Numerical Solution to the Equation of Radiative Transfer. *J. Atmos. Sci.*, 22, 559-566.
- Hobbs, P. V., Bowdle, D. A., and Radke, L. F., 1985: Particles in the Lower Troposphere over the High Plains of the United States. Part II: Cloud Condensation Nuclei. *J. Climate and Appl. Meteo.*, 24, 1358-1369.
- Hobbs, P. V., 1971: Simultaneous Airborne Measurements of Cloud Condensation Nuclei and Sodium-Containing Particles over the Ocean. *Quart. J. R. Met. Soc.*, 97, 263-271.
- Hobbs, P. V., Stith, J. L., and Radke, L. F., 1980: Cloud-Active Nuclei from Coal-Fired Electric Power Plants and Their Interaction with Clouds. *J. Appl. Met.*, 19, 439-451.
- Holben, B. N., and Kaufman, Y. J., 1991: NOAA-11 AVHRR Visible and Near-IR Inflight Calibration. *Int. J. Remote Sens.*, in press.
- Holben, B. N., and Kaufman, Y. J., 1977: NOAA-11 AVHRR Visible and Near-IR Inflight Calibration. *Int. J. Remote Sens.*,
- Howell, H. B., and Jacobowitz, H., 1970: Matrix Method Applied to the Multiple Scattering of Polarized Light. *J. Atmos. Sci.*, 27, 1195-12-6.
- Howell, H. B., and Jacobowitz, H., 1971: On the Matrix Method Applied to the Multiple Scattering of Polarized Light. *J. Atmos. Sci.*, 28, 1301-1303.
- Jursa, A. S., editor, 1985: Handbook of Geophysics and the Space Environment. *Air Force Geophysics Laboratory*.

Justus, C. G., 1989: An Operational Procedure for Calibration and Assessing the Stability and Accuracy of Shortwave Satellite Sensors. *NOAA Cooperative Agreement NA84AA-H-00010*.

Kaufman, Y. J., and Holben, B. N., 1991: Calibration of the AVHRR Visible and Near-IR Bands by Atmospheric Scattering, Ocean Glint, and Desert Reflection. *Int. J. Remote Sens.*, in press.

Kells, L. M., Kern, W. F., and Bland, J. R., 1940: *Plane and Spherical Trigonometry*. McGraw-Hill, New York, 1940.

King, M. D., Nakajima, T., and Radke, L. F., 1990: Optical Properties of Marine Stratocumulus Clouds Modified by Ship Track Effluents. *AMS Seventh Conference on Atmospheric Radiation*, San Francisco, July 23-27.

Kneizys, F. X., Shettle, E. P., Gallery, W. O., Chetwynd, J. H., Abreu, L. W., Selby, J. E. A., Clough, S. A., and Fenn, R. W., 1983: Atmospheric Transmittance/Radiance: Computer Code LOWTRAN 6. *AFGL-TR-83-0187*.

Kneizys, F. X., et al., 1988: Users Guide to LOWTRAN 7. *AFGL-TR-88-0177*.

Koraćin, D., and Rogers, D. P., 1989: Numerical Simulations of the Response of the Marine Atmosphere to Ocean Forcing. *J. Atmos. Sci.*, 47, 592-611.

Laszlo, I., Jacobowitz, H., and Gruber, A., 1988: The Relative Merits of Narrowband Channels for Estimating Broadband Albedos. *J. Atmos. and Ocean. Tech.*, 5, 757-773.

Masuda, K., Takashima, T., and Rao, C. R. N., 1988: Remote Sensing of Atmospheric Aerosols Over the Oceans Using Multispectral Radiances Measured with the AVHRR. In *Aerosols and Climate*, edited by P. V. Hobbs and M. P. McCormick, Deepak Publishing.

McClatchey, R. A., Fenn, R. W., Selby, J. E. A., Volz, F. E., and Garing, J. S., 1972: Optical Properties of the Atmosphere. *AFCRL-72-0497*.

Mergenthaler, J., 1981: *Some Middle Ultraviolet Radiative Effects of Particulate Pollution in the Stratosphere*. Ph.D. Dissertation, University of Arizona, Tucson.

Mergenthaler, J., 1988: Private Communication, Institute of Atmospheric Physics, University of Arizona, Tucson.

Nakajima, T., King, M. D., and Spinhirne, J. D., 1991: Determination of the Optical Thickness and Effective Particle Radius of Clouds from Reflected Solar Radiation Measurements. Part II: Marine Stratocumulus Observations. *J. Atmos. Sci.*, 48, 728-750.

Neckel, H., and Labs, D., 1984: Solar Radiation Between 3000 and 12500 Å. *Solar Physics*, 90, 205-258.

Nianzeng, Che; Grant, B. G., Flittner, D. E., Slater, P. N., Biggar, S. F., 1991: Results of Calibrations of the NOAA-11 AVHRR Made by Reference to Calibrated SPOT Imagery at White Sands, N.M. *SPIE Proceedings*, V1943-21, Orlando, Florida.

NOAA Polar Orbiter Data User's Guide 1988: edited by K. Kidwell, NOAA/NESDIS.

Noonkester, V. R., 1984: Droplet Spectra Observed in Marine Stratus Cloud Layers. *J. Atmos. Sci.*, 41, 829-845.

Paltridge, G. W., 1974: Infrared Emissivity, Short-Wave Albedo, and the Microphysics of Stratiform Water Clouds. *J. Geophys. Res.*, 79, 4053-4058.

Pilewskie, P., and Twomey, S., 1987: Discrimination of Ice from Water in Clouds by Optical Remote Sensing. *Atmos. Res.*, 21, 113-122.

Planet, W. G., 1988: Data Extraction and Calibration of TIROS-N/NOAA Radiometers. *NOAA Technical Memorandum NESS 107-Rev. 1*.

Plass, G. N., Kattawar, G. W., and Guinn, J. A., 1975: Radiative Transfer in the Earth's Atmosphere and Ocean: Influence of Ocean Waves. *Appl. Opt.*, 14, 1924-1936.

Pollack, J. B., et al. 1974: Aircraft Observations of Venus' Near-Infrared Reflection Spectrum: Implications for Cloud Composition. *Icarus*, 23, 8-26.

Pollack, J. B., Stecker, D. W., Witterhorn, F. C., Erikson, E. F., and Baldwin, B. J., 1978: Properties of the Clouds of Venus as Inferred from Airborne Observations of its Near-Infrared Reflectivity Spectrum. *Icarus*, 34, 28-45.

Preisendorfer, R. W., and Mobley, C. D., 1986: Albedos and Glitter Patterns of a Wind-Roughened Sea Surface. *SPIE*, 637, Ocean Optics VIII.

Pruppacher, H. R., and Klett, J. D., 1980: *Microphysics of Clouds and Precipitation*. Reidel Publishing Co., Holland.

- Radke, L. F., Coakley, J. A., and King, M. D., 1989: Direct and Remote Sensing Observations of the Effects of Ships on Clouds. *Science*, 246, 1146-1149.
- Rao, C. R. N., 1987: Pre-Launch Calibration of Channels 1 and 2 of the Advanced Very High Resolution Radiometer. *NOAA Technical Report NESDIS 36*.
- Rao, C. J. N., Stowe, L. L., and McClain, E. P., 1989: Remote Sensing of Aerosols over the Oceans Using AVHRR Data- Theory, Practice and Applications. *Int. J. Remote Sensing*, 10, 743-749.
- Rawlins, F., and Foot, J. S., 1990: Remotely Sensed Measurements of Stratocumulus Properties during FIRE Using the C130 Aircraft Multi-channel Radiometer. *J. Atmos. Sci.*, 47, 2488-2503.
- Riehl, H., 1979: *Climate and Weather in the Tropics*. Academic Press, New York.
- Roach, W. T., Brown, R., Caughey, S. J., Garland, J. A., and Readings, C. J., 1976: The Physics of Radiation Fog: I- A Field Study. *Quart. J. R. Met. Soc.*, 102, 313-333.
- Robinson, I. S., 1985: *Satellite Oceanography, An Introduction for Oceanographers and Remote-Sensing Scientists*. Ellis Horwood Limited, Chichester, 1985.
- Rothman, L. S., et al., 1987: The HITRAN Database: 1986 Edition. *Appl. Opt.*, 26, 4058-4097.
- Rozenberg, G. G., Malkevich, M. S., Malkolva, V. S., and Syachinov, V. I., 1974: Determination of the Optical Characteristics of Clouds from Measurements of Reflected Solar Radiation on the Kosmos 320 Satellite. *Izv. Atmos. Ocean. Phys.* 10, 14-24.
- Sato, M., Kawabata, K., and Hansen, J., 1977: A Fast Invariant Imbedding Method for Scattering Calculations and an Application to Equivalent Widths of CO₂ Lines on Venus. *The Astrophysical Journal*, 216, 947-962.
- Saunders, R. W., and Edwards, D. P., 1989: Atmospheric Transmittances for the AVHRR channels. *Appl. Opt.*, 28, 4154-4160.
- Schnapf, A., editor, 1985: *Monitoring Earth's Ocean, Land, and Atmosphere from Space- Sensors, Systems, and Applications*. American Institute of Aeronautics and Astronautics, Inc., New York.
- Selby, J. E. A., and McClatchey, R. A., 1972: Atmospheric Transmittance from 0.25 to 28.5 μm : Computer Code LOWTRAN 2. *AFCRL 72-0745*, AD 763721.

Selby, J. E. A., Kneizy, F. X., Chetwynd, J. H., and McClatchey, R. A., 1978: Atmospheric Transmittance/Radiance: Computer Code LOWTRAN 4, *AFGL TR-078-0053*, ADA 058643.

Shine, K. P., Henderson-Sellers, A., and Slingo, A., 1984: The Influence of the Spectral Response of Satellite Sensors on Estimates of Broadband Albedo. *Quart. J. R. Met. Soc.*, 110, 1170-1179.

Slater, P. N., Biggar, S. F., Holm, R. G., Jackson, R. D., Mao, Y., Moran, M. S., Palmer, J. M., and Yuan, B., 1987: Reflectance- and Radiance-Based Methods for the In-Flight Absolute Calibration of Multispectral Sensors. *Remote Sens. Environ.*, 22, 11-37.

Slingo, A., Nicholls, S., Schmetz, J., 1982: Aircraft Observations of Marine Stratocumulus During JASIN. *Quart. J. R. Met. Soc.*, 833-856.

Squires, P., 1966: An Estimate of the Anthropogenic Production of Cloud Nuclei. *J. Rech. Atmo.*, 2, 297-308.

Stephens, G. L., and Tsay, S., 1990: On the Cloud Absorption Anomaly. *Q. J. R. Meteorol. Soc.*, 116, 671-704.

Stephens, G. L., and Platt, C. M. R., 1987: Aircraft Observations of the Radiative and Microphysical Properties of Stratocumulus and Cumulus Cloud Fields. *J. Climatol. Appl. Meteorol.*, 26, 1243-1269.

Somerville, R. C. J., and Remer, L. A., 1984: Cloud Optical Thickness Feedbacks in the CO₂ Climate Problem. *J. Geophys. Res.*, 89, D6, 9668-9672.

Taylor, J., 1991: Private Communication. Meteorological Research Flight, Farnborough, Hampshire, GU14 6TD, England.

Teillet, P. M., Slater, P. N., Ding, Y., Santer, R. P., Jackson, R. D., and Moran, M. S., 1990: Three Methods for the Absolute Calibration of the NOAA AVHRR Sensors In-Flight. *Remote Sens. Environ.*, 31, 105-120.

Telford, J. W., and Chai, S. K., 1984: Inversions, and Fog, Stratus and Cumulus Formation in Warm Air over Cooler Water. *Boundary-Layer Met.*, 29, 109-137.

Thekaekara, M. P., Kruger, R., and Duncan, C. H. 1969: Solar Irradiance Measurements from a Research Aircraft. *Appl. Opt.*, 8, 1713-1732.

- Thekaekara, M. P., 1974: Extraterrestrial Solar Spectrum, 3000-6100 Å at 1 Å Intervals. *Appl. Opt.*, 13, 518-522.
- Tsay, S., Jayaweera, K., 1984: Physical Characteristics of Arctic Stratus Clouds. *J. Climate and Appl. Met.*, 23, 584-596.
- Twomey, S., 1959: The Nuclei of Natural Cloud Formation. *Geofis. Pura Appl.*, 43, 227-240, 243-249.
- Twomey, S., Jacobowitz, H., and Howell, H. B., 1966: Matrix Methods for Multiple-Scattering Problems, *J. Atmos. Sci.*, 23, 101-108.
- Twomey, S., and Warner, J., 1967: Comparison of Measurements of Cloud Droplets and Cloud Nuclei. *J. Atmos. Sci.*, 24, 702-703.
- Twomey, S., and Wojciechowski, 1969: Observations of the Geographical Variation of Cloud Nuclei. *J. Atmos. Sci.*, 26, 684-688.
- Twomey, S., 1974: Pollution and the Planetary Albedo. *Atmos. Environ.*, 8, 1251-1256.
- Twomey, S., 1975: Explicit Calculation of Layer Absorption in Radiative Transfer Computations by Doubling Methods. *J. Quant. Spectrosc. Radiat. Transfer*, 15, 774-778.
- Twomey, S., 1977a: *Atmospheric Aerosols*. Elsevier Scientific Publishing Co., Amsterdam, 302 pp.
- Twomey, S., 1977b: The Influence of Pollution on the Shortwave Albedo of Clouds. *J. Atmos. Sci.*, 34, 1149-1152.
- Twomey, S., 1979: Doubling and Superposition Methods in the Presence of Thermal Emission. *J. Quant. Spectrosc. Radiat. Transfer*, 22, 355-363.
- Twomey, S., and Bohren, C. F., 1980: Simple Approximations for Calculations of Absorption in Clouds. *J. Atmos. Sci.*, 37, 2086-2094.
- Twomey, S., Cocks, 1982: Spectral Reflectance of Clouds in the Near-Infrared: Comparison of Measurements and Calculations. *J. Met. Society of Japan*, 60, 583-592.
- Twomey, S., Piepgrass, M., and Wolfe, T., 1984: An Assessment of the Impact of Pollution on Global Cloud Albedo. *Tellus*, 36B, 356-366.

- Twomey, S., Gall, R., Leuthold, M., 1987: Pollution and Cloud Reflectance. *Boundary Layer Met.*, 41, 335-348.
- Twomey, S., 1989: Aerosols, Cloud Physics and Radiation. Plenary Lecture 3, *American Association for Aerosol Research*, Annual Meeting, Reno NV.
- Twomey, S., Cocks, T., 1989: Remote Sensing of Cloud Parameters from Spectral Reflectance in the Near-Infrared. *Beitr. Phys. Atmosph.*, August 1989, 172-179.
- Valley, S. L., editor, 1965: *Handbook of Geophysics and Space Environments*. Air Force Cambridge Research Laboratories.
- Van de Hulst, H. C., 1963: Remarks on Multiple Scattering. *ICES Electromagnetic Scattering*, Proceedings of the Interdisciplinary Conference held at Clarkson College of Technology, Kerker, M., editor, Pergamon Press.
- Van de Hulst, H. C., 1964: Diffuse Reflection and Transmission by a Very Thick Plane-Parallel Atmosphere with Isotropic Scattering. *Icarus*, 3, 336-344.
- Van de Hulst, H. C., 1980: *Multiple Light Scattering*. Academic Press, New York, 739 pp.
- Walton, C., 1987: The AVHRR/HIRS Operational Method for Satellite Based Sea Surface Temperature Determination. *NOAA Technical Report NESDIS 28*.
- Warner, J., and Twomey, S., 1967: The Production of Cloud Nuclei by Cane Fires and the Effect on Cloud Droplet Concentration. *J. Atmos. Sci.*, 24, 704-706.
- Warner, J., 1969: The Microstructure of Cumulus Cloud. Part I. General Features of the Droplet Spectrum. *J. Atmos. Sci.*, 26, 1049-1059.
- World Climate Research Programme 1986: A Preliminary Cloudless Standard Atmosphere for Radiation Computation. *WCP-112, WMO/TD-24*.
- Whittacker, E. T., and Watson, G. N., 1962: *A Course in Modern Analysis*. Cambridge University Press.

Whitlock, C. H., et al. 1988: *AVHRR and VISSR Satellite Instrument Calibration Results for both Cirrus and Marine Stratocumulus IFO Periods*. FIRE Science Team Meeting, Vail, CO., July 11-15.

Wydick, J., Davis, A., and Gruber, A., 1987: Estimation of Broadband Planetary Albedo from Operational Narrowband Satellite Measurements. *NOAA Technical Report NESDIS 27*.