Infrared Propagation Modeling beneath Marine Stratus Clouds

H. G. HUGHES
Science and Technology Corporation, Hampton, Virginia

C. R. ZEISSE
Space and Naval Warfare Systems Center, San Diego, California

(Manuscript received 19 October 1998, in final form 8 June 1999)

ABSTRACT

Airborne measurements of aerosol size distributions are used to determine the vertical profiles of infrared (IR) extinction and absorption coefficients and asymmetry factors in eight different maritime stratus cloud regimes during unstable boundary layer conditions where the sea temperature was greater than the ambient air temperature. The average values of these parameters are determined relative to the level where the air temperature change with elevation was near a moist adiabatic lapse rate. A model to determine the effects of aerosols on IR propagation beneath these types of clouds is presented in terms of multiplying arrays compatible with the input format of the transmittance/radiance computer code MODTRAN. The model is used in a modified version of MODTRAN to test its utility in system performance predictions beneath these types of clouds. The maximum detection range (MDR) of a surface ship by an airborne forward looking infrared (FLIR) system was determined to be a factor of 3 in better agreement with the observed MDR than that determined using the MODTRAN ICLD3 stratus model with the Navy Aerosol Model (NAM) beneath the cloud. The predictions were found to be insensitive to the wave slope model used in the zero-range sea radiance calculations. This is shown to be the result of a compensating effect between increased sea emissions and decreased cloud reflections for the larger wave slope variances associated with unstable boundary layer conditions as compared to those for stable or neutral conditions.

1. Introduction

Of particular importance to fleet operations is the prediction of the performance range of airborne Forward Looking Infrared (FLIR) passive surveillance systems over long slant paths at sea beneath stratus clouds where the vertical gradients in humidity and aerosol structure can be large. Under these full-sky coverage conditions, the sea radiance backgrounds are influenced by the cloud that emits infrared (IR) radiation differently than does the clear sky. The computer codes LOWTRAN 7 (Kneizys et al. 1988) and its successor MODTRAN 2 (Berk et al. 1989) include various structured stratus cloud models, which provide a means for determining the atmospheric transmission and radiance beneath these types of clouds. The default models are for an altostratus cloud (base 2.4 km, top 3.0 km), a stratus cloud (base 0.33 km, top 1.0 km), a stratus/stratocumulus cloud (base 0.66 km, top 2.0 km), and a nimbostratus cloud (base 0.16 km, top 0.66 km). The models are based on the Deirmendjian (1960) modified gamma aerosol size distribution given by

\[ n(r) = A r^{\alpha} \exp(-B r^g), \]

where \( n(r) \) is the number of particles per (cm\(^3\) \(\mu m\)). The model is described by an array of wavelength-dependent extinction and absorption coefficients and asymmetry factors that are precalculated from Mie theory using the appropriate constants of the size distribution models for each type of cloud. For the stratus model (ICLD3), the constants \( A, B, \alpha, \) and \( g \) are equal to 27, 0.6, 2.0, and 1.0, respectively. The calculated extinction coefficient at 0.55 \(\mu m\) is 55.18 km\(^{-1}\). The total number of particles per cubic centimeter is 250, and the modal radius is 3.33 \(\mu m\). The wavelength-arrayed extinction and absorption coefficients are normalized by the extinction coefficient at 0.55 \(\mu m\). To account for the variation of the IR parameters with altitude, the dependence of the 0.55-\(\mu m\) extinction coefficient with altitude must be known. Below the cloud, one is presently relegated to using one of the various aerosol models available within the MODTRAN code to specify the 0.55-\(\mu m\) extinction coefficient. These steady-state models are based on vertical profiles of relative humidity to account for aerosol growth, and the stability of the boundary layer is not specified.
2. Model development

Development of a dynamic model of the cloud-subcloud turbulence effects on aerosol propagation parameters would be impractical for operational uses. In this paper, a steady-state model is presented to describe the effects of aerosols on IR propagation beneath the cloud. The model is based on easily obtainable parameters, such as cloud height and thickness, and is intended to provide improved estimates of atmospheric transmission/radiance in stratus regimes than are presently available within the MODTRAN code.

The MODTRAN code is not configured to accept inputs of user-defined extinction and absorption coefficients and asymmetry factors for the wavelength band of interest at each of the 34 atmospheric levels allowed in the code. It does allow user-defined parameters to be specified within four altitude regions. This is accomplished in the same manner as discussed in the previous section using wavelength-dependent multiplying arrays for each of the four regions. In each array the extinction and absorption coefficients (normalized to the 0.55-μm extinction coefficient) and the asymmetry factors are required for the multiple scattering mode. The scaling of the arrays with altitude is accomplished with the 0.55-μm extinction coefficient for each of the 34 altitudes.

This configuration is used here to develop a stratus cloud aerosol IR propagation model. For this purpose, a dataset of aerosol size distributions measured in eight different maritime stratus regimes off the coast of Southern California in 1981 is used. This dataset consists of airborne (twin-engine Piper Navajo) measurements of the droplet spectra \(N(r)\), [particles per (cm\(^3\) μm\(^{-1}\))], made as a function of elevation \(Z\), along 2-min horizontal runs (6.44 km) at different levels in the eight stratus cloud layers. The measurement area was located approximately 130 km southwest of San Diego, California. The sampling period at each elevation was 5 and 8 s for \(Z\) and \(N(r)\), respectively. Each 8-s spectra, representing 429 m along a horizontal flight, was obtained utilizing an ASSP-100 (0.23 μm ≤ \(r\) ≤ 14.7 μm) and OAP-200 (14.2 μm ≤ \(r\) ≤ 150 μm) particle spectrometers (manufactured by Particle Measuring Systems, Inc.). These spectrometers determine the particle sizes within specified radii intervals by laser beam (0.6328 μm) scatter, assuming the particle refractive index to be that of pure water. Thus, 24 measurements of elevation and 15 droplet spectra were obtained along each 2-min run. The ambient air temperature (HP quartz thermometer, model 2801A) and dewpoint sensor (EG&G, model 137C3) were measured along each horizontal run. An IR radiometer (Barnes PRT-5) was on board the aircraft for measuring the sea surface temperature from low-altitude horizontal flights. Surface wind speeds during the measurements were not available.

Noonkester (1985) first used this dataset to calculate extinction coefficients profiles at three selected laser wavelengths (0.53, 1.06, and 10.6 μm) relative to the elevation above and below cloud base. The cloud base was defined as the altitude where the liquid water content reached 0.02 gm m\(^{-3}\). These reference levels, as well as the absorption coefficients and asymmetry factors, and other wavelengths required to complete the MODTRAN multiply arrays, were not presented. Determination of the cloud-base height from the liquid water content is not practical for shipboard operation. Here, we define a reference level, \(Z_{r}\), as that elevation where the temperature change with elevation was near a moist adiabatic lapse rate. This elevation may not relate to the same cloud base that would be determined by rotating beam ceilometers or laser ceilometers. It is, however, a level that is commonly identifiable from the aircraft measurements on all of the measurement days. The top of the cloud, \(Z_{t}\), was determined to be that level where the temperature increased with elevation. The cloud reference levels and tops for the dataset are shown in Table 1, along with the air and the sea surface temperatures obtained during low-altitude (30–40 m) horizontal flights. The sea surface temperatures were not measured on 14 May.

The average size distribution along the horizontal path at each elevation was calculated and presented in a data report by Noonkester (1982) for each regime. For this study, the \(N(r)\) plots were digitized to provided a text file of the size distribution versus radius for use in a Mie computer code for calculating the IR parameters over the wavelength band of interest. As the plots were logarithmic, a constant digitization rate produced different radii intervals within and between decades. The digitized points for size distributions measured below the reference levels on 29 May and 17 August are shown in Fig. 1. The altitudes above the sea listed for each regime correspond to 48 m (29 May) and 38 m (17 August) below the reference level. For radii less than about 20 μm, the August data indicate a higher total number of particles than that for the May data. Examples of the cloud size distributions for the same days are shown in Fig. 2. The altitudes above the sea listed in the figure correspond to 181 m (29 May) and 197 m (17 August) above the reference level. The major differences between the cloud size distributions and the subcloud distributions occur below about 1-μm radius, where the \(N(r)\)’s have been depleted by about an order

**Table 1. Cloud reference levels, cloud tops, and air and sea temperatures.**

<table>
<thead>
<tr>
<th>Date</th>
<th>(z_r) (m)</th>
<th>(Z_t) (m)</th>
<th>(T_\text{air}) (°C)</th>
<th>(T_\text{sea}) (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>14 May</td>
<td>620</td>
<td>1090</td>
<td>14.7</td>
<td>—</td>
</tr>
<tr>
<td>28 May</td>
<td>369</td>
<td>701</td>
<td>15.1</td>
<td>17.6</td>
</tr>
<tr>
<td>29 May</td>
<td>395</td>
<td>684</td>
<td>14.6</td>
<td>18.3</td>
</tr>
<tr>
<td>11 Aug</td>
<td>352</td>
<td>814</td>
<td>17.8</td>
<td>22.3</td>
</tr>
<tr>
<td>13 Aug</td>
<td>745</td>
<td>876</td>
<td>19.0</td>
<td>19.7</td>
</tr>
<tr>
<td>14 Aug</td>
<td>460</td>
<td>679</td>
<td>19.5</td>
<td>20.2</td>
</tr>
<tr>
<td>17 Aug</td>
<td>396</td>
<td>660</td>
<td>17.5</td>
<td>19.1</td>
</tr>
<tr>
<td>18 Aug</td>
<td>488</td>
<td>673</td>
<td>17.7</td>
<td>20.6</td>
</tr>
</tbody>
</table>
Fig. 1. Examples of substratus cloud particle size distributions measured on 29 May and 17 Aug 1981.

Fig. 2. Examples of stratus cloud particle size distributions measured on 29 May and 17 Aug 1981 compared with the Deirmendjian cloud model.

Fig. 3. The 0.55-μm extinction coefficient variation with distance from the cloud reference level.

of magnitude, and near 10-μm radius, where the $N(r)$'s have been increased by about an order of magnitude. Noonkester (1985) concluded that the May and August data represent different air masses, that is, maritime and continental air masses, respectively, which may account for some of the differences. In any event, the total dataset is not sufficiently large to justify differentiating between the two types of air masses.

The MODTRAN ICLD3 stratus model is also shown in Fig. 2 for comparison. The measured spectra do not approximate the ICLD3 model in the submicron range. The disagreement is more noticeable in the May spectrum where a modal peak occurs near 7 μm.

As the aerosols were sized by the PMS spectrometers assuming their refractive indices to be that of pure water, there is no justification for using a different refractive index when performing Mie calculations with the measured size distributions. The range of relative humidity encountered on all horizontal flights for the measurement days was between 80% and 100%. With the expressions of Hanel (1971) for the refractive indices of a mixture, and the particle growth factor employed in the Navy Aerosol Model (NAM; Gathman and Davidson 1993), we can estimate the uncertainties to be expected if the particles' refractive indices were other than pure water. If the particles were sea salt solutes, the uncertainties in calculated extinction coefficients within the cloud are less than 2% at wavelengths of 0.55 and 10 μm. Beneath the cloud, the uncertainties for both wavelengths are less than 6%. These uncertainties are small considering that the standard deviations of the average size distributions in the dataset are greater than ±50% in many instances.

The database provided 102 averaged spectra (57 below and 45 above the reference levels). Each spectrum was then utilized in the Mie code to calculate the 0.55-μm extinction coefficient and the infrared extinction and absorption coefficient and asymmetry factors for 13 individual wavelengths corresponding to the response of a typical mercury cadmium telluride (HgCdTe) detector from 6.5 to 14.8 μm. The individual spectra were normalized to the elevation below or above the reference level $Z^*$ (in km$^{-1}$) given by,

$$Z^* = Z - Z_R.$$

The results of the calculations at 0.55 μm are shown in Fig. 3. The solid line represents the least squares regression analysis of the data points to the equation

$$Z^* = -71.86 \sigma^{-0.744} + 0.0422 \sigma^{2.09},$$

where $\sigma$ is the 0.55-μm extinction coefficient (km$^{-1}$). From the plot, a subjective decision was made to divide

$\lambda = 0.55$ microns

$\lambda = 10$ microns

$\lambda = 50$ microns
the stratus regime into four separate regions; two subcloud regions \([-500 \text{ m} \leq A < -100 \text{ m}\) and \((-100 \text{ m} \leq B < 0 \text{ m})\)] and two cloud regions \([0 \text{ m} \leq C < 100 \text{ m}\) and \((100 \text{ m} \leq D \leq 500 \text{ m})\)].

Equation (3) is readily solved for the extinction coefficient at \(Z^* = 0\), which is 13.73 \(\text{km}^{-1}\). For the subcloud region \((-500 \text{ m} \leq Z^* \leq -11 \text{ m}\)), the extinction coefficient can be approximated closely by the relationship

\[
\sigma = (-0.0139Z^*)^{-1.34},
\]

and for the cloud region \((11 \text{ m} \leq Z^* \leq 500 \text{ m})\),

\[
\sigma = (23.59Z^*)^{0.477}.
\]

Using Eqs. (4) and (5), an example of the normalized extinction and absorption coefficients at 10.0 \(\mu\text{m}\) is shown in Figs. 4 and 5, respectively, versus the distance from the reference level. The accompanying calculated asymmetry factors are shown in Fig. 6. In these examples, the solid regression lines represent polynomial fits to the data and are shown only to accentuate the general trend of the points with altitude. The average normalized extinction and absorption coefficients and asymmetry factors for each of the four regions were calculated for the wavelength band of interest and are shown in Figs. 7, 8, and 9, respectively. An example of the average normalized extinction coefficients at 10 \(\mu\text{m}\) and their standard deviations for each region are displayed in Fig. 10. The width of the dotted rectangles for each region span the plus and minus standard deviations about the averages depicted by the arrows. The large standard deviations (approximately \(\pm 50\%\) of the mean values) in regions A and B indicate the high variability of the subcloud size distributions. Also shown in Figs. 7, 8, and 9, are the parameters for the ICLD3 cloud model. At a wavelength of 10 \(\mu\text{m}\), the ICLD3...
model normalized extinctions and absorption coefficients and asymmetry parameters are greater than region D of the new model (hereafter referred to as the SSC SD model) by 21.2%, 16.5%, and 3.5%, respectively. These lower normalized extinction and absorption values for the SSC SD model are the result of the larger 0.55-μm extinction coefficients caused by the larger number of submicron particles in the measured distributions compared to the Deirmendjian model.

3. Surface layer stability considerations

The formation, maintenance, and decay of stratus clouds are generally well understood and are adequately discussed by Paluch and Lenschow (1991), and more recently by Tjernstrom and Rodgers (1996). In the present observations, where the sea surface temperatures were warmer than the air, moisture near the surface is transported upward from the surface toward the top of the mixed layer, where condensation takes place leading to the formation of the stratus cloud. During daytime, the short-wave solar heating within the cloud and long-wave cooling at the cloud top tend to destabilizes the entire cloud. If the turbulent flux from the surface is not sufficiently strong, the heating of the lower parts of cloud may generate a secondary temperature inversion close to the cloud base. The cloud will then be decoupled from the surface mixed layer and severed from the transport of moisture from the surface. The temporal and spatial averaging process over long horizontal flights complicates the observation of these processes. However, we see evidence of this decoupling in the temperature and relative humidity profiles of the present dataset. When the air is warmer than the sea, the cloud is not decoupled, and entrained air above the cloud can be transported to the surface layer. The turbulence profiles in these stable conditions will produce different aerosol profiles than do unstable conditions. Whether or not the modeled parameters developed here can be applied in stable or neutral conditions has yet to be determined. Depending on the intensity of the vertical thermal buoyancy forces at the surface, they can either dominate or reduce the mechanical forces influencing the formation of capillary waves on the sea surface and alter its emissivity (Hwang and Shemdin 1988; Shaw and Churnside 1997). Hwang and Shemdin used a refractive laser slope gauge to demonstrate the effects of stability and swell on wave slope measurements. More recently, Shaw and Churnside (1997) measured laser glint patterns at sea from a floating instrumentation platform (FLIP). They derived the along-wind wave slope probability density functions that agreed well with the Cox and Munk (1954) model under stable or near-neutral conditions. However, during unstable conditions (sea warmer than the air), they measured a rougher surface than predicted by the along-wind Cox–Munk model. The along-wind wave slope variance, \( \sigma^2_{\text{sc}} \), dependence on wind speed at 10-m elevation, \( U_{10} \) (m s\(^{-1}\)), derived from the linear regression of the Shaw–Churnside tabulated data is given by

\[
\sigma^2_{\text{sc}} = 2.0 U_{10}^2 + 0.005
\]
The regression line variances exceed the along-wind variances of the Cox–Munk distributions by a factor of 1.8. Shaw and Churnside did not present measurements of cross-wind variances. However, it can be shown (using calculations to be discussed in the next section) that sea background radiances calculated with the Cox–Munk variances are not strongly dependent upon the angular differences between the wind direction and the look angle for zenith angles less than about 92°. Within this limit, the equivalent blackbody temperature of the zero-range sea radiances calculated with variances for along-wind and cross-wind directions differ by less than 0.2°C.

4. Systems performance predictions

During July and August 1992, tests were conducted off the coast of northern California to assess the accuracy of computer models to predict the performance ranges of airborne forward looking infrared (FLIR) systems operating against surface targets. This was a joint effort of the Naval Ocean Systems Center (the predecessor to the Space and Naval Warfare Systems Center, San Diego Division), with the Naval Postgraduate School (NPS), Monterey, California, using the research vessel (R/V) Point Sur. The R/V Point Sur was equipped with a full meteorological suite, radiosonde launches (Vaisala model RS80), and a thermistor system (16 sensors) for measuring the hull temperatures. Stratus cloud conditions existed on seven days of the experiment. An FLIR-equipped navy operational aircraft flew over the ship at specified altitudes during the local midmorning and nighttime hours, and the FLIR sensor operator recorded the maximum detection range (MDR) for each flight. Depending upon the altitude of the aircraft relative to cloud base, the MDR varied between 5 and 9 km on two of the measurement days (30 and 31 July).

An observation made on 30 July, approximately 1-h prior to a radiosonde launch from the R/V Point Sur, is chosen for comparison of the measured and predicted MDRs. For this observation, the sea surface temperature was 16.6°C, the air temperature (14.9°C) was measured from the radiosonde launch altitude of 7 m, and the wind speed (4.8 m s⁻¹) is referenced to a height of 10 m. The average ship temperature (starboard side) at the time of observation was 13.1°C. The sensor’s line of sight relative to the wind direction was 16°. From the radiosonde record, the temperature lapse rate changed from a near-dry to a near-moist adiabatic at 182 m and the cloud top was determined to be at 590 m. The air-mass factor (ICSTL) is a required input for NAM. It is related to the atmospheric radon content expressed in pCi m⁻³ and determines the contribution of continental aerosols to the first component of the size distribution. It ranges between integer values of 1 for the open ocean and 10 for coastal regions. Radon measurements on board the R/V Point Sur determined this parameter to be an integer value of 8. These observations provide the input parameters to MODTRAN for calculating the extinction coefficients as a function of altitude calculated using the ICLD3 model with NAM. In Fig. 11, the extinctions calculated for a wavelength of 0.55 μm are compared with those determined for the new model (SSC SD model) using Eqs. (3) and (4). Near the reference level (182 m), the SSC SD model extinction coefficients are nearly an order of magnitude greater than those for NAM. Within the cloud, the SSC SD extinction coefficients are greater by a factor of 2. The 10-μm extinction coefficients calculated for the two models are then shown in Fig. 12. The SSC SD model extinction coefficients near the reference level are greater than those for NAM by nearly two orders of magnitude. There is an abrupt transition in the data points between regions A and B. Because of the separation between the averaged values for adjacent regions, a
smooth transition of the parameters from one region to another cannot be expected.

To test the models’ utility in predicting FLIR performance, a simple detection task is chosen. For this purpose, the spatial frequency ($v$) (cycles mrad$^{-1}$) of a ship in terms of its angular subtense ($\Theta$) at a range $R$ is defined as

$$v = (2\Theta)^{-1} = R(2000D_c)^{-1},$$

where $D_c$ is the ship’s critical dimension in meters. Here, $D_c$ is related to the ship’s projected area ($A_p$) as viewed by the sensor, that is, $D_c = (A_p)^{1/3}$. The projected area of the ship is determined assuming it to be represented by an equivalent parallelepiped given (for broadside viewing) by

$$A_p = lh \cos \theta + lw \sin \theta,$$

where $h$ is the unobscured ship height, $w$ is the width, $l$ is the length, and $\theta$ is the viewing elevation angle. The range at which the atmosphere transmittance degrades the zero-range ship-background temperature difference down to the sensor’s minimum detectable temperature difference (MDTD) is the maximum detectable range, that is,

$$\text{MDTD} = (T_{\text{SHIP}} - T_{\text{BKG}}) \tau_{\text{MDR}},$$

where $(T_{\text{SHIP}})$ is the average ship temperature, $T_{\text{BKG}}$ is the effective blackbody temperature of the sea at zero range, and $\tau_{\text{MDR}}$ is the atmospheric transmittance at the maximum range of detection. The sensor MDTD($v$) is known such that MDTD($R$) for a particular ship configuration can be determined. The zero-range sea background temperature and the atmospheric transmittance as a function of range are determined using the measured meteorological as inputs to the computer code SEARAD (Zeisse et al. 1999), a modification to MODTRAN. The SEARAD code determines the contributions from the sea surface emissions and the reflected cloud radiances to the band-averaged radiance, leaving a wave facet at zero range and directed toward a sensor.) A Gaussian distribution of wave tilts defines the Fresnel reflection angles at the surface and the zenith angle from where the cloud radiance is reflected into the sensor. In this instance, the sea was warmer than the air, and the Shaw–Churnside wave slope variance model for unstable conditions [Eq. (6)] is applicable. The results of the MDR calculations are shown graphically in Fig. 13. The contrast temperature degradation curve (solid line) was calculated using the Shaw–Churnside variances. Zero-range sea background temperatures were calculated as a function of the viewing zenith angle as the aircraft flew toward the ship at an altitude of 152 m. The temperature difference between the ship and the zero-range sea background was then degraded by the atmospheric transmittance calculated along the line of sight. The point at which this curve intersects the sensor’s MDTD curve (solid line) determines the MDR of the ship. For comparison, the contrast temperature degradation curve (dashed line) was calculated using the Cox–Munk variances for stable or near-neutral conditions. In this case, the angular difference between the sensor look angle and the wind direction was 16°. The MDR observed (marked by the FLIR operator) was 5.5 km. The calculated MDR’s were nearly the same (~8.3 km, as indicated by the arrow) for both variance models. As can be determined from Table 2, this is a result of a compensating effect between wave slope emissivity and reflectivity. In a like manner, the MDR’s calculated using the ICLD3 cloud model with NAM beneath the cloud were nearly the same (~15.3 km) for both variance models. Table 2 was calculated with a sensor altitude of 152 m at a range of 5.86 km, and the zenith angle
of the line of sight was 91.5°. The increase in wave slope variance of the Shaw–Churnside model over that of the Cox–Munk model increases in the average emissivity of the sea surface by 11.8%, thereby increasing sea radiance by the same amount. Then, by Kirchhoff’s law, the increase in emissivity lowers the reflected cloud radiance by 19.4%. The net result is an increase of 0.6% in the total zero-range sea radiance, which corresponds to only a 0.4°C increase in equivalent blackbody temperature. The difference in the two contrast temperature difference curves at ranges less than about 8 km is caused by the absence of a cross-wind component in the Shaw–Churnside variances.

5. Conclusions

This study has demonstrated the utility of the SSC SC model to improve the performance predictions of FLIR systems operating beneath these types of clouds during unstable boundary layer conditions. While the model overestimates the observed MDR by about 3 km in the case presented, its difference from the observed value is an improvement by a factor of 3 over that calculated using the ICLD3 model with NAM beneath the cloud. This result does not distract from the accuracy of the ICLD3 model in this situation. The key role of the cloud in the infrared scene is in its contribution to the zero-range sea radiance. For a thickness of 300 m or greater, these types of clouds are essentially black at long IR wavelengths (Stephens et al. 1978). The discrepancy in the calculated MDRs is caused by the higher transmittances (≈50% at 5.86 km) of the NAM model below the cloud.

Whether the SSC SD model is applicable during stable or neutral conditions has not been determined. However, the results of this analysis have also shown that the modeled MDRs beneath these types of clouds are insensitive to whether the variance model employed in the calculations was for stable or unstable conditions.

Acknowledgments. This work was accomplished under the Space and Naval Warfare Systems Center, San Diego (SSC SD), Contract 66001-94-D-0064, Delivery Order 0009. Funding for the project was provided by the Office of Naval Research Exploratory Development Program.

REFERENCES