Downward Longwave Irradiance at the Ocean Surface From Satellite Data: Methodology and in Situ Validation

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A methodology is presented for estimating downward longwave irradiance at the ocean surface from satellite radiance data. The downward longwave irradiance is computed with a fast and accurate radiative transfer model as a function of temperature, water vapor, ozone and carbon dioxide mixing ratios, fractional cloud coverage, emissivity of clouds, and cloud top and cloud base altitudes. A sensitivity study is performed to assess the relative importance of the model input parameters and devise strategies regarding their retrieval. Ozone and carbon dioxide mixing ratios are consequently fixed at their climatological values, whereas the other parameters, highly variable in space and time, are determined from satellite data. Temperature and water vapor mixing ratio are obtained from NOAA Tiros operational vertical sounder data, and cloud parameters are retrieved from GOES visible and infrared spin scan radiometer data. Several methods are investigated to retrieve the cloud parameters. In the most refined method, cloud base altitude is deduced from cloud top altitude and liquid water path, assuming a vertical liquid water distribution within the clouds. In the other methods, simplifying assumptions are introduced, which include directly relating liquid water path to cloud geometrical thickness, fixing the cloud geometrical thickness to its climatological value, and, finally, parameterizing the cloud effects only as a function of fractional cloud coverage. Satellite-derived irradiances are compared to those measured in situ during the Mixed Layer Dynamic Experiment, conducted in October–November 1983 off the central California coast. The results indicate that the satellite methods perform similarly, with standard errors of estimate ranging from 21 to 27 W m$^{-2}$ on a half-hourly time scale and from 16 to 22 W m$^{-2}$ on a daily time scale. These errors correspond to 6 to 8% and 4 to 6% of the average measured values, respectively.

1. INTRODUCTION

The net radiative flux at the ocean surface, i.e., the net solar radiation absorbed minus the net longwave radiation emitted, is one component of the air-sea heat balance, the other two being the latent and sensible heat fluxes. As a major heat source for the oceans, the net radiative flux constitutes an important boundary forcing for the general ocean circulation and is a crucial parameter for determining meridional oceanic heat transport. Together with the top-of-atmosphere radiative flux, it also provides estimates of tropospheric radiative heating and cloud radiative forcing. Knowing its space and time variability over the world’s oceans is therefore central to questions of climate. On smaller scales, an adequate knowledge of the net radiative flux is necessary to estimate the upper ocean heat content and will contribute to an improved characterization of upper ocean properties.

Direct high-quality radiation measurements at sea are difficult to make and therefore are made only from research vessels. This is particularly the case for the net longwave radiation flux, which requires careful instrument calibration and temperature corrections. Because of the lack of long-term measurements, investigators interested in long-time scale studies [e.g., Budyko, 1963; Wyrhtki, 1965; Esbensen and Kushnir, 1981] have had to rely upon empirical formulas that involve quantities regularly observed from ships (sea surface temperature, air temperature and specific humidity near the surface, fractional cloud coverage) to evaluate the climatological means over large ocean areas. Even though ship observations are extensive along shipping lanes, vast geographical gaps still exist, especially in the southern oceans. Moreover, empirical formulas have been validated only over limited regions but are often applied globally, leading to large and inhomogeneous errors. It has become clear during the last few years that applying the available empirical formulas is not sufficient to achieve the 10 W m$^{-2}$ accuracy required for climatological applications [e.g., Fung et al., 1984].

If operational satellite data could be used for estimating the net radiative flux at the ocean surface, they would remedy some of the empirical formula deficiencies. They would provide, in particular, global spatial coverage and adequate temporal sampling. Attempts have been made to relate the net radiative flux at the surface to the top-of-atmosphere fluxes [Pinker and Corio, 1984]. The problem, however, is complex because no strong correlation generally exists between top-of-atmosphere and surface flux patterns [Ramanathan, 1986].

Another approach is to estimate separately the different components of the net radiative flux. Various methods have been developed for estimating the shortwave component (the largest component). The downward part, or insolation, was derived first [e.g., Tarpley, 1979; Gautier et al., 1980]; the albedo [e.g., Briegleb et al., 1986] and the net solar flux [Gautier, 1984] were obtained later. Insolation estimates are accurate to about 10% on daily time scales, and better on longer...
time scales [Gautier and Katsaros, 1984; Gautier, 1986]. Estimating the net longwave component is still problematic. The upward part can be easily determined from the sea surface temperature, since the ocean surface emits almost as a blackbody. The downward part, however, is more difficult to estimate, since it depends on atmospheric moisture, temperature, and cloud properties. Up to now, a few methods have been developed and can be classified as either statistical or physical. The statistical methods utilize combinations of multispectral infrared radiances [e.g., Smith and Woolf, 1983; Morcrette and Deschamps, 1986], while the physical ones [e.g., Darnell et al., 1983] are based on modeling the radiative processes that occur in the atmosphere. In Schmetz et al.'s [1986] hybrid approach, thermodynamic grid point fields at 850 and 1000 mbar are used together with satellite-derived cloud parameters. Only the methods of Darnell et al. [1983] and Schmetz et al. [1986], in which the clouds are simply parameterized, have actually been tested with in situ measurements, and this was done only over land.

In this paper, our approach to derive downward longwave irradiance at the ocean surface is physical. We utilize a radiative transfer model and determine its most important input parameters from satellite radiance data. Section 2 is devoted to describing the radiative transfer model, a simplified version of Morcrette et al.'s [1986] routine for global circulation models. Section 3 presents the results of a sensitivity study, performed to assess the effects and relative importance of the model input parameters and to devise strategies for their retrieval. In view of the sensitivity study and our knowledge of atmospheric variability, section 4 introduces four methods by which to retrieve model input parameters. The methods differ essentially in the way the cloud parameters are obtained, with satellite observations being employed whenever necessary and possible. Section 5 describes both the in situ and the satellite data used to validate the above retrieval methods. These data were collected during the Mixed Layer Dynamics Experiment (MILDEX), which took place off the central California coast from October 24 to November 13, 1983. In section 6 the satellite-derived irradiances, obtained using the four retrieval methods, are compared with the in situ measurements. Also in section 6 the performance of the satellite methods is compared with that of empirical formulas that employ conventional surface data. The statistical methods utilize combinations of multispectral infrared radiances, while the physical ones [e.g., Darnell et al., 1986] are based on modeling the radiative processes that occur in the atmosphere. In Schmetz et al.'s [1986] hybrid approach, thermodynamic grid point fields at 850 and 1000 mbar are used together with satellite-derived cloud parameters. Only the methods of Darnell et al. [1983] and Schmetz et al. [1986], in which the clouds are simply parameterized, have actually been tested with in situ measurements, and this was done only over land.

2. Radiative Transfer Model

Many techniques exist for solving the radiative transfer equation in clear and cloudy atmospheres [Lenoble, 1977]. Given our objective of an operational algorithm for estimating downward longwave irradiance at the ocean surface from satellite data, the radiative transfer model must be computationally efficient and involve parameters that can be derived from space. These constraints keep us away from the more sophisticated schemes that, in particular, explicitly account for scattering within the clouds. Our model is therefore based on a highly parameterized scheme developed for global atmospheric circulation models [Morcrette et al., 1986].

2.1. Clear Atmosphere

The downward longwave irradiance at the surface, \( F^-(0) \), is calculated from the radiative transfer equation developed into

\[
F^-(0) = 2\pi \int_0^\infty dv \left\{ \int_0^1 \mu \, d\mu \left[ -B_s(z) \mu + B_s(0) + \int_0^Z \frac{dB_s}{dT}(z) \mu \, d\mu \right] \right\}
\]

(1)

where \( v \) is the frequency, \( \mu \) is the cosine of the zenith angle, \( B_s \) is the Planck function, \( t_\mu(z, 0; \mu) \) is the atmospheric transmittance between altitude \( z \) and the surface, \( T \) is the temperature, and \( Z \) is the top-of-atmosphere altitude. The first and second terms on the right-hand side represent the contribution of the upper and lower limits of the atmosphere, respectively, and the last term accounts for radiative processes occurring between these limits.

The integration over \( \mu \) is performed by evaluating \( t_\mu \) in a direction defined by \( \mu = 1/\gamma \), where \( \gamma \) is the radiative diffusivity factor, that is,

\[
\int_0^1 d\mu \, t_\mu(z, 0; \mu) \approx \frac{1}{\gamma} t_\mu(z, 0; 1) \quad (2)
\]

This approximation, introduced by Elsasser [1942], is common in radiative transfer calculations, and its validity has been demonstrated by many authors [e.g., Rodgers and Walthall, 1966; Chou and Arking, 1980]. Although \( \gamma \) may vary between 1 (strong line approximation) and 2 (weak line approximation), we fixed \( \gamma \) at 1.66, the value used in most atmospheric radiation codes. Higher \( \gamma \) values would be more appropriate for calculations at higher levels in the atmosphere. Combining (1) and (2) yields the following expression for \( F^-(0) \):

\[
F^-(0) \approx \int_0^\infty dv \left\{ -\pi B_s(Z, 0; 1) + \pi B_s(0) + \int_0^Z \frac{dB_s}{dT}(z) \mu \, d\mu \right\}
\]

(3)

To evaluate the integral over \( z \) on the right-hand side of (3), the atmosphere is divided into \( N \) layers of arbitrary thickness (Figure 1a). Temperature and absorber amounts are specified at each level separating these layers. In spectral regions where atmospheric absorption is strong (almost everywhere outside the 8–14 \( \mu \)m range), \( t_\mu \) varies strongly with respect to \( z \) near the surface because the radiative energy is exchanged over short distances. Integration over the layer adjacent to the surface must therefore be performed carefully, as was demonstrated by Wu [1980]. In our scheme, a two-point Gaussian quadrature [Carnahan et al., 1969] is used for the layer adjacent to the surface. For the other layers, a simple trapezoidal rule is applied. Thus we have

\[
\int_0^z \frac{dB_s}{dT}(z) \mu \, d\mu \approx \frac{1}{2} \sum_{k=2}^N \left( \int_0^z \frac{dB_s}{dT}(z_{k+1/2}) \frac{dT}{dz}(z_{k+1/2}) \right) \left[ t_\mu(z_{k+1/2}, 0; 1) \right]
\]

(4)

where the summation over \( k \) represents the contribution of
Fig. 1a. Contribution to $F^+(0)$ of the different layers of the model atmosphere for (a) clear and (b) cloudy atmospheres.
distant layers and the summation over \( I \) represents the contribution of the layer adjacent to the surface. In (4), subscript \( k + \frac{1}{2} \) refers to the middle of layer \( k \), \( \zeta_i^* \) is the ordinate of the \( i \)th Gauss angle, and \( w_i \) is the corresponding weight. The vertical integration scheme therefore accounts for the effect of an eventual temperature inversion in the lowest layer. When compared with the more accurate 32-point Gaussian quadrature for all layers, the 2-point Gaussian quadrature for the layer adjacent to the surface gives a maximum 4 W m\(^{-2}\) error in \( F^-(0) \) [Morcrette and Fouquart, 1985].

To perform the integration over \( \nu \), we adopted Rodgers' [1967] emissivity approach. Four spectral intervals are considered (see Table 1), in which atmospheric absorption is due mainly to water vapor, carbon dioxide, and ozone. Absorption by minor gaseous constituents (e.g., methane and nitrous oxide) and aerosols is neglected. The contribution of each spectral interval to \( F^-(0) \) is evaluated using normalized transmissivity functions.

Two normalized transmissivity functions characterize each type of absorption \( j \) in a given spectral interval \( i \). They are defined by

\[
t_i(z, 0) = \int_{\nu_{1i}}^{\nu_{2i}} dvB(z, \nu_{1i}, 0; 1/\gamma)
\]

\[
t_{ij}(z, 0) = \int_{\nu_{1i}}^{\nu_{2i}} dv(dB_i/dT)(z, \nu_{1i}, 0; 1/\gamma)
\]

where \( \nu_{1i} \) and \( \nu_{2i} \) are the lower and upper frequency limits of spectral interval \( i \), respectively. The first one, \( t_i \), is used to calculate the energy exchange with the top of the atmosphere (first term on the right-hand side of (3)); the second one, \( t_{ij} \), is used to calculate the energy exchange with the other layers (last term on the right-hand side of (3)). The functions \( t_i \) and \( t_{ij} \), indeed, depend on the absorber amount along the optical path.

To evaluate \( t_i \) and \( t_{ij} \), \( t_j \) is modeled according to Morcrette [1984]. Absorption by water vapor is treated with the statistical band model of Goody [1952], whereas absorption by carbon dioxide and ozone is modeled according to Malek [1967]. At high altitudes, these models are modified to account for the Voigt profile of the absorption lines. The strong line approximation is assumed for water vapor and carbon dioxide absorption, while the weak line approximation is supposed to hold for ozone absorption. To account for temperature and pressure dependence on absorption, the Curtis-Godson approximation [e.g., Goody, 1964] is used, and equivalent absorber amounts are defined accordingly (for more details, see Morcrette et al. [1986]). Since we are dealing with moist boundary layers, absorption by water vapor continuum must be treated properly. This is accomplished according to Clough et al. [1980]. Utilization of Clough et al.'s [1980] material is further warranted by the good agreement obtained with results from narrow-band and line-by-line models [Morcrette et al., 1986].

Morcrette and Fouquart [1985] have shown that the strong and weak line approximations overestimate gaseous absorption, introducing significant (several watts per square meter) errors in \( F^-(0) \). These deficiencies are remedied following Garand [1983]: corrections to the absorber amounts are applied in those spectral regions where the absorption is overestimated. The corrections are obtained from regressions between transmittances calculated with, on the one hand, the original statistical band models and, on the other hand, the approximate models.

The downward longwave irradiance at the surface is finally expressed numerically as

\[
F^-(0) = \sum_{i=1}^{N} \left\{ -\pi B_i(Z) \sum_{j} t_{ij}(Z, 0; 0) + \pi B_i(0) \right. \\
+ \frac{\pi}{2} \sum_{k=1}^{N} \frac{dB_i}{dT}(z_{k+1/2}) \frac{dT}{dz}(z_{k+1/2}) \\
\cdot \prod_{j} (t_{ij}(z_{k}, 0) + t_{ij}(z_{k+1}, 0)) \left( z_{k+1} - z_k \right) \\
+ \pi \sum_{i=1}^{2} \sum_{k=1}^{N} w_i \frac{dB_i}{dT}(z_{k}^*) \frac{dT}{dz}(z_{k}^*) \prod_{j} t_{ij}(z_{k}^*, 0) \right\}
\]

(7)

where \( B_i \) is the black body function integrated over spectral interval \( i \). Figure 1a shows how the different terms of (7) are vertically distributed; that is, to which layer or level of the model atmosphere they correspond.

### 2.2. Cloudy Atmosphere

In our model, clouds are assumed to occur in one layer, and plane parallel theory is used to parameterize their effect on the downward longwave irradiance. This simple approach is justified because from space it is not easy to distinguish overlying clouds and to retrieve cloud geometrical properties. Even though plane parallel theory must be regarded as suspect in the presence of broken clouds [e.g., Harshvardhan and Weinman, 1982], it ought to be employed when cloud geometry is not available. This is in fact the approach used in most global circulation models. For consistency, cloud scattering is also neglected. To include cloud scattering would have implied specifying the microphysical characteristics of the cloud particles (size distribution, complex refractive index, and phase function), and these parameters are even less accessible from satellite measurements. In any case, tests performed by Morcrette [1978] have shown that cloud scattering effects are small (≈2-3 W m\(^{-2}\)), even at small optical thicknesses. In these conditions, cloud effects are described with parameters that can be derived from space with some accuracy, namely, fractional coverage and emissivity, and the downward longwave irradiance is a linear combination of clear and cloudy irradiances.

In practice, the scheme first ignores the clouds and computes the corresponding flux, \( F_{o_1}^-(0) \), as described in the previous section. The flux that would be observed if the cloud fractional cover and emissivity were equal to 1 is then evalu-
obtained from the following combination of fluxes calculated during winter, but is more moderate ($< 100$ W m$^{-2}$). The downward flux for the actual cloudy atmosphere is finally obtained from the following combination of fluxes calculated in the previous steps:

$$F_{c-}^{(0)} = A_c F_{c-}^{(0)} + (1 - A_c) F_{c-}^{(0)}$$  

(9)

where $A_c$ is the cloud amount (product of fractional coverage and emissivity) in layer $n$. Figure 1b gives the contribution of the different model layers to $F_{c-}^{(0)}$ when (9) is expressed numerically. Above the cloud base, the clear sky terms are now multiplied by ($1 - A_c$).

### 2.3. Accuracy and Computational Efficiency

The longwave radiative transfer model has been compared with a more detailed, broadband model [Morcrette and Fouquart, 1985] and with Scott and Chedin's [1981] line-by-line model. Comparisons have also been performed with broadband pyrgeometer measurements over land by Morcrette and Deschamps [1986]. In all these comparisons, the agreement is within 2-3% ($4-7$ W m$^{-2}$). While maintaining a good level of accuracy, the model keeps the computational burden as small as possible (e.g., 3 ms on Cray 1 for a 15-layer clear atmosphere) and therefore is very suitable for monitoring $F_{c-}^{(0)}$ over large extents of ocean.

### 3. Sensitivity of $F_{c-}^{(0)}$ to Model Input Parameters

We examine in this section the sensitivity of $F_{c-}^{(0)}$ to temperature, water vapor and ozone mass mixing ratio profiles, carbon dioxide volume mixing ratio, cloud base altitude, and cloud amount. The effect of cloud top altitude will not be considered, since once cloud base altitude is fixed, the only important cloud parameter is cloud amount. The downward flux $F_{c-}^{(0)}$ is computed for five environments: summer and winter conditions in the mid-latitudes and subarctic and mean annual conditions in the tropics. Temperature, water vapor, and ozone mass mixing ratio profiles are taken from McClatchey [1971], whereas carbon dioxide volume mixing ratio is fixed at 340 ppm, the 1982 globally averaged value [Komhyr et al., 1985]. Because radiosonde observations are mainly land based, McClatchey's [1971] profiles may not be representative of oceanic conditions, especially in the lower troposphere from which most of the longwave radiation contributing to $F_{c-}^{(0)}$ originates. Nevertheless, calculations with the above model atmospheres are useful to provide information about the sensitivity of $F_{c-}^{(0)}$ to changes or uncertainties in the atmospheric column. This information is necessary for assessing the importance of parameters, their required accuracy, and strategies for their retrieval.

Table 2 gives the clear sky $F_{c-}^{(0)}$ values. The downward longwave irradiance is maximum in the tropics and minimum in the winter subarctic. There is a strong meridional gradient during winter: $F_{c-}^{(0)}$ increases by more than 200 W m$^{-2}$ from subarctic to tropical regions. A meridional gradient subsists during winter, but is more moderate ($< 100$ W m$^{-2}$).

### Table 2. Downward Longwave Irradiance at the Surface Under Clear Skies for Different Model Atmospheres

<table>
<thead>
<tr>
<th>Model Atmosphere</th>
<th>Longwave Irradiance, W m$^{-2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tropical</td>
<td>390.6</td>
</tr>
<tr>
<td>Mid-latitude summer</td>
<td>343.0</td>
</tr>
<tr>
<td>Mid-latitude winter</td>
<td>233.4</td>
</tr>
<tr>
<td>Subarctic summer</td>
<td>296.5</td>
</tr>
<tr>
<td>Subarctic winter</td>
<td>171.6</td>
</tr>
</tbody>
</table>

For each of the five model atmospheres, variations in the mean conditions were introduced, ranging from $-5$ to $5$ K for temperature, and from $-50$ to $50\%$ for both water vapor and ozone mass mixing ratios and carbon dioxide volume mixing ratio. The results are presented in Figure 2. When the atmospheric column is warmer (Figure 2a) or more humid (Figure 2b), $F_{c-}^{(0)}$ increases. The reverse is true when the conditions are colder or drier, except that the water vapor effect, as opposed to the temperature effect, is not as linear. The sensitivity of $F_{c-}^{(0)}$ to temperature and water vapor mixing ratio is higher in the tropics than in the subarctic. For instance, a 5 K warmer atmosphere increases $F_{c-}^{(0)}$ by about 20 W m$^{-2}$ in the tropics and 10 W m$^{-2}$ in the winter subarctic. In the case of a 50% more humid atmosphere, the values become 40 and 35 W m$^{-2}$, respectively. Variations in the ozone and carbon dioxide mixing ratio (Figures 2c and 2d) have a small impact on $F_{c-}^{(0)}$. A 50% change in mixing ratio modifies $F_{c-}^{(0)}$ by 1 W m$^{-2}$ or less. The seasonal and latitudinal variability of carbon dioxide concentration, however, is only a few percent of the globally averaged value [Komhyr et al., 1985] and, therefore, negligibly affects $F_{c-}^{(0)}$.

We now investigate the sensitivity of $F_{c-}^{(0)}$ to a hypothetical cloud of given downward emissivity $e_c$, occupying one layer of the model atmosphere from altitudes $z_b$ (base) and $z_t$ (top), and covering a fraction $C$ of the sky. Figure 3 shows the effect of cloud amount $A_c$ (product of $e_c$ and $C$) for clouds with bases at 1 and 5 km. As expected from (9), $F_{c-}^{(0)}$ increases linearly with $A_c$, the slope depending on the type of overlying atmosphere and on $z_b$. The sensitivity of $F_{c-}^{(0)}$ to $A_c$ is minimum in the tropics and maximum in the winter subarctic and mid-latitudes. These last two environments respond very similarly to changes in $A_c$. The effect of a cloud with $A_c = 1$ and $z_b = 1$ km is about 50 W m$^{-2}$ in the tropics and 80 W m$^{-2}$ in the winter subarctic (Figure 3a), reducing to 35 and 45 W m$^{-2}$ when $z_b$ increases to 5 km (Figure 3b). The effect of $z_b$ is more clearly illustrated in Figure 4. As $z_b$ increases, $F_{c-}^{(0)}$ generally decreases, except for $z_b < 1$ km in the winter subarctic because of a temperature inversion in the lower troposphere. Changes in $z_b$ affect $F_{c-}^{(0)}$ similarly for the five model atmospheres. For instance, the difference between cloud effects in the tropics and winter mid-latitudes is 25 W m$^{-2}$ when $A_c = 1$ and $z_b = 0.5$ km, decreasing to 20 W m$^{-2}$ when $z_b$ increases to 5 km (Figure 4b). The sensitivity of $F_{c-}^{(0)}$ to $z_b$ increases noticeably with $A_c$. It is about 1.3 W m$^{-2}$/km when $A_c = 0.5$ (Figure 4a) and becomes 3.5 W m$^{-2}$/km when $A_c = 1$ (Figure 4b).

Simulations of the cloud effect on $F_{c-}^{(0)}$ were also performed with typical clouds. We used Stephens' [1978a] model, which describes most liquid water clouds. The results are given in Table 3. The average effect is minimum in the tropics ($\sim$50 W m$^{-2}$) and maximum in the winter mid-latitudes ($\sim$70 W m$^{-2}$). Except for the tropical atmosphere, the values are
similar for all environments. Compared with the clear sky $F^{-}(0)$ values, the cloud effect is more important at high and middle latitudes. The correction represents, for instance, 39% of the clear sky value in the winter subarctic but only 12% in the tropics. Table 3 also indicates that the effect of various clouds may differ by 15 W m$^{-2}$ (1σ) from the average values. This suggests that the same treatment for all clouds might be appropriate in the tropics where the clear sky flux is generally large but that distinction between clouds should be made in other regions, especially during winter.

4. DETERMINATION OF THE MODEL INPUT PARAMETERS

Our primary objective is to estimate $F^{-}(0)$ for oceanographic studies with a 10 W m$^{-2}$ accuracy on daily and 50- to 100-km spatial scales. This objective, in view of the sensitivity study presented above and our knowledge of atmospheric
variability, determines the scales at which the various model input parameters need to be specified. For carbon dioxide and ozone mixing ratios, it is clearly sufficient to use climatological data. For the other parameters, however, which are highly variable in space and time and have the strongest influence on $F^{-}(0)$, frequent monitoring at scales smaller than 50 km is required.

Only satellites can provide adequate sampling at the required spatial and temporal scales. Existing operational satellites, however, only make observations in regions of the electromagnetic spectrum (visible and infrared) that are largely opaque to clouds. These observations therefore do not give direct access to some of the parameters to which $F^{-}(0)$ was previously shown to be sensitive (e.g., cloud base altitude, emissivity of clouds, temperature, and water vapor profiles below clouds), and consequently one must investigate indirect methods to retrieve these parameters.

Since cloudy regions behave differently than clear air regions with regard to radiative transfer, studied areas are partitioned accordingly. Advantage is taken of the highest spatial resolution in the satellite data to estimate cloud coverage and most of the relevant cloud properties. Because of their high repetitiveness, geostationary satellite measurements are used to monitor natural cloud variability. This reduces the uncertainty and potential bias on time-averaged estimates, which would result from too infrequent sampling (e.g., when using polar orbiter data). In the presence of clouds, the properties of the atmosphere above, within, and below the clouds are estimated by interpolating data from nearby clear sky regions.

Several methods are investigated to retrieve the most variable and important input parameters. In all these methods, temperature and water vapor mixing ratio are obtained from infrared sounder data. Depending on the method, however, some of the cloud parameters are extracted differently. In the most refined method (hereinafter referred to as method A), our governing assumptions are that cloud reflectance is strongly related to liquid water path [Stephens and Webster, 1984] and that liquid water concentration within the clouds has a reasonably well known law of variation with temperature [Feigelson, 1978; Jeck, 1983]. These assumptions allow one to deduce cloud emissivity from cloud reflectance as well as cloud base altitude from cloud top altitude, parameters that can be inferred with reasonable accuracy from satellite data. In the other methods, a hierarchy of simplifying assumptions is introduced. These include directly relating liquid water path to cloud geometrical thickness (method B), assuming clouds to be blackbodies and specifying cloud geometrical thickness from climatological data (method C), and finally, parameterizing cloud effects only as a function of fractional coverage (method D). Table 4 gives a schematic of the four basic retrieval methods considered.

### 4.1. Most Refined Method

Air temperature and water vapor mixing ratio can be obtained from Tiros operational vertical sounder (TOVS) and visible and infrared spin scan radiometer (VISSR) Atmospheric Sounder (VAS) data using inversion techniques developed by, among others, Smith and Woof [1976], Chedin and Scott [1984], and Susskind et al. [1984] for TOVS and by Smith [1983] for VAS. VAS data are more appropriate to monitor atmospheric variability because of the instrument's high sampling frequency (one observation every half hour). Unfortunately, VAS is not continuously operated in sounding mode and was not operated at all during the experimental period considered in our later analysis. We therefore use TOVS data, in particular, the operational sounding products made available by the National Oceanic and Atmospheric Administration (NOAA). TOVS, however, only provides soundings twice a day. The retrievals are limited to clear or partly cloudy conditions, and space and time interpolation is necessary for local estimations in overcast conditions. We do not attempt to account for the particular shape of the temperature and water vapor mixing ratio within the clouds. Retrievals' accuracy is typically 2-3 K for temperature and 30% for water vapor mixing ratio. According to Figure 2, this may introduce 20-60

<table>
<thead>
<tr>
<th>Cloud Effects, W m$^{-2}$</th>
<th>Model Atmosphere</th>
<th>Average</th>
<th>Standard Deviation</th>
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<tbody>
<tr>
<td>Tropical</td>
<td>48.7</td>
<td>±13.4</td>
<td></td>
</tr>
<tr>
<td>Mid-latitude summer</td>
<td>60.7</td>
<td>±15.4</td>
<td></td>
</tr>
<tr>
<td>Mid-latitude winter</td>
<td>68.2</td>
<td>±15.6</td>
<td></td>
</tr>
<tr>
<td>Subarctic winter</td>
<td>64.3</td>
<td>±16.7</td>
<td></td>
</tr>
<tr>
<td>Subarctic winter</td>
<td>66.4</td>
<td>±13.2</td>
<td></td>
</tr>
</tbody>
</table>
where fl is the backscattered fraction of solar radiation. The nonabsorbing medium, rc is \cite{Coakley and Chylek, 1975} significantly above 0.8 /m \cite[e.g., Welch et al., 1980]{}. Knowing R c in the 0.5- to 0.75-ttm range, while clouds only absorb significantly because the VISSR visible channel includes wavelengths or quasi-transparent clouds \cite[e.g., certain types of cirrus]{}. Absorption coefficients for ozone of direct and diffuse reflection coefficients, respectively, and a and a’ are absorption coefficients for ozone of direct and diffuse solar flux, respectively. Since clouds, even very small ones, greatly increase the energy to the satellite, I p is used as a threshold to detect the presence of clouds. The threshold allows one to derive the number of cloudy pixels within subsets of images \cite{elementary areas over which F -0(0) is calculated} and therefore the fractional cloud coverage. The expected error in cloud coverage is 10%, except in the presence of transparent or quasi-transparent clouds \cite[e.g., certain types of cirrus]{}. According to Figure 3, such an error affects the accuracy of F -0(0) computations by 5-10 W m \^-2 for a cloud base at 1 km.

The broadband downward emissivity of clouds, e c-, is also derived from VISSR data in the visible. A several-step procedure is employed that involves Stephens’ \cite{1978b} theoretical results. The first step is to compute cloud reflectance, R c. For each pixel classified as cloudy, R c is obtained from the calibrated energy flux to the satellite, I c, by solving the following quadratic equation \cite{Diak and Gautier, 1983}:

\[
I_c = S_0 \mu_0 \left[1 - \alpha(1 - \alpha')(1 - \alpha)^2 + (1 - \alpha)(1 - \alpha')R_c \right] (11)
\]

where S_0 is the top-of-atmosphere solar irradiance in the VISSR visible channel, R c is the surface reflectance \((R_c \approx 0.06)\), \mu_0 is the cosine of the solar zenith angle, and a and a’ are absorption coefficients for ozone of direct and diffuse solar flux, respectively. Since clouds, even very small ones, greatly increase the energy to the satellite, I c is used as a threshold to detect the presence of clouds. The threshold allows one to determine the number of cloudy pixels within subsets of images \cite{elementary areas over which F -0(0) is calculated} and therefore the fractional cloud coverage. The expected error in cloud coverage is 10%, except in the presence of transparent or quasi-transparent clouds \cite[e.g., certain types of cirrus]{}. According to Figure 3, such an error affects the accuracy of F -0(0) computations by 5-10 W m \^-2 for a cloud base at 1 km.

The broadband downward emissivity of clouds, e c-, is also derived from VISSR data in the visible. A several-step procedure is employed that involves Stephens’ \cite{1978b} theoretical results. The first step is to compute cloud reflectance, R c. For each pixel classified as cloudy, R c is obtained from the calibrated energy flux to the satellite, I c, by solving the following quadratic equation \cite{Diak and Gautier, 1983}:

\[
I_c = S_0 \mu_0 \left[1 - \alpha(1 - \alpha')(1 - \alpha)^2 + (1 - \alpha)(1 - \alpha')R_c \right] (11)
\]

In this expression, cloud absorption is neglected. This is justified because the VISSR visible channel includes wavelengths in the 0.5- to 0.75-\mu m range, while clouds only absorb significantly above 0.8 \mu m \cite[e.g., Welch et al., 1980]{}. Knowing R c then allows us to compute cloud optical thickness \(\tau_c\). For a nonabsorbing medium, \(\tau_c\) is \cite{Coakley and Chylek, 1975}:

\[
\tau_c = \mu_0 \frac{R_c}{\beta (1 - R_c)} (12)
\]

where \(\beta\) is the backscattered fraction of solar radiation. The values of \(\beta\) are taken from Stephens et al.’s \cite{1984} tables, obtained from the two-stream model of Coakley and Chylek \cite{1975}. The calculations, however, were performed between 0.3 and 0.75 \mu m, which is not exactly the spectral interval of the VISSR visible channel, but cloud reflectance below 0.5 \mu m is fairly constant, to within a few percent for most clouds \cite[e.g., Welch et al., 1980]{}. The next step is to compute liquid water path \(W\) from \(\tau_c\). We use Stephens’ \cite{1978b} formula, established from theoretical calculations with standard cloud models. In the case of conservative scattering, the formula reads as follows:

\[
\log_{10} W = \exp \left[ \log_{10} (\tau_c) - 0.2633 \right] / 1.7095 (13)
\]

Finally, \(e_c\) is deduced from \(W\), still following Stephens \cite{1978b}:

\[
e_c = 1 - \exp (-0.158W) (14)
\]

The error in estimating \(e_c\) using (11) through (14) is difficult to assess, since on the one hand, Stephens’ \cite{1978b} calculations were performed for typical clouds and, on the other hand, the uncertainty on \(\tau_c\) is not readily known. Assuming, however, a reasonable 15% uncertainty on \(\tau_c\), which includes model and calibration errors, the error induced on \(e_c\) remains less than 5% for \(\tau_c > 0.2\) and most solar zenith angles. For optically thin clouds \((R_c < 0.2)\), larger errors are expected because of the exponential form of (13).

To estimate cloud base altitude, cloud top altitude together with information about cloud geometrical thickness is necessary. Our approach is as follows. For each elementary area, we consider the average radiances measured in the VISSR infrared channel, \(L^i\). Assuming nonreflecting clouds, \(L^i\) can be written as

\[
L^i = L_s^+ + e_c^+ C(L_c^+ - L_s^+ + L_s^+ - L_c^+) (15)
\]

where \(L_s^+\) is the clear sky radiances, \(L_c^+\) is the upwelling radiances from the clouds, and \(e_c^+\) is the upward emissivity of the clouds in the 10- to 12-\mu m region. In order to solve (15) for \(L_c^+, e_c^+\) is obtained from \(W\) and Stephens’ \cite{1978b} theoretical formula:

\[
e_c^+ = 1 - \exp (-0.13W) (16)
\]

Fractional cloud coverage \(C\) is determined as indicated above, and \(L_s^+\) is computed from TOVS temperature and water vapor mixing ratio retrievals by modeling the radiative transfer in the atmosphere. The \(L_s^+\) value is then used to derive
cloud top temperature and, given the atmospheric temperature profile, cloud top altitude. We assume that the clear air temperature profile inferred from satellite data is the same in the neighboring cloudy region. Atmospheric absorption above the clouds is also accounted for in the calculation. Note that once the pixels are determined to be clear or cloudy, \( L_0^+ \) and \( L_0^- \) can be obtained directly from infrared radiance data. This simpler procedure should be adopted whenever possible. In the present study, however, we were obliged to devise the more indirect approach described above, which is also more subject to uncertainty, because of the degraded resolution of the VISSR infrared data (about 8 km at the equator). Given cloud top altitude, cloud base altitude is finally obtained from \( W \), the retrieved temperature profile, assuming a vertical liquid water distribution within the clouds. We use Feigelson's [1978] statistical relationship between liquid water concentration and temperature. This relationship was established from aircraft soundings at altitudes up to 7-8 km over the United States and is shown in Figure 5. Even though the standard deviations around the mean values are large, it is expected that over large enough elementary areas the liquid water variability will be, at least partly, smoothed out. Also reported in Figure 5 are more recent results obtained by Jeck [1983] over the United States with a variety of cloud types and weather conditions up to 3 km. These results are in fairly good agreement with Feigelson's [1978]. Unfortunately, Feigelson's [1978] relationship applies only to liquid water clouds. No alternative is envisaged for icy clouds. Therefore clouds located above 8 km, where supercooling is unlikely to be encountered, will not be considered. Using typical errors for \( \varepsilon_0^+ \), \( C_0 \), and \( L_0^+ \), 10 to 20% errors on \( L_0^- \) are estimated when \( A_0 \) varies from 1 to 0.5. This translates into errors of 1.5 km to 3 km on \( z_w \) which affects \( F^- (0) \) by 5-10 W m\(^{-2}\). As \( A_0 \) decreases further, the error on \( z_w \) increases rapidly, but the effect on \( F^- (0) \) becomes smaller and even large errors in \( z_w \) do not change \( F^- (0) \) much (see Figures 3 and 4).

4.2. Simpler Methods

The method described above has several limitations. First, since visible data are used mostly to derive the cloud parameters, it cannot be applied at night. Second, the procedure to compute cloud base altitude is complex and requires several steps, each adding a potential source of error. Because of the relative uncertainties in the relationships between visible brightness and cloud parameters, it is not obvious that such complexity is necessary for most applications. Therefore simpler methods have been investigated.

Method B takes advantage of the strong correlation existing between the total liquid water content of a cloud, \( W \), and its geometrical thickness \( H \). Paltridge [1974a, b] proposed the following relationship, which is based on knowing the cloud thickness alone:

\[
H = \left( \frac{W}{b} \right)^{1/2}
\]

where \( b = 180 \text{ g m}^{-2} \text{ km}^{-1} \). This formula was developed from aircraft measurements in tropical clouds of low liquid water content and therefore might not apply to all types of clouds. Devault and Katsaros [1983], in fact, showed that during the Joint Air-Sea Interaction (JASIN) experiment the value of \( b \) given by Paltridge [1974a, b] was much too small; but their data were somewhat limited, and further studies are necessary to determine the variability of \( b \). The utilization of (16), however, highly simplifies the computation of \( z_w \) since neither the temperature profile nor the liquid water profile within the clouds needs to be known. For consistency, \( \varepsilon_0^- \) is related to \( W \) by

\[
\varepsilon_0^- = 1 - \exp (0.05W)
\]

where the constant 0.05 has been determined experimentally by Paltridge and Platt [1976]. It is important to note that (18) differs notably from (14), which was derived theoretically by Stephens [1978b]. For instance, the discrepancy in \( \varepsilon_0^- \) is 0.4 when \( W = 10 \text{ g m}^{-2} \). In summary, method B differs from method A only in the determination of \( \varepsilon_0^- \) and \( z_w \) from \( W \) and \( z_w \) which is accomplished in method B directly via (17) and (18).

In method C it is assumed that all clouds are sufficiently thick and dense for them to be optically black; their emissivity is therefore equal to 1. Even though this assumption is unlikely to hold for high clouds, in particular cirrus clouds, it remains statistically a good approximation for middle and low clouds (see, for example, Paltridge and Platt, 1976). For further simplification, cloud thickness is fixed at 0.5 km, which according to various climatologies [e.g., Teleadas and London, 1954; Sasamori et al., 1972] is typical of midtropical clouds. This climatological value, however, is not representative of nimbostratus and cumulonimbus type clouds, which can extend well beyond several kilometers in the vertical. Underestimation of \( F^- (0) \) is therefore expected in the presence of such clouds, for instance by about 10 W m\(^{-2}\) if \( H \) is 2 km too small (see Figure 4). Method C is similar to Darnell et al.'s [1983], which considers clouds as black bodies of 50-mbar pressure thickness. The main difference, however, is that in our case determining the cloud parameters is done with satellite data of much higher space and time resolution. Since method C does not require \( W \), which is derived in methods A and B from satellite data in the visible, it is applicable at night. This is a considerable advantage over methods A and B. For the nighttime case, the cloud cover is determined from satellite data in the infrared.

Method D is based on the theoretical simulations reported in section 3. These simulations suggest that \( F^- (0) \) can be simply parameterized as a function of the clear sky flux \( F_0^- (0) \) and the cloud cover \( C \):

\[
F^- (0) = F_0^- (0) + cC
\]

where \( c \) is an empirical coefficient that depends on latitude, season, and cloud type (see Table 3). Even though \( c \) may vary

![Fig. 5. Statistical relationship between liquid water concentration and temperature within clouds.](image-url)
by 30 W m\(^{-2}\) depending on cloud type, a single average value representing all clouds is assigned for \(c\). In the study case of October–November 1983 at 34\(^\circ\)N that will be discussed later, \(c = 62\) W m\(^{-2}\). In (19), \(F_0(0)\) is computed from temperature and humidity profiles retrieved from TOVS data, and the cloud cover \(C\) is determined from VISSR visible data during the day and from infrared data at night. For global applications, which generally require day and night estimation, only one data set, infrared data, should be utilized to derive \(c\). Paltridge [1970] proposed a similar parameterization, in which \(F(0)\) is also expressed as the sum of a clear sky contribution and of a correction due to the presence of clouds. From observations at Aspendale, Australia, he found that \(c = 60\) W m\(^{-2}\). This value compares favorably with the results of Table 3. In Paltridge’s [1970] approach, however, \(F_0(0)\) is estimated using empirical formulas and surface temperature and humidity measurements made aboard ships. In method D we use only satellite-derived parameters.

5. Data

In order to test the methods described above, we use surface and satellite data collected during MILDEX. The experiment was conducted from October 24 through November 13, 1983, approximately 300 km off the central California coastline. R/P Flip, a large manned spar buoy, served as the central coordinate for the experiment, and R/V Acania traveled in boxlike patterns at distances of a few kilometers around R/P Flip. R/P Flip drifted freely during the experiment, generally towards the north, with positions ranging from 33.43\(^\circ\)N, 126.44\(^\circ\)W at the beginning of the experiment to 34.12\(^\circ\)N, 126.26\(^\circ\)W at the end.

5.1. Surface Measurements

Measurements of hemispheric downward longwave irradiance were made on both R/P Flip and R/V Acania with Eppley pyrgeometers provided by K. B. Katsaros and R. J. Lind of the Department of Atmospheric Sciences, University of Washington, Seattle. These instruments respond to radiation in the 5–50 \(\mu\)m range and their nominal accuracy is 1.5\% (3–6 W m\(^{-2}\)). They were gimbal mounted to remain level under roll and pitch actions and installed at carefully selected locations to minimize contamination by radiation emanating from ship structures. Other relevant near-surface data collected during MILDEX include observations of water temperature, air pressure, temperature, and humidity. Visual observations of fractional cloud coverage were logged hourly during the experiment, and radiosondes provided air temperature and humidity profiles up to 300 mbar. Approximately 40 radiosonde launches were made on R/V Acania (2–3 launches per day).

After the experiment, the pyrgeometer data were calibrated and corrected for radio frequency noise and differences between dome and housing temperatures [Lind and Katsaros, 1987]. The corrected values were then processed into half-hour averages at 15 and 45 min past the UT hour (VISSR acquisition times). A comparison was made of R/P Flip and R/V Acania half-hourly averages of longwave irradiance (Figure 6). The data used were collected during the period from October 26 to November 9, when R/P Flip and R/V Acania were within 2 km of each other. The correlation coefficient and rms difference are respectively 0.96 at the 95\% confidence level and 8.6 W m\(^{-2}\), respectively (Figure 6a). The slope of best linear fit, 0.97, is close to 1, the 3\% difference introducing relative errors of less than 5 W m\(^{-2}\) in the range of observed irradiances. These figures attest to the quality of the surface measurements. Time series of the two data sets (Figure 6b) indicate the conditions under which the two sets of measurements differ most. No obvious pattern can be detected, but the discrepancies are often larger when abrupt changes in longwave irradiance are measured. R/P Flip’s values, however, are generally higher than R/V Acania’s. The bias, approximately 4.6 W m\(^{-2}\), is not negligible and needs to be understood. Unfortunately, the tests performed, namely comparisons with empirical formulas and examination of longwave irradiances in heavy rain conditions, were inconclusive.

Also, exposure errors do not give a systematic bias, since they depend on the highly variable temperature of contaminating surfaces (see the appendix). Another possibility is that the bias originates from the calibration. Intercalibrating the instruments, however, would be unavailing at this time. Yet, assuming that R/P Flip and R/V Acania measurements are subject to statistically uncorrelated and identical errors, an order of magnitude for the uncertainty in half-hour averages of longwave irradiance (including the bias) is given by

\[
(S_e^2 - R_e^2)/2^{1/2} \approx 7\text{ W m}^{-2} \tag{20}
\]

where \(S_e\) and \(R_e\) respectively denote the standard difference between R/P Flip and R/V Acania data sets and their residual differences after best linear fit.

5.2. Satellite Observations

During MILDEX, the VISSR/VAS instrument onboard the GOES 6 satellite was operated in VISSR mode, the normal cloud mapping mode, viewing the Earth in the visible (0.5–0.75 \(\mu\)m) and infrared (10.4–12.1 \(\mu\)m) regions of the electromagnetic spectrum. Images in these spectral bands were acquired at hourly intervals, from 1545 to 2145 UT, and covered a 600 \times 600 km area centered at 34\(^\circ\)N and 126\(^\circ\)W. Each image contained noncalibrated brightness counts coded on respective 6-bit (visible) and 8-bit (infrared) scales. At 34\(^\circ\)N and 126\(^\circ\)W the spatial resolution of the data is about 1.2 km in the visible and 9.6 km in the infrared.

Since the VISSR/VAS Dwell-Sound mode of operation was not activated during MILDEX, sounding data were acquired from the TOVS instruments aboard NOAA 7 and 8. NOAA provided these data as products containing surface temperature and pressure, a 15-layer tropospheric and stratospheric temperature profile, a 3-layer tropospheric water vapor profile, fractional cloud coverage, cloud top pressure, and total ozone content. Only temperature and water vapor profiles are used in the present study. The spatial resolution of TOVS product is approximately 250 km, and typical accuracies are 2.5 K for mean layer temperature and 30\% for layer water vapor [Kidwell, 1981]. During MILDEX, 605 TOVS products were acquired in the area from 30\(^\circ\)–38\(^\circ\)N and 122\(^\circ\)–130\(^\circ\)W. Unfortunately, no data were available between October 29 and November 6.

The VISSR/VAS data, noncalibrated brightness counts, were converted into geophysical units. Visible brightness counts were transformed into radiances using space and a sunlit area on the Earth surface, the White Sands Monument area in New Mexico, as calibration targets [Frouin and Gau-tier, 1987]. The accuracy of the calibration is about 10\%, which translates into \(F(0)\) errors of only a few watts per square meter. Infrared brightness counts were converted into radiometric temperatures using calibration constants provided.
by NOAA. A calibration check was performed by comparing radiometric temperatures observed in clear sky conditions with those calculated from radiosonde data. For the 15 clear sky observations selected, the rms difference is 0.7 ± 0.2 and the bias negligible. Earth location was accomplished in two stages: first, by using the navigation provided by NOAA and, second, by visually correcting for discrepancies between the predicted position of landmarks on the California coastline and their actual location on the images. The accuracy of the procedure is within ± 3 pixels for the visible images.

The TOVS sounding products were spatially interpolated to the MILDEX location. We applied a standard interpolation technique [Akima, 1976]. In the scheme, all TOVS retrievals corresponding to a particular day are used to produce daily-averaged temperature and mixing ratio profiles. Table 5 gives the weighted distance $d$ between TOVS retrievals and MILDEX location for each day. In general, $d < 100$ km, but in some instances $d > 300$ km. Therefore the TOVS retrievals corresponding to the farthest distances might not represent the actual atmospheric conditions at the experimental site.

Note that when some of the mixing ratio information is missing, $d$ differs for temperature and mixing ratio. From the daily averages, MILDEX-average TOVS profiles (averaged over the entire MILDEX period) were also constructed.

The last preprocessing stage prepared the space- and time-averaged results for the radiative transfer code. The atmosphere was divided into 34 layers between the surface and 1 mbar, for which temperature and water vapor mixing ratio were computed. The vertical resolution is 25 mbar below 300 mbar and 50 mbar above, but the highest layer is 99 mbar thick. To interpolate vertically, a procedure similar to Darnell et al.'s [1983] was employed. The temperature and water vapor mixing ratio were assumed to vary exponentially and linearly with pressure, respectively. Denoting the water vapor mixing ratio at pressure level $p$ by $q$, it can be shown [Smith, 1966] that

$$q = (\lambda + 1)q(p/p_0)^u_0/p_0$$

with

$$\lambda + 1 = \ln(u^*/u_0)/\ln(p^*/p_0)$$

where $p_0$ is the surface pressure, $u_0$ is the total precipitable water in the atmosphere, $u^*$ is the precipitable water above the pressure level $p^*$ (an arbitrary atmospheric level), $g$ is the acceleration due to gravity, and $\lambda$ is a parameter depending on atmospheric and water vapor scale heights. The surface pressure was taken from R/P Flip and R/V Acania data. Values of $u_0$ were computed as the sum of precipitable water vapor burdens in the three TOVS layers (surface to 700 mbar, 700–500 mbar, and 500–300 mbar), while those of $u^*$ involved only the upper two TOVS layers, with $p^*$ = 500 mbar. The amount of water vapor above 300 mbar (top of the highest TOVS layer) was considered insignificant. Figure 7 shows the resulting daily-averaged and MILDEX-averaged profiles. When compared with radiosonde observations (Figure 8), the retrieved temperatures and water vapor mixing ratios are respectively higher by 2 to 3 K and 10 to 15% on the average. According to Figure 2, it suggests that $F(0)$ will be overestimated by about 10 W m$^{-2}$. This, however, might not be true on hourly or even daily time scales, because of the high atmospheric variability during MILDEX. Also, a 2 K temperature
overestimation results in clouds located too high in the atmosphere by up to 0.5 km, which decreases $F^{-}(0)$ by several watts per square meter (Figure 4) and therefore contributes to attenuate the discrepancy. Figure 8 further indicates that the atmospheric variability, well reproduced in the temperature retrievals, is not well reproduced in the water vapor retrievals. This might be due to the poor quality of the water vapor retrievals, which, in particular, do not adequately resolve variations in the vertical.

6. COMPARISON OF SATELLITE ESTIMATES WITH IN SITU MEASUREMENTS

6.1. Numerical Procedures

Once the surface and satellite data have been preprocessed as outlined above, the procedure described in section 4 is applied. The satellite-derived longwave irradiances are estimated over areas of approximately 25 x 25 km, centered on the ship locations. The size of the areas is large enough for the

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Fig. 7. TOVS atmospheric profiles interpolated to the ship locations during MILDEX.

Fig. 8. Comparison of averaged TOVS and radiosonde atmospheric profiles during MILDEX. Profiles corresponding to plus and minus 1 standard deviation from the average are also indicated.
cloud cover computations to be statistically significant but still remains as small as possible, since the highest spatial resolution is sought. Also, the satellite measurements provide instantaneous information over small viewing angles, while the pyrgeometer measurements are integrated over time and a 2π solid angle. We note, in this respect, that 25 km is the distance that mid-latitude clouds would typically travel during 30 min, therefore justifying comparisons between satellite estimates averaged over 25 × 25 km and half-hourly averages of measured longwave irradiance.

To evaluate liquid water path, the visible reflectance of individual GOES VISSR pixels is averaged over the 25 × 25 km areas. We therefore assume that the results obtained in this way do not differ too much from those derived by first computing liquid water path for individual pixels and then averaging (the most logical procedure). Such a procedure, however, would have increased the computational burden considerably, and this has to be avoided for the approach to be applicable over large oceanic areas.

To determine cloud top altitude, precalculated functions relating the cloudy radiance \( L_c^+ \) to cloud top altitude are used. These functions are shown in Figure 9 for each TOVS atmosphere (daily averaged and MILDEX averaged). In the calculations the radiative transfer in the atmosphere is simulated according to Kneizys et al. [1980]. Figure 9 indicates that the sensitivity of \( L^+ \) to cloud top altitude is about twice as much above 3 km than near the surface. When \( L^+ < L_c^+(1 - e_c^- C) \), the considered case is eliminated because \( L_c^- \) cannot be negative. Such negative values are sometimes obtained because on the one hand, clear sky radiances computed with the TOVS profiles only represent average atmospheric conditions and on the other hand, the radiative transfer modeling is subject to uncertainty.

To infer cloud base altitude in method A, a simple integration scheme is employed. First, cloud top altitude replaces the closest atmospheric level of the model atmosphere. Vertical integration of liquid water content over the various atmospheric layers is then performed successively until the resulting amount is greater than the liquid water path determined from the visible reflectance data. Cloud base altitude is finally obtained by linear interpolation between the levels delineating the lowest layer. Since Feigelson's relationship is not valid for high (>7–8 km) or icy clouds, cases for which cloud top altitude is greater than 8 km were not treated.

The longwave irradiance is computed with both daily-averaged and MILDEX-averaged TOVS profiles. This is done to assess whether using daily-averaged TOVS profiles or TOVS profiles averaged over a longer period gives better results, which is not obvious, a priori, in view of the retrievals' large spatial resolution and the interpolation scheme. In the following, the prime denotes that MILDEX-averaged TOVS profiles were used. Longwave irradiiances derived at hourly intervals are also time averaged over about 8 consecutive hours for comparisons with situs measurements. The 8-hour averages are hereinafter referred to as “daily averages.” These daily averages, however, are not expected to provide a good estimate of the actual daily-averaged values, since only daytime satellite observations are selected in the present study. Our intention here is only to study time averaging effects.

6.2. Results

Comparisons of satellite-derived longwave irradiances with surface-measured values are presented in Figures 10 through 13 and the statistics are summarized in Table 6. One hundred four pairs of \( F^-(0) \) values were used in the half-hourly comparisons, and 16 pairs were used in the time-averaged comparisons.

We first examine the performance of methods A, B, C, and D (Figures 10a through 13a). The best statistics are obtained with method A, for which the correlation coefficient and rms difference are respectively 0.69 and 22.6 W m⁻² for half-hourly comparisons, and 0.73 and 18.0 W m⁻² for daily comparisons. The rms differences represent respectively 6.4 and 5.1% of the average \( F^-(0) \) value measured during MILDEX. The slopes of best linear fit are 0.54 (half-hourly comparisons) and 0.52 (daily comparisons), indicating that only 54 and 52% of the observed \( F^-(0) \) variability is reproduced. A bias of −6.5 W m⁻² (underestimation) is computed for both half-hourly and daily comparisons. Compared with method A, method B exhibits slightly better correlation coefficients but slightly larger rms differences (2–3 W m⁻²), similar slopes of best fit, and much larger biases (−13 W m⁻²). The results obtained with method C are similar to those of methods A and B, except that the correlation coefficients and slopes of best fit are definitely smaller. For half-hourly comparisons, the correlation coefficient (0.55) and rms difference (26.6 W m⁻²) compare favorably with those reported by Darnell et al. [1983] for a similar method (0.87 and 20 W m⁻², respectively). A significant difference, however, exists between the two correlation coefficients, which might be due to the fact that the data used by Darnell et al. [1983] describe a wider range of \( F^-(0) \) values (180 W m⁻² instead of 100 W m⁻²). Surprisingly, methods C and D, which involve much less physics, are less biased. The bias is reduced to about −4 W m⁻² for method C and is only −1 W m⁻² for method D. Of all methods, however, methods C and D provide the smallest correlation coefficients, the largest rms differences, and slopes of best fit farthest from 1.

When MILDEX averaged TOVS profiles are used (Figure 10b through 13b), the statistics are better for all methods. The rms differences, in particular, are smaller by 2–3 W m⁻². Methods A’ and B’ are less biased than methods A and B by about 2 W m⁻². The bias computed for method D’ is positive,
but when compared to that of method D, its magnitude is also 1 W m\(^{-2}\). The half-hourly comparisons show degraded results in clear sky conditions but improved results in the presence of clouds (Table 7). The rms differences are larger by 5 W m\(^{-2}\) under clear skies and smaller by 1 to 3 W m\(^{-2}\) under cloudy skies. Since only one set of profiles is used in the primed methods, clear sky conditions are represented by a single value (322.6 W m\(^{-2}\)) for the entire experiment, whereas the corresponding range of variations in the surface measurements is about 65 W m\(^{-2}\). This single value is indeed replaced by values scanning part of the \(F^-(0)\) variability when daily-averaged TOVS profiles are used. In cloudy conditions, the prevailing air mass at the comparison site appears to be better represented by the MILDEX-averaged profiles.

6.3. Discussion
A noticeable feature revealed by the comparison statistics is that all the methods exhibit biases. These range from 1 to \(-13\) W m\(^{-2}\). Except for method D', the biases are negative, indicating that \(F^-(0)\) is generally underestimated. Several origins may be attributed to these biases. First, they might be linked to uncertainties in the temperature and humidity retrievals; but Figure 8 indicates that the retrieved atmospheric profiles are generally warmer and more humid than those obtained from radiosondes, which produce an overestimation of \(F^-(0)\) on the average. Another possibility is that the clouds are located too high in the atmosphere and/or that the cloud amounts are undervalued. We note in this respect that the cloud thicknesses computed with methods A and A' are gener-
ally smaller than those expected for the type of clouds observed. Furthermore, the one-layer assumption for the clouds can lead, in methods A and A', to an overestimation of the cloud base altitude in the case of multilayered clouds. The much larger biases of methods B and B' result most likely from uncertainties in Paltridge's [1974a, b] formulas. Using (18) instead of (14), in particular, gives much smaller cloud emissivities. For methods C and C', the selected cloud geometrical thickness, 0.5 km, represents well the mean conditions during MILDEX, inducing smaller biases. The simpler methods (methods D and D') provide the smallest biases, which could be explained by their use of cloud coverage to characterize the effect of clouds, and this parameter is directly estimated from the satellite observations. Indirectly derived parameters, such as cloud base altitude and atmosphere profiles below clouds, which are more prone to errors, are not needed in these methods. The small biases of methods D and D' might be fortuitous, however, and valid only for the type of clouds prevailing during MILDEX.

Another feature common to all methods is that the slopes of best fit are close to 0.5. Thus the satellite methods are only able to reproduce about 50% of the actual $F^{-}(0)$ variability. This can be partly attributed to uncertainties in the slope estimates, but another explanation is that the TOVS retrievals poorly describe the atmospheric changes encountered.

Interestingly, the daily comparisons yield smaller rms differences than the half-hourly comparisons by about 4-5 W m$^{-2}$. Since the effect of time averaging on the noise in the surface data only reduces the rms differences by about 1 W m$^{-2}$, we conclude that time averaging improves the accuracy of the
methods. On time scales other than those considered here, the 10 W m⁻² accuracy for climate studies will be approached.

Even though differences exist in the various methods' statistics, the performance is similar for all methods. The small disparities thus do not warrant favoring one method over the others. It is important to note, however, that methods A and A' have more potential for improvement, since they are based on modeling the physical processes within the atmosphere and clouds. Furthermore, an improvement of a few watts per square meter can be significant for some applications (e.g., interannual variability). The advantage of methods C, C', D and D', however, is that they are applicable at night. In addition, methods D and D' are much easier to implement and much faster computationally.

The accuracy of the methods depends largely on the quality, resolution, and sampling of the retrieved atmospheric profiles and cloud parameters. A primary source of error is linked to inherent uncertainties in the TOVS products. These uncertainties, which typically amount to ±2–3 K for the temperature and 30% for the water vapor mixing ratio, translate directly into $F^-(0)$ errors of ±20–30 W m⁻² in the MILDEX conditions. They also affect indirectly the computations of $F^-(0)$ through the cloud parameters. For instance, a ±2 K uncertainty in the atmospheric temperature induces errors of ±0.2 to 0.5 km on cloud base altitude, which reduces the accuracy of $F^-(0)$ by several watts per square meter. A second important source of error stems from the coarse spatial resolution of TOVS and the need to interpolate TOVS retrievals in cloudy conditions. As a result of cloudiness, the nearest retrievals are made far from the comparison site and therefore might not represent the instantaneous atmospheric conditions existing at the site. Limitations resulting from time sampling
are illustrated in Figure 14, which presents the TOVS retrievals for a particular day, November 9, 1983, together with the evolution of the actual profiles during that day as measured by radiosondes. The situation on November 9, 1983, is characterized by a frontal passage and, consequently, heavy cloudiness and air mass changes. The temperature and water vapor mixing ratio changed respectively by 3 to 7 K and 30 to 50% in the lower troposphere over 24 hours. These changes affect \(-F(0)\) by 30 to 60 W m\(^{-2}\). Important errors can therefore result from the large variability of the profiles during the day, which cannot be adequately described from one polar orbiting satellite. In this respect, retrievals from the VAS instrument aboard a geostationary satellite should provide better monitoring of the daily variability.

Uncertainties in the radiative transfer modeling also constitute a potential source of error in computing \(-F(0)\), but the magnitude of these uncertainties is difficult to assess. Even though the model compares well with more accurate models, few direct comparisons with broadband flux measurements have been performed. Final assessment of the model accuracy cannot be made until more absolute validations are performed. Thus for the time being, further refinement of the model must await these validations.

Another way of assessing the performance of the satellite methods is to compare them with other methods, in particular those which utilize empirical formulas and conventional parameters observed from ships. Such comparisons are of great interest since empirical formulas are employed operationally. We have selected five commonly used formulas and computed \(-F(0)\) for the same situations as above. Table 8 presents statis-
tics when comparing measured $F^-(0)$ with $F^-(0)$ estimated using the empirical formulas. In general, the rms differences are similar to those obtained with the satellite methods, ranging from 21 to 28 W m$^{-2}$ for half-hourly comparisons and from 16 to 23 W m$^{-2}$ for daily comparisons. The empirical formulas, however, except Efimova's [1961], give biases comparable to or larger than that of the most biased satellite method. The bias reaches $-23$ W m$^{-2}$ when Clark et al.'s [1974] formula is employed. Even though Efimova's [1961] formula yields the smallest bias ($-5$ W m$^{-2}$), it predicts poorly the $F^-(0)$ variability. Bunker's [1976] formula appears to provide the best overall results. But even compared with Bunker's [1976] formula, all the satellite methods except B and B' are favored because of their smaller biases. Note that the results of Table 8 cannot be generalized because the empirical formulas involve coefficients that are subject to latitudinal and seasonal variations. On the contrary, the satellite methods are applicable globally, which makes them particularly suitable for long-term variability studies, such as climate studies.

7. SUMMARY AND CONCLUSIONS

Several methods for estimating the downward longwave irradiance at the ocean surface, $F^-(0)$, have been investigated. These methods involve a radiative transfer model and satellite radiance data, and they are intended to be generally applicable.

The radiative transfer model is a simplified version of Morcrette et al.'s [1986] routine for global circulation models. The longwave spectrum is divided into four spectral intervals over which spectral irradiances are evaluated using transmissivity-type functions. Only absorption by the main atmospheric absorbers is taken into account, namely water vapor, carbon dioxide, and ozone. Vertical integration is performed numerically after dividing the atmosphere into layers of arbitrary thickness. Each layer is characterized by its mean temperature, water vapor, carbon dioxide and ozone mixing ratios, fractional cloud coverage, and downward emissivity of clouds. The clouds are assumed to occupy only one layer of the model atmosphere. Therefore the case of clouds located at different levels in the atmosphere and having different radiative properties is not explicitly parameterized. The radiative transfer model is computationally efficient and fairly accurate (a few watts per square meter when compared with line-by-line models), making it very suitable for global applications. To run the model, it is necessary to know instantaneous water vapor and ozone mass mixing ratio profiles, carbon dioxide volume mixing ratio, temperature profile, cloud amount (product of fractional coverage and emissivity), and cloud top and cloud base altitudes.

A sensitivity study was first performed to assess the relative importance of the different model input parameters and to devise strategies for their retrieval. The results show that the most important effect is that of temperature and water vapor mixing ratio variations. Changes in carbon dioxide and ozone amounts affect $F^-(0)$ very little. For a given atmosphere, $F^-(0)$ varies linearly as a function of cloud amount, with the slope depending on the cloud base altitude. Values of $F^-(0)$ in cloudy conditions may depart from clear sky values by 80 W m$^{-2}$. Consequently, carbon dioxide and ozone mixing ratios are specified from climatological data. The other model input parameters, which have the strongest influence on $F^-(0)$ and are highly variable in space and time, are determined from satellite radiance data. Temperature and water vapor mixing ratio profiles are retrieved from NOAA TOVS data; cloud parameters are determined from GOES VISSR observations in the visible and infrared.

Four basic methods, with varying degrees of sophistication, have been presented and discussed. In the most refined method (method A), cloud reflectance in the visible is first determined according to Diak and Gautier [1983]. Stephens' [1978b] theoretical results are then applied to estimate cloud liquid water path and downward and upward cloud emissivities. Fractional cloud coverage is determined from visible data using Gautier et al.'s [1980] threshold technique. While

### Table 6. Performance of the Different Satellite-Based Methods

<table>
<thead>
<tr>
<th>Method</th>
<th>$r$, %</th>
<th>$S_e$, W m$^{-2}$</th>
<th>Bias, W m$^{-2}$</th>
<th>Slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>69</td>
<td>22.6</td>
<td>$-6.5$</td>
<td>0.53</td>
</tr>
<tr>
<td>B</td>
<td>74</td>
<td>24.0</td>
<td>$-13.3$</td>
<td>0.54</td>
</tr>
<tr>
<td>C</td>
<td>55</td>
<td>26.6</td>
<td>$-4.3$</td>
<td>0.46</td>
</tr>
<tr>
<td>D</td>
<td>58</td>
<td>24.9</td>
<td>$-1.2$</td>
<td>0.45</td>
</tr>
<tr>
<td>A'</td>
<td>73</td>
<td>20.6</td>
<td>$-3.9$</td>
<td>0.56</td>
</tr>
<tr>
<td>B'</td>
<td>78</td>
<td>21.8</td>
<td>$-11.0$</td>
<td>0.55</td>
</tr>
<tr>
<td>C'</td>
<td>64</td>
<td>23.6</td>
<td>$-3.8$</td>
<td>0.49</td>
</tr>
<tr>
<td>D'</td>
<td>62</td>
<td>23.7</td>
<td>1.5</td>
<td>0.46</td>
</tr>
</tbody>
</table>

$r$, $S_e$, and Bias are root-mean-square differences between measured and model-derived downwelling longwave irradiance at the ocean surface. Slope is the correlation coefficient.

### Table 7. Comparison of Half-Hourly Averages of Downward Longwave Irradiance

<table>
<thead>
<tr>
<th>Method</th>
<th>Clear</th>
<th>Cloudy</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>11.6</td>
<td>24.3</td>
</tr>
<tr>
<td>B</td>
<td>11.6</td>
<td>24.9</td>
</tr>
<tr>
<td>C</td>
<td>11.6</td>
<td>27.6</td>
</tr>
<tr>
<td>D</td>
<td>11.6</td>
<td>25.8</td>
</tr>
<tr>
<td>A'</td>
<td>16.2</td>
<td>20.9</td>
</tr>
<tr>
<td>B'</td>
<td>16.2</td>
<td>22.3</td>
</tr>
<tr>
<td>C'</td>
<td>16.2</td>
<td>24.2</td>
</tr>
<tr>
<td>D'</td>
<td>16.2</td>
<td>24.3</td>
</tr>
</tbody>
</table>

$r$, $S_e$, and Bias are root-mean-square differences between measured and model-derived downwelling longwave irradiance at the ocean surface. Slope is the correlation coefficient.

Se is standard error, and $r$ is the correlation coefficient.
accounting for atmospheric absorption above the clouds and for nonblackness of clouds, cloud top altitude is obtained from infrared data in a several-step procedure involving the retrieved temperature and water vapor mass mixing ratio profiles, fractional cloud coverage, and upward emissivity of clouds. Finally, cloud base altitude is deduced from cloud top altitude and cloud liquid water path, assuming that the liquid water within the clouds is vertically distributed according to Feigelson [1978]. The method does not apply to icy clouds and is not valid at night. In the other methods, several of the cloud parameters are derived differently. In method B, downward emissivity and geometrical thickness of clouds are obtained from cloud liquid water path via formulas established experimentally byPaltridge [1974a, b]. In method C, cloud thickness is assumed constant (0.5 km), as it was by Darnell et al. [1983]. Finally, in method D, the effect of clouds is simply parameterized as a function of fractional cloud cover. Although the cloud radiative properties are crudely parameterized in methods C and D, the advantage of these methods is that they are applicable at night.

The different methods were validated with surface data acquired during the MILDEX experiment, conducted from October 24 through November 13, 1983, off the central California coast. Comparisons of satellite-derived irradiances with in situ pyrgeometer measurements show that all the satellite methods perform similarly. The best statistics are obtained with the most sophisticated method when TOVS retrievals averaged over the entire period of the experiment are used. For this method, the correlation coefficients and standard errors of estimate are respectively 0.73 and 20.6 W m\(^{-2}\) (half-hourly comparisons) and 0.83 an 15.7 W m\(^{-2}\) (daily comparisons). Compared with this method, the simplest method yields standard errors of estimate that are larger by only 4 W m\(^{-2}\). When daily-averaged TOVS profiles are used, the comparisons give better results in clear sky conditions but degraded results in cloudy conditions, which yield slightly larger standard errors of estimate (1–2 W m\(^{-2}\)). Compared with bulk formula techniques that employ conventional surface data, the satellite methods exhibit similar standard errors of estimate. The satellite methods, however, are favored, since they are generally less biased and theoretically applicable over the global oceans.

Among the satellite methods, methods C and D are recommended for global applications, since they are easier to implement and faster computationally. Method A, which is more complex and not valid at night, should be considered for more regional studies. This method, however, constitutes a research tool which has great potential for improvement. Method B, because of its large bias, appears to be the least attractive of all methods. Still, a better knowledge of the relationships between cloud characteristics should reduce this bias.

A major part of the errors on the satellite methods orig-
inates from uncertainties in the TOVS retrievals. Means of reducing the errors are linked to improving the temperature and water vapor mass mixing ratio profiles in clear and cloudy conditions. Unless a significant improvement in retrieving atmospheric profiles takes place, determining cloud parameters might not require further attention. Recent advances in retrieval procedures [e.g., Chedin et al., 1985; Uddstrom and Wark, 1985] indicate that such an improvement might be possible for clear and partly cloudy conditions. For overcast conditions, however, spatial interpolation is still necessary, which can be improved by using additional physical constraints (e.g., accounting for higher moisture content below clouds). The use of microwave data or new techniques capable of identifying air masses from cloud characteristics can also be envisioned for cloudy conditions. In fact, the advanced microwave sounding unit to be launched in future NOAA satellites will provide all-weather temperature and humidity profiles, giving important support to our methodology. At night, liquid water content can be estimated from microwave data. This approach can be tested now with Nimbus 7 scanning multichannel microwave radiometer data and, in the near future, with special sensor microwave/imager data on DMSP satellites.

So far, the methods presented have only been validated for one oceanic region over a restricted time period. Obviously, they require further validation before they can be applied globally. This will be accomplished in part with data sets collected during the recent Frontal Air-Sea Interaction Experiment (FASINEX) (February 1986) in the subtropical Atlantic.

APPENDIX: EXPOSURE ERRORS IN LONGWAVE IRRADIANCE MEASUREMENTS

The pyrgeometers on board R/P Flip and R/V Acania during MILDEX were deployed to provide measurements of atmospheric downward longwave irradiance. The instrument on board R/P Flip was attached to one of the booms, 18 m from the hull and 3.5 m above mean sea level; the instrument onboard R/V Acania was installed on the rear upper deck at 8 m above mean sea level. These locations for the instruments were carefully selected in order to minimize exposure errors. The instruments, however, did not view the sky over an entire 2π solid angle. Part of the sky was always obscured by emitting surfaces, namely the hull of R/P Flip, a radiosonde hut, and a sodar instrument on the rear upper deck of R/V Acania. These surfaces introduce measurement errors because they are generally warmer than the portion of the sky they obscure.

To evaluate the errors, we first consider the longwave irradiance that would be measured if there were no emitting surface above the pyrgeometers (the ideal case). It can be expressed as

\[ F_\nu^{-}(0) = \int_0^\infty dv \int_0^{2\pi} d\phi \int_0^{\pi} d\psi \frac{\mu L_{\nu}^{-}(\mu, \phi)}{\cos \psi} \]

where \( L_{\nu}^{-} \) is the monochromatic radiance at frequency \( \nu \) received in the direction characterized by zenith angle \( \mu \) and azimuth angle \( \phi \). It is assumed in (A1) that the pyrgeometers measure thermal radiation in the entire electromagnetic spectrum. In fact, these instruments respond to radiation with wavelengths in the 5–50 \( \mu \)m range, but thermal emission outside this range represents only 3% of the total (i.e., entire spectrum) thermal emission and is negligible for the purpose of the study.

Now, if a surface emitting as a blackbody at temperature \( T_s \) occupies a solid angle defined by \( \mu_1, \mu_2, \phi_1, \) and \( \phi_2, \) the longwave irradiance received within this solid angle is

\[ F_\nu^{-}(O) = \int_0^\infty dv \int_{\mu_1}^{\mu_2} d\mu \int_0^{2\pi} d\phi [\mu L_{\nu}^{-}(\mu, \phi)] \]

\[ = \sigma T_s^4(\phi_2 - \phi_1)(\mu_2^2 - \mu_1^2)/2\pi \]  

(A2)

where \( \sigma \) is the Stefan-Boltzmann constant. In (A2), atmospheric effects along the path between the emitting surface and the pyrgeometers are neglected.

Within the same solid angle, the pyrgeometers would receive from the atmosphere

\[ F_\nu^{-}(0) = \sigma T_s^4(\phi_2 - \phi_1)(\mu_2^2 - \mu_1^2)/2\pi \]  

(A3)

where \( T_s \) is an effective emitting temperature for the portion of the sky obscured by the surface.

Therefore the error introduced in the pyrgeometer measurements is

\[ F_\nu^{-}(O) - F_\nu^{-}(0) = \sigma(T_s^4 - T_s^4)(\phi_2 - \phi_1)(\mu_2^2 - \mu_1^2)/2\pi \]  

(A4)

Equation (A4) was applied to our case. Values of 0.45 and 0.0 (R/P Flip) and of 0.34 and 0.0 (R/V Acania) were estimated for \( \mu_1 \) and \( \mu_2 \), respectively, and values of 29° (R/P Flip) and 86° (R/V Acania) were estimated for \( \phi_1 \) and \( \phi_2 \). Temperatures \( T_s \) and \( T_s \) were more difficult to estimate, since they depend on the variable atmospheric conditions and on R/P Flip and R/V Acania heat sources. We note, however, that \( \mu_1 \) and \( \mu_2 \) correspond to zenith angles not far from the horizontal. Therefore \( T_s \) must be close to the air temperature in the vicinity of the instruments and emitting surfaces (ambient air temperature). On the other hand, \( T_s \) is generally higher than the ambient air temperature. Large \( T_s \) values occur during the day under clear skies; small \( T_s \) values occur at night when there is no solar heating. Thus we expect \( T_s \) to be generally higher than \( T_s \) i.e., the pyrgeometers will measure larger longwave irradiances than the ones that would be measured if there were no contaminating surfaces.

Figure A1 gives \( F_\nu^{-}(O) - F_\nu^{-}(0) \) as a function of \( T_s - T_s \) for \( T_s = 291 \) K (the mean near-surface air temperature observed during MILDEX). The \( F_\nu^{-}(O) - F_\nu^{-}(0) \) values are below the noise level of the instruments (about 7 W m\(^{-2}\)) when \( T_s - T_s \) is less than 37 K (R/P Flip) and 56 K (R/V Acania). Larger \( T_s - T_s \) values would give significant errors but can only occur during periods of low wind speed and intense insolation. In fact, the actual \( T_s - T_s \) values are generally much smaller,
likely to remain within the range 0–10 K, introducing negligible exposure errors (<1 W m⁻²). It is also clear that the exposure errors, because of fluctuations in $T_e$ and $T_o$, cannot produce a significant bias between R/P Flip and R/V Acania measurements.

The present calculations, based on very simple physics, can only provide an order of magnitude for the exposure errors. In particular, it was not possible to adequately estimate $T_s$ from the MILDEX data set. Reflection effects by the contaminating surfaces were not explicitly considered; it was only assumed that the surfaces emit as blackbodies. Also, the fact that the sensitivity of the pyrgeometers drops considerably at high zenith angles was not taken into account, which results in an overestimation of the actual errors. The assumptions and approximations that were made, however, have no dramatic consequence on the general conclusions of the study.

**NOTATION**

$a$ ozone absorptance for direct solar energy flux.

$a'$ ozone absorptance for diffuse solar energy flux.

$A_i$ cloud amount.

$B_i$ Planck function integrated over spectral interval $i$.

$B_c$ monochromatic Planck function.

$C$ fractional cloud coverage.

$d$ weighted distance from TOVS retrievals to comparison site.

$F_0$ downward longwave irradiance.

$F_0'$ clear sky downward longwave irradiance.

$F_a$ atmospheric downward longwave irradiance that would be observed within the solid angle occupied by contaminating surfaces.

$F_c$ downward longwave irradiance for cloudy layer $n$ (clouds are assumed black).

$F_s$ downward longwave irradiance within the solid angle occupied by contaminating surfaces.

$g$ acceleration due to gravity.

$H$ cloud geometrical thickness.

$i$ index of model spectral interval.

$I_c$ cloudy sky VISSR visible irradiance.

$I_s$ clear sky VISSR visible irradiance.

$j$ index of absorption type.

$k$ index of model atmospheric layer.

$l$ index of Gaussian angle.

$n$ index of cloudy layer.

$L_c$ downward monochromatic infrared irradiance.

$L_c^*$ VISSR infrared irradiance.

$L_e$ VISSR infrared irradiance emitted by black clouds.

$L_s$ clear sky VISSR infrared irradiance.

$N$ number of model layers.

$p$ pressure.

$p_0$ surface pressure.

$p^*$ arbitrary atmospheric pressure level.

$q$ water vapor mixing-ratio.

$R_c$ cloud reflectance.

$R_s$ surface reflectance.

$t_{i0}$ transmissivity-type function for energy exchange between top of atmosphere and surface.

$t_{ij}$ transmissivity-type function for energy exchange between model layers and surface.

$t_c$ monochromatic transmittance.

$t_s$ monochromatic transmittance for absorption type $j$.

$S_0$ extraterrestrial solar irradiance in the VISSR visible channel.

$T$ temperature.

$T_e$ effective sky temperature.

$T_o$ temperature of contaminating surfaces.

$u_0$ total precipitable water.

$u^*$ precipitable water above $p^*$.

$w$ weight associated with $i$th Gaussian angle.

$W$ vertically integrated liquid water amount.

$z$ vertical coordinate.

$z_b$ cloud base altitude.

$z_k$ altitude of model layer base.

$z_t$ ordinate of $i$th Gaussian angle.

$z_c$ cloud top altitude.

$Z$ altitude of top of atmosphere.

$\alpha$ reflection coefficient for direct solar radiation.

$\alpha'$ reflection coefficient for diffuse solar radiation.

$\beta$ backscattered fraction of solar radiation.

$\epsilon_1^+$ 10- to 12-µm upward emissivity of clouds.

$\epsilon_1^-$ broadband downward emissivity of clouds.

$\gamma$ diffusivity factor.

$\mu$ cosine of zenith angle.

$\mu_1$, $\mu_2$ cosine of zenith angles characterizing the solid angle occupied by contaminating surfaces.

$\mu_0$ cosine of solar zenith angle.

$\nu$ radiation frequency.

$v_{16}$, $v_{21}$ lower and upper frequency limits of spectral interval $i$.

$\sigma$ Stefan-Boltzmann constant.

$\phi$ azimuth angle.

$\phi_1$, $\phi_2$ azimuth angles characterizing the solid angle occupied by contaminating surfaces.

$\tau$ cloud optical thickness.

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