A Model of Longwave Irradiance for Use with Surface Observations

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ABSTRACT

Longwave irradiance is often poorly estimated in heat budget calculations for the sea surface. A model using the same hourly surface data as a successful shortwave irradiance parameterization (Lumb, 1964) is developed. Cloud layers in each hour's observation are assigned effective emittances. Cloud temperatures are calculated from temperature profiles constructed using heights and temperatures interpolated from synoptic charts. Contributions from the water vapor band are calculated using the temperature profile to the height of the lowest cloud base (up to 1000 m) and a linear profile of mixing ratio calculated from the surface value. The model compares to within 10 W m⁻² over 2-week periods with measurements of longwave irradiance on two ships in the North Atlantic during three months in the summer season. Existing shortwave models and this model can be combined to give better estimates of the radiation balance at the sea surface.

1. Introduction

Radiation measurements are not always available at sea. Instrumentation is difficult to maintain and ocean weather ships are few and far between. Shortwave irradiance parameterizations using surface weather observations have shown some success (Lumb, 1964; Kasten and Czepalak, 1980; Lind, 1981).

Variations in the net longwave radiation balance are primarily due to changes in atmospheric irradiance $E_l$. Hourly mean $E_l$ can change by up to 100 W m⁻² over short periods in midlatitudes depending primarily upon cloud cover and atmospheric temperature profiles. Use of long-term mean atmospheric irradiances sacrifices potentially important transient information. A model estimating long-wave irradiance using routinely available ship surface observations and upper air data was developed. Each reported cloud layer contributes to the total irradiance according to its coverage, temperature and effective emittance. Water vapor is treated as a grey body and considered to the height of the lowest cloud base, or 1000 m. Atmospheric irradiance outside the water vapor window region is considered constant (Swinbank, 1963; Paltridge, 1974).

Formulas introduced to estimate clear sky atmospheric irradiance using combinations of water vapor pressure $e$, water surface temperature $T_s$, and air temperature $T_a$ were tested by Simpson and Paulson (1979). They found that all of these formulas gave results to within 20 W m⁻² of their measurements.

Simpson and Paulson applied linear “cloud factors” to clear sky formulas and found them accurate to 20 W m⁻² for daily averages in cloudy skies.

With estimates made using surface data alone, no account is taken of variations in profiles of temperature or water vapor mixing ratio. Reed (1975) found that common subtropical inversions could introduce an error of up to 15 W m⁻² in the irradiance estimate.

Our model uses effective cloud emittances $\epsilon$, estimated cloud base temperatures and boundary-layer water vapor effects from surface observations and synoptic charts. This model is compared to measurements of longwave irradiance by two ships during the 1978 JASIN Experiment over the Eastern North Atlantic. Results show the model is accurate to better than 10 W m⁻² for daily averages.

2. Cloud temperature estimates

A profile of atmospheric temperature was constructed for each hour using observed air temperatures and lapse rates determined by using heights $Z$ and temperatures $T$ from twice-daily hemispheric analyses of 850, 700 and 500 mb surfaces.

A lapse rate of $6^\circ$C km⁻¹ is assumed for the lowest 800 m. From 800 m to the height of the 850 mb surface, a lapse rate of $[(T(850\text{ mb})-T_s)/Z(850\text{ mb})]$ was used. Lapse rates for the layers from 850 to 700 mb and 700 to 500 mb were determined using $\Delta T/\Delta Z$ interpolations taken from the upper air analyses.

Hourly cloud base temperatures were calculated by using the cloud base heights reported in the surface weather observation and the temperature profile. WMO cloud height codes represent a height range (e.g., 50–100 m). For the model, mid-points of each range were used.
3. Low-level cloud effective emittance

Eq. (1) defines effective emittance as discussed by Stephens (1980), Cox (1976) and others, which includes effects of cloud transmission, but ignores infrared reflection from cloud to cloud and ground to cloud which, however, amounts to only a few percent (Yamamoto et al., 1970):

\[
\epsilon = \frac{E_{\text{bot}} - E_{\text{top}}}{\sigma T_{\text{aq}}(\text{bot}) - E_{\text{top}}},
\]

(1)

where top and bot refer to top and bottom of cloud, \( \sigma \) is Stefan-Boltzmann constant and \( E_t \) is irradiance observed at subscripted level.

Relationships between liquid water path LWP of low-level layer clouds and effective emittance have been observed by Paltridge (1974), Stephens (1978) and others. Fitted relations take the form of

\[
\epsilon = 1 - \exp(a_0 \text{ LWP}),
\]

(2)

where \( a_0 = -70 \text{ m}^2 \text{ kg}^{-1} \) was used in the model.

Paltridge noted that measured liquid water content increased almost linearly with height for stratiform water clouds. Assuming this generally applies to all stratiform clouds, the total LWP is then proportional to the square of the cloud layer thickness \( \Delta Z \) as in (3):

\[
\text{LWP} = b_0 (\Delta Z)^2.
\]

(3)

Paltridge found \( b_0 = 1.8 \times 10^{-7} \text{ kg m}^{-4} \) when \( \Delta Z \) is given in meters.

Cloud thicknesses for stratus (St) and stratocumulus (Sc) were estimated from their coverage assuming that as coverage decreases, thickness decreases in proportion. Table 1 shows low-level cloud layer thicknesses assigned to layer coverages and corresponding effective emittances from (2).

Cumulonimbus (Cb, coded CL.3) and towering cumulus (Cu, coded CL.2) emittances are close to unity. Their emittances were set to 0.96, the same as a 500-m thick deck of Sc. Other cumulus (coded CL1 or genus code C8) have a greater LWP than the same amount of Sc or St. Emittances for these clouds were assigned from Table 1 using a cloud cover two oktas higher than reported. By this method, four oktas of Cu have the same effective emittance as six oktas of Sc or St. The method was used only to assign cumulus emittances. Irradiance calculations used the actual reported cloud cover for the layer.

4. Middle and high cloud effective emittance

No relationship between upper-level cloud thickness and emittance was found in the literature. Platt (1973) attempted to find a correlation between these two parameters for cirrus, but only found a linear correlation of 0.19 for 209 cloud cases. Stephens (1980) showed that ice water content was mildly correlated to emittance for cirriform clouds. Emittance measurements of cirriform clouds reveal a wide range of values (Platt and Dilley, 1981) with a mean of \( \sim 0.50 \). Allen (1971) found cirriform cloud effective emittances to average somewhat less than this, \( \sim 0.35 \).

Middle-level clouds have broad ranges of both thickness and water content (Platt and Bartusek, 1974). Relations between effective emittance and middle cloud thickness were not found. Since surface observations have no information on cloud LWP, effective emittances were assigned to middle and high clouds by cloud type as shown in Table 2. A discussion of the sensitivity of the model to middle- and upper-level cloud effective emittances can be found in Section 8.

5. Water vapor effects

The model assumes that water vapor mixing ratio \( q \) decreases linearly with height to the lowest cloud base or 1000 m, whichever is smaller. Eq. (4) de-

\[
\text{LWP} = b_0 (\Delta Z)^2.
\]
scribes \( q(Z) \) from the measurement of \( q(\text{sfc}) \):

\[
q(Z) = -1.4 \times 10^{-3} Z + q(\text{sfc}).
\]  

(4)

Water vapor flux emittance \( \epsilon_f \) can be determined if the water vapor optical path length \( u \) is known. Water vapor mixing ratio and \( u \) are related by

\[
u = \int_{\rho_0}^{\rho} \rho \, dZ = \frac{1}{g} \int_{p}^{\infty} qdp,
\]

(5)

where \( \rho_0 \) is density of water and \( g \) is gravitational constant.

The flux emittance of a layer of water vapor can be computed from (6) if the relationship between \( \epsilon_f \) and \( u \) is known:

\[
de \epsilon_f = \gamma \, du,
\]

(6)

where \( \gamma \) is a constant of proportionality which is only a function of optical depth.

The relation between \( \epsilon_f \) and \( u \) has been evaluated at 293 K (Fleagle and Businger, 1980). It is approximated by

\[
\epsilon_f = 0.157 \log_{10}(u) + 0.14.
\]

(7)

A hypothetical sounding was constructed with profiles of temperature and mixing ratio assuming \( \Delta Z/ \Delta p = -8 \text{ m mb}^{-1} \). A range of surface water vapor mixing ratios were inserted and curves were calculated using Eqs. (4) through (7) to relate \( \epsilon_f \), \( Z \) and \( q(\text{sfc}) \). Eq. (8) approximates these curves:

\[
\epsilon_f = C_1 \log_{10}(Z) + C_2 + C_3 \left[ q(\text{sfc}) - 7.5 \right],
\]

(8)

where \( C_1 = 0.157 \), \( C_2 = 0.14 \) and \( C_3 = 0.02 \).

The water vapor layer is now considered as a greybody slab with emittance \( \epsilon_{fs} \) and a mean equivalent radiative temperature \( T \). For computational purposes, the simple mean temperature of the layer below the lowest reported cloud base or 1000 m, whichever is smaller, was used. This slab absorbs longwave irradiance from above \( (1 - \epsilon_f)E_i \) and emits \( \epsilon_{fs}T^4 \).

This method, treating the below-cloud layer as a grey body slab, is oversimplified for use in tropical applications as no allowance is made for the 8–12 micrometer water vapor pressure broadened continuum absorption.
### Table 3. Daily mean hourly estimated and measured atmospheric irradiance for M/V Endurer (60°15′N, 14°30′W) and R/V Meteor (59°N, 12°30′W) during JASIN, Phase I. All values are in W m⁻² or percentages.

<table>
<thead>
<tr>
<th>Date (1978)</th>
<th>Endurer</th>
<th></th>
<th></th>
<th></th>
<th>Meteor</th>
<th></th>
<th></th>
<th></th>
</tr>
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<tbody>
<tr>
<td></td>
<td>Model</td>
<td>Measured</td>
<td>Error</td>
<td>Percent</td>
<td>Model</td>
<td>Measured</td>
<td>Error</td>
<td>Percent</td>
</tr>
<tr>
<td>7/23</td>
<td>358.2</td>
<td>357.2</td>
<td>1.0</td>
<td>0.28%</td>
<td>360.1</td>
<td>362.7</td>
<td>-2.6</td>
<td>-0.72%</td>
</tr>
<tr>
<td>7/24</td>
<td>353.0</td>
<td>352.9</td>
<td>0.1</td>
<td>0.03%</td>
<td>362.5</td>
<td>364.1</td>
<td>-1.6</td>
<td>-0.45%</td>
</tr>
<tr>
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<td>345.6</td>
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<td>-0.57%</td>
<td>356.8</td>
<td>358.8</td>
<td>-2.0</td>
<td>-0.58%</td>
</tr>
<tr>
<td>7/26</td>
<td>346.1</td>
<td>348.6</td>
<td>-2.5</td>
<td>-0.71%</td>
<td>363.1</td>
<td>368.5</td>
<td>-5.4</td>
<td>-1.46%</td>
</tr>
<tr>
<td>7/27</td>
<td>350.1</td>
<td>346.7</td>
<td>3.4</td>
<td>0.98%</td>
<td>364.1</td>
<td>369.4</td>
<td>-5.4</td>
<td>-1.44%</td>
</tr>
<tr>
<td>7/28</td>
<td>327.5</td>
<td>309.2</td>
<td>18.2</td>
<td>5.90%</td>
<td>329.5</td>
<td>331.0</td>
<td>-1.5</td>
<td>-0.46%</td>
</tr>
<tr>
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<td>323.8</td>
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<td>9.7</td>
<td>3.08%</td>
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<td>329.2</td>
<td>-1.0</td>
<td>-0.32%</td>
</tr>
<tr>
<td>7/30</td>
<td>326.3</td>
<td>314.4</td>
<td>11.9</td>
<td>3.78%</td>
<td>314.6</td>
<td>313.6</td>
<td>1.0</td>
<td>0.32%</td>
</tr>
<tr>
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<td>326.7</td>
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<td>324.2</td>
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<td>0.76%</td>
</tr>
<tr>
<td>8/01</td>
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<td>357.6</td>
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</tr>
<tr>
<td>8/02</td>
<td>340.9</td>
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<td>346.6</td>
<td>354.0</td>
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<td>-2.11%</td>
</tr>
<tr>
<td>8/03</td>
<td>*-</td>
<td>*-</td>
<td>*-</td>
<td>*-</td>
<td>362.1</td>
<td>364.2</td>
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<td>-0.58%</td>
</tr>
<tr>
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</tr>
<tr>
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<td>351.2</td>
<td>-0.9</td>
<td>-0.25%</td>
</tr>
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<tr>
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<td>350.5</td>
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</tr>
<tr>
<td>Mean**</td>
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<td>336.6</td>
<td>5.6</td>
<td>1.66%</td>
<td>348.8</td>
<td>351.1</td>
<td>-2.3</td>
<td>-0.64%</td>
</tr>
</tbody>
</table>

* No data available.
** Mean hourly values for Phase I.

### 6. Calculation of atmospheric irradiance

With the relations estimating cloud base temperature, cloud effective emittance and water vapor flux emittance, the cloud coverage information from each hour's surface observation is used to estimate each layer's contribution to the total longwave irradiance. The model equations are

\[
E(3) = N_s(3)\epsilon(3)\sigma T(3)^4 + [1 - N_s(3)]E(4), \quad (11)
\]

\[
E(2) = N_s(2)\epsilon(2)\sigma T(2)^4 + [1 - N_s(2)]E(3), \quad (12)
\]

\[
E(1) = N_s(1)\epsilon(1)\sigma T(1)^4 + [1 - N_s(1)]E(2), \quad (13)
\]

\[
E(\text{tot}) = (1 - \epsilon_f)E(1) + E(\text{sky}) + \epsilon_f\sigma T^4, \quad (14)
\]

where

\[
\epsilon_f = \text{cloud effective emittance}
\]

\[
\epsilon_s = \text{fractional coverage of subscripted cloud layer}
\]

\[
N_s = \text{fractional cloud coverage}
\]

\[
N_s = \text{total fractional cloud coverage}
\]

### Table 4. As in Table 3, but for Phase II.

<table>
<thead>
<tr>
<th>Date (1978)</th>
<th>Endurer</th>
<th></th>
<th></th>
<th></th>
<th>Meteor</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Model</td>
<td>Measured</td>
<td>Error</td>
<td>Percent</td>
<td>Model</td>
<td>Measured</td>
<td>Error</td>
<td>Percent</td>
</tr>
<tr>
<td>8/22</td>
<td>345.6</td>
<td>347.9</td>
<td>-2.3</td>
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<td>351.0</td>
<td>356.4</td>
<td>-5.4</td>
<td>-1.51%</td>
</tr>
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<td>8/23</td>
<td>331.1</td>
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<td>350.1</td>
<td>-4.4</td>
<td>-1.27%</td>
</tr>
<tr>
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<td>319.9</td>
<td>310.6</td>
<td>9.3</td>
<td>2.99%</td>
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<td>331.9</td>
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<td>-0.16%</td>
</tr>
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<td>336.9</td>
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<td>364.2</td>
<td>-4.0</td>
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</tr>
<tr>
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<td>355.9</td>
<td>2.3</td>
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<td>358.1</td>
<td>361.8</td>
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<td>-1.02%</td>
</tr>
<tr>
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<td>361.9</td>
<td>0.8</td>
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<td>358.4</td>
<td>360.5</td>
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</tr>
<tr>
<td>8/29</td>
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<td>359.7</td>
<td>365.0</td>
<td>-5.3</td>
<td>-1.44%</td>
</tr>
<tr>
<td>8/30</td>
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<td>1.6</td>
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<td>366.6</td>
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</tr>
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<td>354.0</td>
<td>-2.3</td>
<td>-0.64%</td>
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<td>1.03%</td>
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<td>342.6</td>
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<td>-0.25%</td>
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<tr>
<td>9/02</td>
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<td>350.5</td>
<td>2.3</td>
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<td>358.5</td>
<td>357.2</td>
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<td>0.36%</td>
</tr>
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<td>5.6</td>
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<tr>
<td>Mean*</td>
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<td>0.41%</td>
<td>350.5</td>
<td>353.3</td>
<td>-2.8</td>
<td>-0.79%</td>
</tr>
</tbody>
</table>

* Mean hourly values for Phase II.
\( T = \text{cloud base temperature of subscripted cloud layer} \)
\( \varepsilon = \text{effective emittance of subscripted cloud layer} \)
\( \varepsilon_f = \text{flux emittance of layer from surface to lowest reported cloud base} \)
\( T = \text{mean equivalent radiative temperature of same layer. For computational purposes, this has been replaced by the simple mean temperature of layer.} \)
\( \sigma = \text{Stephan-Boltzmann constant (5.67 \times 10^{-8}} \text{ W m}^{-2} \text{ K}^{-4}) \)
\( E(\text{sky}) = \text{irradiance originating from clear sky between cloud layers centered at 500 mb.} \)
\( E(4) \) through \( E(1) \) are cumulative irradiances from cloud layers above and including the subscripted level.

The model, simplified to \( E_t = \varepsilon_f \sigma T^4 \), was compared to clear sky irradiances and it was found that the contribution from low-level water vapor did not account for the total irradiance by a deficit of 50–100 W m\(^{-2}\). No account was made for wavelengths outside the water vapor window region and all constituents above 1000 m were neglected. Without radiosondes, water vapor profiles above the atmospheric boundary layer are unknown. Keeping the model as simple as possible, it was found that if a level in the atmosphere was considered black and its irradiance (attenuated by the water vapor layer) were added, the clear sky irradiance could be more closely approximated. This added irradiance was calculated by using the 500 mb temperature. In cases with scattered or broken clouds, only the clear sky portion of the sky was represented by the irradiance calculated in this way.

Fig. 1 shows the origins of each term in Eqs. (9)–(14). Profiles of temperature and water vapor mixing ratio are shown at the left. In this example, Eqs. (10) and (11) would not be used and \( E(4) \) and \( E(3) \) would be set to zero.

7. Model comparisons with JASIN data

The Joint Air–Sea Interaction (JASIN) Experiment (Pollard, 1978) took place in the Eastern North Atlantic from mid-July to early September 1978. Radiation measurements were made on two ships, the M/V Gardline Endurer (60°15'N, 14°30'W) and the R/V Meteor (59°N, 12°30'W). Both ships were equipped with Eppley pyrgeometers, model PIR (Eppley Laboratories, 1971; Albrect and Cox, 1977). A mirrored silicon hemisphere with a nearly constant spectral response from 4 to 50 micrometers was used. Compensation for detector temperature was achieved by use of a thermistor–resistor–battery circuit. Corrections for radiative imbalances within the housing were made using bead thermistors mounted on the inner surface of the dome and within the base housing the thermopile. Accuracies were found to be better than 1% during JASIN ship to ship intercomparisons.

a. M/V Endurer comparison

The M/V Endurer used a standard ship height code (WMO code FM-11) for reporting cloud height estimates. This code does not allow accurate height estimates for clouds higher than 2500 m; therefore, clouds coded 98 and 99 were assigned heights of 3200 and 5000 m, respectively. The resolution with each coded value decreases as cloud height increases. Heights and temperatures were interpolated to the ship position from 0000 and 1200 GMT analyses of 850, 700 and 500 mb surfaces.

b. R/V Meteor comparison

After the model was finalized using the M/V Endurer data set, the model was independently tested with the surface observations and data from the R/V Meteor. R/V Meteor cloud heights were reported in hundreds of meters giving equal resolution for all heights. Heights and temperatures were interpolated to the ship position as for M/V Endurer. Hourly measured atmospheric irradiances from the R/V Meteor were provided by M. Gube (1979).

Tables 3 and 4 list model and measured atmospheric irradiances plus errors averaged over each day for JASIN, Phases I and II, respectively. Daily mean model-measurement errors ranged from \(-2.5 \text{ to } +18.2 \text{ W m}^{-2} (-0.71\% \text{ to } +5.90\% \text{ for M/V Endurer and } -7.6 \text{ to } 3.8 \text{ W m}^{-2} (-2.15\% \text{ to } +1.15\% \text{ for R/V Meteor).} \)
for R/V Meteor. An error summary for both ships over the entire experiment is shown in Table 5.

Hourly measured and estimated atmospheric irradiances as well as model-measurement differences from M/V Endurer are shown in Fig. 2, and from R/V Meteor in Fig. 3. Overestimation by the model on 28, 30, 31 July and 24 August for the M/V Endurer data set occur at lower irradiances (see also Table 6). Checks of these data showed that the cloud observations contained cirriform clouds. The method of assigning heights of 5000 m to these upper-level clouds could have been improved by using a greater height. Overestimation from using emittances that are too large for high clouds can be ruled out as the

Fig. 2. Time series of hourly mean longwave irradiance measurements and model estimates (dotted) from M/V Endurer during JASIN 1978. (a) Phase I, (b) Phase II. Model-measurement differences are shown above.
errors from the R/V Meteor data set were of the same magnitude for high clouds as when low clouds dominated the observed sky conditions.

8. Model sensitivity

The atmospheric irradiance model is influenced most strongly by cloud-base height estimates which affect estimates of cloud-base temperature and water vapor flux emittance. Cloud coverage also strongly affects the model, as one okta (12.5%) is the smallest reported resolution. For a single low cloud layer case characteristic of JASIN, a change of one okta coverage changes the model results by about 10 W m\(^{-2}\). Variations in coverage of cloud at 3200 m do not affect the model longwave irradiance received by the surface as much as the low cloud coverage. As coverage decreases, clear sky irradiance, which is calculated by the approximation described in Section 6, increases. The magnitudes per unit sky cover are nearly equal, so the effect of coverage is diminished.

The model is more sensitive to cloud base tem-

Fig. 3. As in Fig. 2, but for R/V Meteor.
temperatures for higher clouds. A change in cloud base temperature from 285 to 280 K results in a 13 W m\(^{-2}\) difference in the calculated irradiance, whereas a change from 270 to 265 K results in a 19 W m\(^{-2}\) change in model irradiance. Errors in cloud height estimates for low clouds are generally less than 200 m, or \(\sim 1\)°C. Upper-level cloud height estimate errors can be large, \(\sim 1\) km, which is approximately equivalent to 6°C. In addition, irradiance variations due to changes in low cloud heights are somewhat compensated by changes in the depth of the water vapor layer.

A sensitivity test was performed setting upper-cloud emittances to 1.0 and then 0.0, as large ranges of observed upper-level cloud emittances have been reported in the literature. Unfortunately, upper-level clouds were only reported in \(\sim 10\)% of the cloud observations. Of these, the large majority were over the M/V Endurer position. The test showed that changing the upper-level cloud emittance from 1.0 to 0.0 changed the model irradiances by \(\sim 12\) W m\(^{-2}\) from the original model irradiance for six oktas of upper-level cloud. This test clearly demonstrates that upper-level clouds are least significant in contributions to the surface observed atmospheric irradiance for this particular region.

9. Summary of Model Results

A model estimating atmospheric irradiance at the ocean surface has been compared to measurements. The model has an average absolute error of 10.3 W m\(^{-2}\) with surface observations from the M/V Endurer and 6.3 W m\(^{-2}\) with the observations from the R/V Meteor over 30 days. Hourly means averaged over the experiment are accurate to better than 4 W m\(^{-2}\). Errors ranged from 0 to 18 W m\(^{-2}\) in the calculation of daily mean averages. Root-mean-square errors for three-week periods ranged from 8 to 15 W m\(^{-2}\). The standard deviation of measured atmospheric irradiance during JASIN was 22 W m\(^{-2}\).

The use of instantaneous surface observations to estimate hourly means of atmospheric irradiance does not create the problems it does for short-wave models (Kasten and Czeplak, 1980) as cloud position with respect to the sun is not as important. Surprisingly, model irradiances for observations obtained during darkness, when compared with measure-

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mements, have nearly the same accuracy as during daytime.

Errors in reported cloud coverage have the strongest effect in the model-computed irradiances. A large percentage of model-measurement differences could be due to cloud observation errors, as cloud coverage errors of one okta are common. The coverage code requires coverages with any breaks, no matter how small, to be coded other than eight oktas.

Upper-level cloud emittances are difficult to assign from surface observations. Wide ranges of measured emittances for these upper-level clouds show that their parameterization is difficult. Fortunately, tests of model sensitivity showed that for surface-measured atmospheric irradiance, high clouds have the smallest effect.

The use of an approximate temperature profile helps our model to better estimate the mean atmospheric irradiance. Simpler models (see Simpson and Paulson) depend on surface information alone.

Extension of the model to other latitudes and seasons must await further testing. Water vapor above the lowest cloud level will have to be included in tropical applications. Since no allowance was made for the 8 to 12 micrometer water vapor pressure broadened continuum absorption, a more realistic model of the boundary-layer water vapor will also have to be used. Cox (1973) describes a model which could be adapted for use with readily and routinely available data for tropical applications. Verification for other latitudes and seasons would make the model a stronger tool for regional, seasonal or climatic calculations of the radiative balance at the sea surface.

A method of modeling atmospheric irradiance from routinely available data sources has been shown to verify well for summer and early fall conditions over the North Atlantic. The model was developed with the M/V Endurer data base and tested independently with surface observations and data from the R/V Meteor. Model irradiances verify well with measurements from both JASIN ships for time scales from a few hours to weeks.

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