An Optical Model for Heat and Salt Budget Estimation for Shallow Seas

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An optical model for heat and salt budget estimation for shallow seas

Hari Warrior1 and Kendall Carder2

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The effects of the underwater light field on heat-budget calculations for shallow waters are developed and applied for the region of Bahamas. Most of the general circulation models use a simplified heat budget scheme based on Jerlov water types, and do not account for optical bottom effects. By optical bottom effect, we mean the bottom absorption and reflection of the short-wave radiation, which in turn affects the thermal stratification and heat exchange with the atmosphere. In this paper, this optical bottom effect is added to a 3D turbulence model (a 1D model called GOTM is coupled to a 3D model called POM) and the evolution of the temperature structure studied. We call the coupled model 3DGOTM. This optical bottom effect is found to be important in the areas with clear water, shallow depths and small solar zenith angle. On the basis of the coastal meteorological measurements from Andros Island, we have used this three-dimensional turbulence closure model (3DGOTM) to show the influence of bottom reflection and absorption on the sea surface temperature field. The final temperature of the developed water column depends on water depth and bottom albedo. Effects of varying the bottom albedo were studied by comparing results for coral sand and sea grass bottoms. This has an appreciable contribution to the heat budget and salt budget of the shallow waters in these coastal regions. The salinities of the shallow regions near Andros Island have been found to reach as high as 46 psu by summer. In addition to the thermohaline plumes generated by these bottom effects, this warming process has an impact on the moisture feedbacks into the atmosphere due to evaporation.


1. Introduction

Study of the heat budget of the oceans is important to correctly simulate the ocean circulation. Heat budget of the oceans is the sum of the long-wave and short-wave components. Clear parameterization of the effects of this radiation and its penetration of the water column is important to improve models of heat budget, coral bleaching and benthic photosynthesis. Because of high absorption by water molecules at longer wavelengths, about 57% of the total radiation (the long-wave component) is absorbed in the top meter or less. Therefore the heating rate from infrared radiation is usually treated separately from absorption of visible (400–700 nm) radiation which is much more penetrative and thus sensitive to the water constituents (Smith and Baker [1981] and Pope and Fry [1997], both dealing with pure water). In this paper, we are concerned with the short-wave (400–700 nm) radiation and its fate as it passes the air-sea interface down to the shallow bottoms.

In this paper, we bring to attention of the scientific community a neglected term in the heat budget equations of general ocean models, which is especially important in certain regions. This term is negligible when one considers waters deeper than 10–15 m but quite significant in shallow areas with clear water and smaller solar zenith angle. Through its influence on the salt budget, this shallow water heating affects the salinity of the adjoining deeper seas.

This neglected term is the light (visible radiation) that reaches the bottom. Depending on the albedo of the bottom (sea grass is less reflective than coral sand), a part of this light is absorbed by the bottom. The rest is reflected back up. The absorbed part of the radiation is converted into heat and is reradiated into the water column. This convectively heats up the water column from below. The temperatures of the shallow bottom waters can be significantly warmer due to this effect, as we show in the following sections.

This paper will be concerned with study of the water masses near the Andros Island of Bahamas. Waters over these shallow sub-tropical banks can become much hotter in the spring than waters found in adjacent deeper regions (e.g., Figures 1a and 1b). In addition to being warmer than the surrounding waters, some of these shallow regions are known to produce hyper-saline waters (waters with salinity greater than 37 psu) due to excessive evaporation. The presence of such hyper-saline waters as high as 46 psu has been observed by Cloud [1962] and later by many other scientists (Figure 2).

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Figure 1. AVHRR image of SST for (a) 15 April 2001 at 6:00 am. (b) 17 April 2001 at 6:00 pm, color scale in °C (AVHRR images courtesy of F. Muller-Karger).
[6] The presence of the Andros Island provides shelter to these waters from the effects of trade winds. There is also a less than moderate tidal exchange with the deeper seas. Broecker [1966], showed that these high-salinity pools of water have residence times of longer than 300 days. Physical observations have shown the seasonal pattern of these sinking waters. For the year 2000 of observations, it was found that these hyper-saline waters are so dense that they can sink to depths of at least 45 m in summer and 75 m in winter [Smith, 1995; Hickey et al., 2000; Otis et al., 2004]. These particular depths of penetration are functions of the increase in salinity over the shallow banks, and the depth of the thermocline in the deeper waters. Since the atmospheric forcings are typically similar over the years, it can be assumed that these depths of penetration are repeatable over the years.

[7] The actual thermal effects of shallow waters often have important consequences for the salt budget, chemical and biologic formations like “whitings”, aragonite precipitation [Broecker et al., 2000], hot brine effects on coral [Lang et al., 1988], high humidity, and even on cloud formation over shallow banks. There have been numerous observations of coral bleaching due to the excessive warming of shallow areas. Coral reef bleaching has also been ascribed to increased ultra-violet radiation together with temperatures greater than 30°C [Humann, 1993]. Chiappone et al. [1997], studying the coral reefs around the greater Bahamas banks, ascribe factors associated with the lack of reef development to include turbidity, sediment transport, and fluctuations in water temperatures.

[8] Figure 3 is a chart of the location of study. The locations of Andros Island, the Great Bahamas Banks, the
TOTO (Tongue of the Ocean) etc. can be seen relative to the state of Florida in the Atlantic Ocean. Figure 4 is a MODIS ocean color image of the Bahamas. For coral sand areas as much as 50% of the light striking the bottom is reflected, while for thick grass regions only about 5% is reflected [Mobley, 1994], and up to 40% of the reflected energy exits the ocean surface, depending on the water column depth and bottom albedo. The regions of sea grass and coral sand bottom based on their reflectivity are marked in the figure (Figure 4).

Some investigations of coastal areas have been performed recently using a black bottom, which assumes that all of the light energy hitting the bottom is placed in the bottom layer as heat (Weisberg et al. personal communication). They have also used a 100% reflecting bottom and have shown [Weisberg, 1996; Weisberg et al., 2001] how the generation of buoyancy (by heating and fresh-water input from the rivers) induces a baroclinic circulation that modifies the wind-driven circulation along the West Florida Shelf. The true effect of bottom absorption lies somewhere between the extremes of “no bottom” and a “black bottom”.

This paper shows that neglecting the optical effects of a bottom can lead to large errors when predicting the heat budgets and thermal structure of shallow oceanic areas. The importance of shallow-bottom reflection on the under-water light field and the subsequent effect on interpretations of ocean-color remote sensing has already been demonstrated [Lee et al., 1998, 1999, 2001]. A similar approach to that used by Lee is adopted here for determining bottom-reflected irradiance and its contribution to shallow-water heating.

2. Methods

2.1. Inherent Optical Properties

Morel Case 1 waters are waters in which the phytoplankton concentration dominates variations in absorption and scattering. Absorption by chlorophyll and related pigments therefore plays a major role in determining the total absorption coefficient in such waters. Case 1 waters can vary from very clear (oligotrophic) waters to very turbid (eutrophic) waters, depending on the phytoplankton concentration. Prieur and Sathyendranath [1981] developed a pioneering bio-optical model for spectral absorption of Case 1 waters. The essence of the Prieur and Sathyendranath model is contained in a more recent and simpler variant given by Morel [1991]:

\[
\alpha(\lambda) = \{ a_w(\lambda) + 0.06a_c(\lambda)C^{0.65} \} \\
\times \{ 1 + 0.2 \exp(-0.014(\lambda - 440)) \}
\]
where, $a_w(\lambda)$ is the absorption coefficient of pure water and $a_c^*(\lambda)$ is the non-dimensional statistically derived chlorophyll-specific absorption coefficient. One of the limitations of this model is that the model assumes that the absorption by yellow matter co-varies with that due to phytoplankton. Note that the formulation of Smith and Baker [1981] must be used in this expression rather than the clearer values of Pope and Fry [1997]. Otherwise the 0.2-factor must be changed to compensate for the yellow matter inherent in the Smith and Baker [1981] clear-water values.

For more turbid waters with higher concentrations of gelbstoff and suspended sediments, the absorption and scattering are calculated based on field measurements. The total absorption coefficient can be expressed as a sum of the absorption coefficients of pure water, gelbstoff and phytoplankton pigments respectively:

$$a_t(\lambda) = a_w(\lambda) + a_g(\lambda) + a_f(\lambda)$$ (2)

Absorption values for pure water are taken from Pope and Fry [1997], whereas absorption for phytoplankton pigments and gelbstoff has been modeled with simple bio-optical models, which give a realistic simulation for a variety of waters. The phytoplankton pigment absorption coefficient is simulated using the empirical model,

$$a_f(\lambda) = a_0(\lambda) + a_1(\lambda) \ln[a_f(440)]$$ (3)

where values for $a_0(\lambda)$ and $a_1(\lambda)$ are provided by Lee et al. [1998, 1999]. $a_f(440)$ can be input by itself or can be linked to another parameter such as the chlorophyll a concentration. To be consistent with the calculations of other researchers such as Morel [1988], and to compare with the POM, [chl-a] has been used as an input to determine the $a_f(440)$ and the particle scattering ($b_p(550)$) values:

$$a_f(440) = 0.06[chl - a]^{0.65}$$ (4)

$$b_p(\lambda) = B[chl - a]^{0.62}(550/\lambda)$$ (5)

Here B is an empirical value, which was 0.3 in the work of Gordon and Morel [1983]. It could vary among 0.3,1.0 and 5.0 to simulate a range from normal to highly turbid waters. It was set to 0.3 m$^{-1}$ for this study.

The gelbstoff absorption is expressed as [Bricaud et al., 1981; Carder et al., 1991]:

$$a_g(\lambda) = a_g(440) \exp[-0.014(\lambda - 440)]$$ (6)

Based on the input absorption and scattering, Hydro-light [Mobley, 1994] can be used to calculate the attenuation of light with depth for these waters. An appropriate expression for more rapidly determining the diffuse attenuation function can be derived. Kirk [1984] found an analytical relationship by systematically varying the optical properties which was comparable to a Monte Carlo study:

$$K_d(\text{avg}) = \frac{1}{\mu_0} \left[ a^2 + (0.425\mu_0 - 0.19ab) \right]^{1/2}$$ (7)

where, $K_d(\text{avg})$ is the average value of the attenuation coefficient in the euphotic zone, which is actually the value of $K_d$ at the midpoint of the euphotic zone. This equation has been further extended by Bissett et al. [1999] to incorporate the effects of the variation in the average cosine
with depth. In the present simulations, this equation for $K_{d}$ is adopted. $\mu_{d}$ is the cosine of a constant 45° in our simulations.

[16] For a listing of the symbols used in this study see Table 1.

[17] The average cosine used is the cosine of the subsurface solar zenith angle, and it provides the effective slant path of photons to depth for a shallow water column [Lee et al., 1999].

### 2.2. Quantifying the Optical Effects

[18] Lee et al. [1999, 2001] have modeled the effect of shallow waters to derive bottom depth and albedo using the in-water remote sensing reflectance $r_{s}$. They simulated the output from a large number of Hydrolight [Mobley, 1994] runs for a wide variety of bottom and water types. The Lee model is used to calculate the irradiance reflectance $r$ (Note that here, above surface value is designated as ‘R’ while below surface value by ‘r’).

$$
r = E_{u}/E_{d} = Q^{*}r_{s}; \text{ where } r_{s} = L_{u}/E_{d}. \quad (8)
$$

$E_{u}$ and $E_{d}$ are the upwelling and downwelling irradiances and $L_{u}$ is the upwelling radiance. While for shallow waters, $2.5 < Q < 3.5$, $Q$ is expressed as $\pi$ as a simplification and is used as such in our model. According to Voss et al. [2000] this assumption is approximately valid for shallow, low-absorbing sandy environments (e.g., 400–560 nm in the Bahamas). At longer wavelengths, $Q < \pi$, reaching values near 1.7 at 670 nm. It is assumed that the total reflectance at any depth is due to the sum of the reflectance due to the water column below, and the reflected light from the bottom [Lee et al., 1999]. Extending Lee’s equations to calculate $r$ (reflectance) provides

$$
r = r_{c} + r_{b}
$$

where $r_{c}$ is the contribution from water column and $r_{b}$ is the contribution from the bottom. The list of symbols is given in Table 1.

$$
rc = r_{dp}(1 - \exp\{-[\frac{1}{C0} + C]n(H - z)\})
$$

$$
r_{b} = \rho_{b}\exp\{-[\frac{1}{C0} + b]n(H - z)\}
$$

where $u = b_{b}/(a + b_{b})$, $\kappa = a + b_{b}$, $\rho_{b}$ is the bottom albedo

and

$$
C_{D_{a}} = 1.03(1 + 2.4u)^{0.5}
$$

$$
b_{D_{a}} = 1.04(1 + 5.4u)^{0.5}
$$

$H$ is the total water column height and $\cos \theta$ is the cosine of the subsurface solar zenith angle,

$$
r_{dp} = (0.084\pi + 0.17u)\frac{u}{2}
$$

[19] Once the reflectance ‘$r$’ is determined, the upwelling light field can be calculated using $E_{u} = r^{*}E_{d}$ [Mobley, 1994]. The net irradiance at any depth is $E_{net} = E_{d} - E_{u}$ and the $dE_{u}/dz$ is converted into heat [Kirk, 1988] and is added as a heat source for each water layer.

[20] The amount of light absorbed by the bottom per second is assumed to be radiated and conducted into the bottom layer and is determined by,

$$
Q_{bottom} = (1 - \rho_{b})E_{d}(\Delta H_{b})
$$

where $\Delta H_{b}$ is the depth of the bottom layer.

[21] A portion of the total upwelling light is internally reflected at the water-air interface, while the rest escapes back into the atmosphere. Gordon [1991] expressed this relationship and Lee et al. [1999] used it in a semi analytical model for nadir-viewing remote-sensing reflectance $R_{rs}$ using Hydrolight 3.0,

$$
R_{rs} = 0.5r_{rs} \frac{1}{1 - 1.5r_{rs}}
$$

### 2.3. Surface-Flux Calculations

[22] The heat fluxes have been parameterized by many scientists. In this paper the state-of-the-art formulae developed by various scientists and summarized by Doney [1996] have been adopted. $Q_{ad}$ is the short-wave radiation measured at the surface of the water column.

$$
Q_{sens} = \rho_{s}e_{s}C_{H}U_{10}(T_{a} - T_{s})
$$

$$
Q_{lat} = \rho_{s}L_{s}C_{W}U_{10}(q_{a} - q_{s})
$$

$$
Q_{lw (net)} = Q_{a} - Q_{b}
$$

$$
= -e_{g}\sigma[T_{s}^{4}(0.39 - 0.05e_{s}^{0.5})F(CI) + 4T_{s}^{3}(T_{s} - T_{a})]
$$

$$
F(CI) = 1 - 0.63CI
$$

Table 1. List of Symbols

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>$E_{u}$, $E_{d}$, $E_{out}$</td>
<td>Downwelling, upwelling and net radiation</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
<td>$a_{d}($λ$)$</td>
<td>Absorption due to water</td>
<td>m$^{-1}$</td>
</tr>
<tr>
<td>$a_{d}($λ$)$</td>
<td>Absorption due to gelbstoff</td>
<td>m$^{-1}$</td>
</tr>
<tr>
<td>$a_{d}($λ$)$</td>
<td>Absorption due to phytoplankton</td>
<td>m$^{-1}$</td>
</tr>
<tr>
<td>$chl-a$</td>
<td>Chlorophyll a</td>
<td>mgm$^{-3}$</td>
</tr>
<tr>
<td>$\mu_{d}$</td>
<td>Average cosine</td>
<td>-</td>
</tr>
<tr>
<td>$r$</td>
<td>Reflectance</td>
<td>-</td>
</tr>
<tr>
<td>$r_{c}$</td>
<td>Reflection from the water column</td>
<td>-</td>
</tr>
<tr>
<td>$r_{b}$</td>
<td>Reflection from the bottom</td>
<td>-</td>
</tr>
<tr>
<td>$r_{dp}$</td>
<td>Deep water reflectance</td>
<td>-</td>
</tr>
<tr>
<td>$r_{s}$</td>
<td>Remote sensing reflectance</td>
<td>-</td>
</tr>
<tr>
<td>$Q_{bot}$</td>
<td>Heat generated due the bottom</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
<td>$Q_{lost}$</td>
<td>Heat lost from the water column</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
<td>$Q_{u}$</td>
<td>Downwelling longwave radiation</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
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<td>Upwelling longwave radiation</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
<td>$Q_{a}$</td>
<td>Latent heat flux</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
<td>$Q_{bt}$</td>
<td>Sensible heat flux</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
<td>$Q_{u}$</td>
<td>Downwelling shortwave radiation</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
<td>$R_{B}$</td>
<td>Bowen’s ratio</td>
<td>-</td>
</tr>
<tr>
<td>$T_{a}$</td>
<td>Air temperature</td>
<td>°K</td>
</tr>
<tr>
<td>$T_{w}$</td>
<td>Water temperature</td>
<td>°K</td>
</tr>
<tr>
<td>$e_{s}$</td>
<td>Saturated vapor pressure</td>
<td>mbars</td>
</tr>
<tr>
<td>$e_{a}$</td>
<td>Vapor pressure of air</td>
<td>mbars</td>
</tr>
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<tr>
<td>$T_{a}$</td>
<td>Air temperature</td>
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</tr>
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<tr>
<td>$e_{a}$</td>
<td>Vapor pressure of air</td>
<td>mbars</td>
</tr>
</tbody>
</table>
where \( \rho_a \) is the air-density (1.22 kg m\(^{-3}\)), \( c_{pg} \) is the specific heat of air at constant pressure (1003 J kg\(^{-1}\) K\(^{-1}\)), \( C_H \) and \( C_E \) are bulk transfer coefficients (9.7 \( \times \) 10\(^{-4}\) and 1.5 \( \times \) 10\(^{-3}\), both unitless), \( U_{10} \) is the wind speed at 10 m above the sea surface (ms\(^{-1}\)) \( T_a \) and \( T_s \) are air and sea surface temperatures in K respectively, \( L_v \) is the latent heat of vaporization (2.45 \( \times \) 10\(^6\) J kg\(^{-1}\)), \( q_a \) and \( q_s \) are air and sea surface humidity respectively (both unitless), \( \varepsilon_o \) is emissivity of sea surface (0.985, untiless), \( \sigma \) is Stephan Boltzmann constant (5.7 \( \times \) 10\(^{-8}\) Wm\(^{-2}\) K\(^{-4}\)), \( e_a \) is the vapor pressure, \( C_l \) is the cloud index.

[23] The net rate of heat uptake by the surface layer, resulting from the combined effects of solar energy absorption and these various surface heat exchange processes is

\[
Q_{\text{net}} = Q_{\text{solar}}(j) + Q_a - Q_b - Q_c - Q_e + Q_{\text{bottom}}(1)
\]

where \( Q_{\text{solar}}(j) \) is the short-wave flux (from 400-700 nm) at the jth level of depth obtained from the downwelling light field and \( (Q_a - Q_b = Q_{\text{sw}}) \) is the remaining long wave radiation. \( Q_{\text{bottom}}(1) \) is a new extra term that we are emphasizing in this paper. This \( Q_{\text{bottom}}(1) \) is the sum of the upwelling light flux from the bottom that is absorbed by the jth layer and the heat diffused upwards from the bottom (heat that is radiated by the bottom), in this case, the top layer.

[24] The total energy that is lost from the water column as light from the surface is designated as \( Q_{\text{bottom}} \) in the Table 2 and Table 3. These calculations have been performed by the 1D GOTM. A typical curve showing the diurnal variation of the short-wave radiation is given in Figure 5. The background radiation value just taking into account nocturnal infrared radiation is about 30 Wm\(^{-2}\). The rate of heat uptake by a typical 2.5 m water column is given in Table 2. This experiment was carried out for 12 h of diurnal heating. The \( Q_{\text{bottom}} \) term is the term that is ignored by many modelers. This is the heat that is lost out of the water column. Mostly, the loss is through the surface as light (which the remote sensing instruments measure as the water-leaving radiance, \( L_w \)). In case of a transparent bottom (when the bottom is ignored), there is a very large heat loss of about 218 Wm\(^{-2}\), showing that ignoring the bottom creates quite erroneous results. Also the water column is about 0.2°C warmer after 12 h of heating when the bottom is black as opposed to coral sand but only 0.03°C warmer than when the bottom is grass. A similar run is carried out for a 10 m water column (seen in Table 3). The \( Q_{\text{bottom}} \) term for 10 m water depth (70 Wm\(^{-2}\)) is not as high as for the 2.5 m depth since more light is absorbed by the deeper water column.

### 2.4. Model Details

[25] We have already mentioned how the geographical features of the shallow waters for our study area make it relatively stagnant and isolated with minimum advection. For a one-dimensional analysis of the development of the thermal structure, like that used in Tables 2, 3, and 4, a 1-D turbulence model called the General Ocean Turbulence Model (GOTM) was adopted [Burchard et al., 1998; Burchard and Petersen, 1999].

[26] The \( k-\varepsilon \) and k-L models are the only two two-equation models that have been extensively applied to geophysical flows. In our simulations, the model was run with the Mellor-Yamada, k-kL and not the k-\( \varepsilon \) turbulence closure, since most ocean models follow MY (k-kL) scheme for calculating the turbulent kinetic energy. There are a total of 20 sigma levels in the vertical.

[27] This 1D GOTM was then incorporated or coupled with a 3D model POM (Princeton Ocean Model) to include the advective effects. We will hereafter call the modified model 3DGOTM. 3DGOTM was modified by including the new features of the light model as described in sections 2.1 and 2.2. The light model used here is an extension of the bottom-reflection optical model developed by Lee et al. [1999] for finer estimation of bottom depths. We extended the model to improve the heat budget calculations in general circulation models, which ordinarily use a simplified Jerlov classification of water types. The three new features added to 3DGOTM are

1) Heat input to the water column from the bottom through conduction (the most important term, which depends on the light reaching the bottom and the bottom albedo. This heat then mixes upwards through vertical diffusion and convection),
2) spectral calculation of \( K_d \)

### Table 2. Flux Rates Averaged Over a 12 h Heating Cycle During Model Simulation for a 2.5 m Deep Water Column (in Wm\(^{-2}\))^a

<table>
<thead>
<tr>
<th>Bottom Type</th>
<th>( Q_a (+) )</th>
<th>( Q_b (-) )</th>
<th>( Q_c (-) )</th>
<th>( Q_e (-) )</th>
<th>( Q_{\text{solar}} (+) )</th>
<th>( Q_{\text{bottom}} (-) )</th>
<th>( Q_{\text{net}} (+) )</th>
<th>( \Delta \text{SST in 12 h} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Black</td>
<td>365.07</td>
<td>440.18</td>
<td>75.02</td>
<td>14.49</td>
<td>375.10</td>
<td>10.3</td>
<td>200.18</td>
<td>0.83</td>
</tr>
<tr>
<td>Sea grass</td>
<td>365.07</td>
<td>440.11</td>
<td>74.97</td>
<td>14.43</td>
<td>375.10</td>
<td>12.00</td>
<td>198.66</td>
<td>0.80</td>
</tr>
<tr>
<td>Coral sand</td>
<td>365.07</td>
<td>439.59</td>
<td>74.59</td>
<td>13.97</td>
<td>375.10</td>
<td>40.8</td>
<td>171.22</td>
<td>0.61</td>
</tr>
<tr>
<td>Clear</td>
<td>365.07</td>
<td>437.76</td>
<td>73.28</td>
<td>12.35</td>
<td>375.10</td>
<td>218.86</td>
<td>-2.08</td>
<td>-0.04</td>
</tr>
</tbody>
</table>

*a"Clear" represents a transparent bottom. Wind speed is the actual measured value.*
(downwelling attenuation coefficient) using 21 wave bands in the visible range. 3) An additional upwelling radiation term $E_u$ from the bottom. This is the fraction of the incoming light that got reflected off the bottom.

Due to solar (short-wave+ long-wave) heating, the water column develops a well mixed temperature profile in these shallow regions [Kirk, 1988]. The resulting temperatures attained were found to depend on the water depth and bottom albedo.

3. Results and Discussion

3.1. Effect of Bottom Albedo

Sensitivity to bottom albedo was one of the main interests of the simulations. The spectral albedos for the two types of bottom (sea grass and coral sand) are provided in Figure 6. Simulations with black bottoms produce water columns that are warmer than using the other bottom properties. Coral sand is highly reflective (albedo of about 0.5) while sea grass is not (about 0.05). The bottom serves two purposes: it reflects the short-wave radiation incident on it depending on its albedo, and absorbs the rest of the radiation. This absorbed radiation is then conducted and radiated back as long-wave radiation and immediately absorbed in the bottom layer. Simulations were run with various bottom conditions in the 1Dimensional GOTM for 1 m, 2.5 m, 5 m, 7.5 m, 10 m, and 20 m deep water columns for 12 h of solar heating. The results for the various types of bottom are tabulated in Table 4. Note that the effects observed in the simulation were accentuated for a shallow (1 m) water column with an almost 0.9°C temperature difference between the black and coral-sand bottom cases. Ignoring the bottom in heat budget models results in water temperature some 2.5°C cooler than one would expect with a grass bottom. In 20 m of water, however, the difference becomes negligible over a single heating cycle.

Figure 7 shows a simulation for a 7.5 m (using 1D GOTM) water column, with the winds set to a very low value of 0.5 m/s to facilitate the comparison among various bottom types for a stratified water column. These shallow regions completely mixed when the winds exceeded 1 m/s for this dry (70% RH) spring period. For this 7-day simulation, the starting temperature profile at 6 am was set constant with depth at 20.5°C. The water column temperature was then allowed to develop for a week reproducing temperature profiles at the end of the seventh day.

Table 4. Final Temperatures (°C) Obtained by Model Simulation After 12 h of Heating

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Black (°C)</th>
<th>Sea Grass (°C)</th>
<th>Coral Sand (°C)</th>
<th>Transparent Bottom (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.0</td>
<td>27.90</td>
<td>27.80</td>
<td>27.04</td>
<td>25.32</td>
</tr>
<tr>
<td>2.5</td>
<td>27.32</td>
<td>27.30</td>
<td>27.11</td>
<td>26.46</td>
</tr>
<tr>
<td>5.0</td>
<td>27.00</td>
<td>26.99</td>
<td>26.92</td>
<td>26.67</td>
</tr>
<tr>
<td>7.5</td>
<td>26.86</td>
<td>26.86</td>
<td>26.83</td>
<td>26.65</td>
</tr>
<tr>
<td>10.0</td>
<td>26.79</td>
<td>26.78</td>
<td>26.77</td>
<td>26.66</td>
</tr>
</tbody>
</table>

The solar radiation values and fluxes are as before. The starting temperature at 6:00 am in the morning of simulation was 26.5°C.

Figure 6. Spectral albedo curves for sea grass and coral sand bottoms.
day. Most light hitting a sea-grass bottom for example is absorbed and radiated as long-wave radiation or conducted into the bottom of the water column. Convection then mixes this hot water upwards due to buoyancy effects producing a well-mixed water column. The bottom of the water column was heated more for the sea grass than for coral sand, and less light was reflected upwards. At the end of a week, the sea grass produced a warmer water column than did the coral sand bottom. This is because more light is lost by bottom reflection through the air-sea interface when the bottom is coral sand (e.g., $Q_{\text{loss}}$ in Table 2).

Additional 3 Dimensional simulations were made using 3DGOTM with bathymetry contours for the study area as shown in Figure 8. This is the same location as in Figure 1, to the north west of Andros Island. There are regions of extremely shallow depth (note that the regions below 1 m depth are treated as 1 m by the model). The model domain has been set up with 4 km bins in the x and y directions. There are 50 grids in the x and 60 grids in the y direction. Figure 9 shows a 3D simulation (using 3DGOTM) for conditions measured from 15 to 17 April 2001 for comparison with the AVHRR images shown in Figures 1a and 1b. The simulation was for 60 h starting at 0600, 15 April with an initial uniform 26.5°C water column with a coral sand bottom. The final sea-surface temperatures, for the coral-sand bottom simulation are provided in Figure 9.

Results from similar runs are shown in Figure 10a (the temperature anomaly) where coral sand is replaced by sea grass. Figure 10b shows the simulated temperature anomalies when the bottom is completely black. As can be seen from Figure 10b, there is a very appreciable change in the horizontal temperature structure after only 60 h when results from a black bottom are compared to those with coral-sand bottom. Note that the shallow-water temperatures can be 1.55°C warmer for black rather than for coral sand bottoms.

Figure 9 can be compared with the bathymetry map (Figure 8) to locate the shallow and deep areas. The regions where the bathymetry falls off to very deep regions are found to be regions of sharp temperature fronts when lateral convection and mixing are neglected. A temperature contrast of about 1.7°C (warmer at the shallow end) can be observed at the interface of these thermal fronts. Note that high temperatures of up to 27.7°C in Figure 9 are in the smaller, very shallow regions with depths less than 1.5 m. The black-bottom temperatures are modeled as high as 29.2°C in Figure 10b. As a reality check, Figure 1b shows AVHRR temperatures exceeding 28.5°C, falling between sea-grass and coral-sand temperatures.

Note that, much of the shallow area near Andros Island and the Berry Islands is, actually less than 1 m in depth, which was modeled as 1 m deep. Also, the skin temperature viewed by AVHRR on calm days can be 0.5 to 1°C higher than the bulk temperature [Brown and Minnett, 1999]. However, vast areas with depths of about 3–7.5 m can be seen to be at about 27°C in both Figure 9 and Figure 1b.
3.2. Effect on Salinity

From the above discussions, the effect of bottom albedo on heat budget calculations and hence on the temperature of the water column has been demonstrated. In addition to this direct effect on water column temperature, the heat budget also has an indirect effect on salinity. As the water evaporates, salinity increases, as does density. In this section, the effects of net evaporation on salinity changes are revisited during both spring and summer.

A comparison between transparent and sea-grass bottoms shows the strong influence of bottom absorption and reflection on the salinity and corresponding density. As expected, salinity changes are greater for sea grass bottom, since thermal heating is greater with a sea grass bottom and hence so is net evaporation. There is an almost 0.4 psu difference between the transparent-bottom model and the implementation of the sea grass bottom over a period of 24 days. This could lead to serious errors when we try to simulate circulation over coastal areas. The salinity change represents then, a maximal salinity change possible for each depth. It is seen that the salinity for these water columns can reach about 46 psu in late spring, and is corroborated from observations [Smith, 1940; Cloud, 1962]. The roughly concentric arrangement of the isohalines was observed in the pool of hyper-saline waters, which develops in the summer months off the west coast of Andros (recall Figure 2). As observed from the bathymetry (Figure 8) these are the shallowest isolated regions along Andros Island with depths less than 2 m.

These pools of hyper saline waters formed in spring and summer are never completely destroyed by the winds in winter and they escape dilution by waters of lower salinity due to their isolation from the ocean. Broecker [1966] estimated residence times up to 300 days for this hyper-saline region west of Andros Island. The presence of these hyper-saline plumes extends to winter, as seen from sampled data, though the salinity values are much lower (about 39 ppt). The reasons for the persistence of this salinity pattern are primarily due to the moderate tidal exchange with deeper waters and the shelter from the full vigor of the

Figure 9. 3DGOTM modeled SST (°C) with uniform coral sand bottom after a 60 h April simulation of the Bahamas banks.
Figure 10. SST anomalies (°C) after 60 h when a sand bottom is replaced by (a) sea grass or (b) black surface.
trades provided by Andros Island. So, even though there is a slow intrusion of less saline water from the Tongue Of The Ocean, it is clear that the dilution is not enough to offset the effects of evaporation on salinity.

4. Conclusions

This paper emphasizes the errors that can be expected when running an Ocean Model without inclusion of proper bottom absorption and reflection of short wave radiation in certain locations of the world (when some conditions as described before prevails). The sea surface temperature in shallow areas can be severely underestimated if the bottom is ignored and is overestimated if it is assumed to be black.

There is a significant fraction of the light energy reaching the bottom in these shallow waters. This fraction can decrease to some extent in case of waters with very high turbidity or high chlorophyll content. The current model takes into consideration the effect of variable chlorophyll conditions while the original POM and 3DGOTM model do not. A fraction of the light hitting the bottom is diffusely reflected back depending on the bottom albedo (high for coral sand and low for sea grass and algal mats). The rest of the light gets absorbed by the bottom and is reemitted as heat convectively heating up the water column from below. Light reflected through air-sea interface is considered and does not heat the water column.

It can be safely concluded that the surface temperature in the shallow areas is wrongly predicted if the bottom effects are neglected or inaccurately represented for waters shallower than 20 m. In any simulations over a period of a couple of days, the errors can be expected to be appreciable if the bottom absorption and radiation are not properly quantified and accounted for. This, in turn, will have effects on the salinity by evaporation and hence density variations of the water column, thus changing the hydrodynamics as well. Thus it can be said with certainty that shallow-water effects are important when dealing with the heat budget of shallow estuaries and lagoons where the depths average less than 10 m. These heating effects and the corresponding salinity increase are expected to have strong implications to the existence of the biological ecosystems at that location. It is known, that these salinity excursions can cause damage to the coral reefs at that depth [Lang et al., 1988]. The damage to coral reefs is not an isolated phenomenon, for the entire bio-diversity in such locations is affected.

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References


