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Ocean Radiant Heating

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by

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Abstract

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Global climate is largely influenced by upper ocean evolution which, in turn, is extremely sensitive to the vertical distribution of the solar energy available for ocean radiant heating. Here, the transmission of solar radiation through the water column, it's variations and their effects are examined on a variety of scales. A global model is developed and used with climatological data to determine the importance of solar penetration through the mixed layer on seasonal scales. In-situ data from the western equatorial Pacific are used to address variations in solar transmission on daily to synoptic scales, to find factors which explain these variations, and to determine the influence of these variations on sea-surface temperature and mixed layer depth. Finally, a full spectral in-water radiative transfer model is used to investigate solar transmission in the upper few meters of the ocean, with particular emphasis on the near-infrared wavebands.

Results suggest that solar radiation penetrating the mixed layer can be a significant term (20 W m\(^{-2}\)) in the mixed layer heat budget for tropical regions. In mid and high latitude regions, solar penetration is important on seasonal timescales, as annual cycles in incident solar irradiance, upper ocean chlorophyll concentration, and mixed layer depth cause trapping of penetrating solar energy of O(10 W m\(^{-2}\)) within the seasonal pycnocline. Variations in solar transmission on mixed layer depth scales can be explained primarily by upper ocean chlorophyll concentration, and to second order by cloud amount. A one-dimensional mixed layer model gives SST errors near 1.0°C when forced with a typical Jerlov water type solar transmission profile. Net irradiance at depth, within the upper 5 m, can vary by 37 W m\(^{-2}\) (based on a climatological surface irradiance of 220 W m\(^{-2}\)), due to changes in chlorophyll biomass concentration, clouds, and solar zenith angle. A strawman parameterization which relates in-water and surface irradiance values through these parameters is developed. The parameterization improves the accuracy of irradiance at depth by more than 10 W m\(^{-2}\) when compared to existing solar transmission models which are completely invariant or depend only on chlorophyll concentration.
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Overview

1. Introduction

Air-sea interaction processes must be better understood if the predictive ability of the Earth’s climate system is to be improved. General circulation models (GCMs) rely on proper thermal forcing of the upper ocean and subsequent heat exchange with the atmosphere. Solar radiation is the primary heat source for the upper ocean, and is a unique term in the heat equation. Unlike other terms which act solely at boundaries, solar radiation penetrates the air-sea interface, and through vertical divergence, directly heats water beneath the ocean surface. For upper ocean modeling it is insufficient to consider only the incident solar irradiance and to assume the thermal energy is instantaneously uniformly distributed throughout the mixed layer. It is necessary to consider the flux divergence, or solar transmission, within the water column, and the possibility that some fraction of the incident irradiance penetrates beyond the mixed layer base, becoming lost to further direct air-sea exchange processes. This dissertation is concerned with the transmission of solar radiation through the water column over a variety of time and space scales, its variations, and its parameterization. The primary goals of this dissertation are:

• To provide an indication of the ocean regions where the fluxes of solar radiation which pass beyond the mixed layer base are a significant portion of the incident irradiance or the net-air sea heat exchange.

• To determine the role of various physical and biological processes in regulating the transmission of solar radiation through the water column on mixed layer depth scales.

• To examine solar transmission and its regulating factors on centimeter to meter depth scales with particular emphasis on the role of sea-surface albedo.

• To explore the sensitivity of upper ocean evolution to changes in solar transmission.

This study provides a climatology of solar flux values at the mixed layer base, the first such global data set, identifies solar transmission variations and their causes, and quantifies how variations in solar transmission influence mixed layer temperature and depth. Most importantly, the results of this work will play a key role in developing
improved solar transmission parameterizations which must abandon Jerlov water type and rely upon continuous measurable quantities such as upper ocean chlorophyll concentration, cloud amount, and solar zenith angle.

Each chapter in this thesis has been written to stand independently of the others. When placed together, the three chapters provide a comprehensive study of in-water solar transmission, or ocean radiant heating.

Chapter 1

Models describing the transmission of solar radiation through the water column date back nearly 100 years (e.g. Schmidt 1908, Defant 1961, Kraus and Turner 1967, Jerlov 1976). However, the importance of solar radiation which penetrates the mixed layer base to the upper ocean heat balance and air-sea heat exchange was only recently identified (Woods et al. 1984, Lewis et al. 1990). Penetrating solar fluxes have been quantified for only a small portion of the world’s oceans, although existing climatologies enable this calculation to be made for all ocean regions.

To calculate climatological values of solar penetration a model which parameterizes in-water solar fluxes in terms of available parameters is developed. The model is constructed for use with existing data as well as with forthcoming remotely sensed products. The model begins with incident irradiance values determined from cloud reflectance. A radiative transfer model is then used to break up the total irradiance spectrally. Spectral values are propagated to the mixed layer depth which is computed from temperature and salinity profiles given in a climatological atlas. Spectral propagation is based upon the diffuse attenuation coefficient spectrum which is parameterized in terms of mixed layer chlorophyll concentration.

Model results indicate that solar penetration can be a significant term (20 W m$^{-2}$) in the mixed layer heat budget in tropical regions. In the mid- and high-latitudes, little solar radiation passes the surface of the permanent thermocline. However, solar penetration beyond the seasonal mixed layer depth is important, as thermal energy can become trapped within the seasonal pycnocline for months. A coupling between solar penetration and mixed layer depth exists. Results of this global study point toward the tropical regions for a more detailed study of solar penetration on shorter timescales.
Chapter 2

In-situ optical data sets are generally composed of only a few profiles widely interspersed in both time and space (e.g., Paulson and Simpson 1977). To examine short term variations in solar transmission and to determine physical and biological factors which regulate in-water solar fluxes a more cohesive data set is required. For this study specifically, more than 750 daytime profiles of spectral irradiance were made in the western equatorial Pacific over a 30 day period. Additionally, a suite of physical, optical, and biological parameters were coincidentally sampled. This provides a data set which can be used to test existing in-water solar transmission parameterizations, to develop an empirical regional parameterization, and to determine the role of various physical and biological processes in regulating transmission on mixed layer depth scales ($z > 10$ m).

The mean penetrative solar flux value for the western equatorial Pacific determined from the in-situ data set is 20 W m$^{-2}$ (based on a climatological incident flux of 220 W m$^{-2}$ and a 30 m mixed layer depth). This value is similar to climatological penetrative solar flux values determined in Chapter 1 for the region. However, solar fluxes at 30 m range from 7 to 44 W m$^{-2}$ (based on an incident flux of 220 W m$^{-2}$). When in-situ solar flux values are compared with solar flux values parameterized with existing solar transmission parameterizations which rely on Jerlov water type indices, errors of O(10 W m$^{-2}$) often arise.

This data set is used to develop an empirical model for solar transmission in the form of a single exponential. Model parameters are calculated by day and compared (statistically) with presumably related parameters. Results indicate that variations in solar transmission are well explained ($r^2=0.83$) by the mixed layer chlorophyll concentration which influences the attenuation of irradiance. Clouds play a secondary role by altering the spectral shape of the incident irradiance. As cloud amount increases, a relatively greater portion of the irradiance exists in the blue-green spectral region for which the ocean is most transparent. To show the sensitivity of the upper ocean evolution to solar transmission a one-dimensional mixed layer model is used to simulate the sampling period. Compared to the use of solar transmission profiles determined empirically by day, use of a Jerlov water type based parameterization results
in mixed layer temperature differences which reach 0.83 C and mixed layer depth differences as large as 3 m.

Chapter 3

The in-situ optical data used in this study covers the visible and near-ultraviolet wavebands, and thus completely resolves solar fluxes only beneath ~10 meters. For depths shallower, a greater portion of the solar spectrum must be resolved. Both full spectral measurements, and measurements within the upper few meters are difficult to accurately obtain. Atmospheric and oceanic radiative transfer models are relied upon to extend this study to the near-surface layer. Such a modeling effort enables solar transmission to be assessed on shallow mixed layer (O(m)) scales, warm layer (O(cm)) scales, and directly at the air-sea interface (albedo).

To model in-water solar fluxes the Santa Barbara DISORT Atmospheric Radiative Transfer model is run for a variety of solar zenith angles and cloud indices to generate spectral radiance distributions just above the sea-surface. This radiance distribution is then used as input to the HYDROLIGHT in-water radiative transfer model which includes a wind speed dependent air-sea interface. HYDROLIGHT is run for each of the surface radiance distributions, various sea-surface roughnesses (wind speed), and upper ocean chlorophyll concentrations.

Model results indicate that in-water solar fluxes vary by as much as 40 W m\(^{-2}\) at depth, within the top few meters of the ocean (based on a climatological incident irradiance of 220 W m\(^{-2}\)). Variations in near surface solar transmission are due primarily to upper ocean chlorophyll concentration, cloud cover, and solar zenith angle. Chlorophyll concentration significantly affects solar transmission only below 1 m. An order of magnitude increase in chlorophyll concentration (from 0.03 to 0.30 mg m\(^{-3}\)) gives only a 2% increase in the radiant heating rate of the top meter of the ocean. Clouds increase solar transmission in the upper few meters, and alter the mean cosine of the light field. Under clear skies, a 40% reduction in the incident irradiance due to clouds results in more than a 45% decrease in the radiant heating rate for the upper meter. The influence of solar zenith angle is greatest under clear skies when a change from 10 to 60° gives a near 5% reduction in the near surface radiant heating rate.
The set of modeled solar transmission values also enables the role of sea surface albedo to be examined and provides sufficient qualitative information from which a strawman parameterization can be developed. This parameterization expresses in-water solar fluxes in terms of the surface irradiance, solar zenith angle, cloud amount, and upper ocean chlorophyll concentration. Even in its preliminary state, this is an improvement over existing near-surface parameterizations which are generally invariant or depend only on Jerlov water type.

Synthesis

This work shows the sensitivity of upper ocean models to ocean radiant heating rates and indicates the importance of accurately parameterizing solar transmission on various temporal scales. More significantly, it lays the foundation for development of the next generation of in-water solar transmission models by identifying causes of variation in solar transmission, and quantifying dependencies of solar transmission on upper ocean chlorophyll concentration, cloud amount and solar zenith angle. A thorough analysis of ocean radiant heating is now complete. An all-condition in-water solar transmission and ocean radiant heating parameterization which relies upon easily measured quantities is presently being developed. This parameterization will enhance both ocean and atmospheric modeling efforts, ultimately leading to an improved predictive understanding of climate variations.

References
Chapter #1

Ocean Mixed Layer Radiant Heating and Solar Penetration: A Global Analysis

Abstract

A hybrid parameterization for the determination of in-water solar fluxes is developed and applied to compute the flux of solar radiation which penetrates beyond the upper ocean mixed layer into permanent pycnocline waters on global space and climatological time scales. The net flux of solar radiation at depth is modeled using values of the solar flux incident at the sea surface, derived from the International Satellite Cloud Climatology Project data set, and in-water attenuation coefficients, determined using upper ocean chlorophyll concentration supplied by Coastal Zone Color Scanner imagery. Solar radiation penetration can be a significant term (20 W m$^{-2}$) in the mixed layer heat budget for tropical regions. In mid and high latitude regions, the annual solar flux entering permanent pycnocline waters is small (< 5 W m$^{-2}$). However, solar penetration in these regions is important on seasonal timescales, as annual cycles in incident solar flux, upper ocean chlorophyll concentration, and mixed layer depth cause trapping of penetrating solar energy of O(10 W m$^{-2}$) within the seasonal pycnocline. This trapped thermal energy is unavailable for atmospheric exchange until winter, a period as long as nine months. A nondimensional parameter is introduced which quantifies the fraction of incident solar radiation contributing to mixed layer radiant heating. This parameter can be used to characterize the relative importance of solar penetration to ocean mixed layer thermal climate.
1. Introduction

Quantitative knowledge of momentum, heat, and buoyancy fluxes at the air-sea interface is necessary for proper forcing of upper ocean models. Air-sea interaction processes generally act upon the sea-surface itself, applying a shear stress, cooling, or altering salinity. Fluxes of solar radiation differ. Solar radiation passes beyond the air-sea interface, and through its vertical divergence can directly heat well below the ocean surface. Solar energy can penetrate beyond the base of the oceanic mixed layer, becoming lost to immediate mixed layer heating and further sea-air exchanges. The quantification of solar radiation penetrating the mixed layer base is critical to the accurate determination of mixed layer heating rates. Lewis et al., (1990) calculated the climatological mean flux of solar radiation penetrating the mixed layer in the equatorial Pacific to be as large as 50 W m\(^{-2}\). Compared to the case of complete solar absorption, this penetrative flux results in a reduction in the heating rate of a 20 m mixed layer of more than 1.5 °C month\(^{-1}\). The penetrative component of solar radiation can vary significantly on synoptic to seasonal time scales due to planetary solar cycles, mixed layer evolution, and changes in radiation attenuation due to variations in algal pigment concentrations. For example, a typical change in mixed layer phytoplankton pigment concentration of 0.10 mg m\(^{-3}\) results in a corresponding change in the penetrative solar flux at 20 m of order 10 W m\(^{-2}\) (Lewis 1987; Lewis et al., 1990; Siegel et al., 1995).

The attenuation of downwelling solar radiation with depth in the upper ocean is generally described by the sum of vertically decaying exponential functions. The ocean mixed layer model of Kraus and Turner (1967) employs a single exponential function while more recent parameterizations address particular spectral regions, each with differing attenuation lengths (Kraus, 1972; Paulson and Simpson, 1977; Zaneveld and Spinrad, 1980; Woods et al., 1984; Morel, 1988; Morel and Antoine, 1994). The number of spectral regions considered by these models ranges from two to more than 100. The evolution of upper ocean structure from mixed layer model results has been found to be extremely sensitive to e-folding lengths used in the solar attenuation parameterization (e.g. Kraus and Turner, 1967; Denman 1973; Simpson and Dickey, 1981; Woods et al., 1984; Price et al. 1986). Attenuation coefficients are functions of the light attenuating materials within the water column and thereby (for climatological timescales) location. A recent global study of solar transmission characteristics uses
data (mostly Secchi disk depth observations) dating back to the 1940's to determine solar extinction coefficients (Simonot and LeTreut 1986). These extinction coefficients are based on Jerlov water type, a subjective index used to parameterize solar transmittance, and a single exponential decay function (Jerlov, 1976). The importance of solar penetration to upper ocean modeling along with the existence of new climatologies suggests that global estimates of solar radiation penetration should be made.

To assess the climatological importance of solar penetration, the seasonal evolution of upper ocean structure must be evaluated. For example, thermal energy associated with solar radiation absorbed within the near-surface mixed layer is immediately available for exchange with the atmosphere. However, thermal energy within the seasonal pycnocline (associated with solar penetration beyond the seasonal mixed layer) will be lost to atmospheric exchange until entrainment associated with winter mixing occurs. This coupling of seasonal pycnocline dynamics and solar radiation penetration effectively puts a seasonal time lag on the availability of thermal energy associated with seasonal penetration for atmospheric exchange. On climatological timescales, the flux of solar radiation penetrating beyond the permanent pycnocline is the relevant quantity. The solar flux penetrating the maximum monthly mixed layer depth is lost to further local atmospheric exchange on annual timescales and is the primary consideration of this work.

Here, a hybrid parameterization of the net flux of solar radiation at depth is developed and applied to examine global fluxes of solar radiation penetration on climatological timescales. The solar spectrum is divided into two regions, the near-infrared which is quickly attenuated, and the visible which penetrates well into the water column. Solar energy in the near-infrared is assumed to be converted to thermal energy within the upper few meters and cannot contribute to the penetrating solar flux. Solar energy in the visible region is treated spectrally, with attenuation coefficients determined from remotely sensed upper ocean chlorophyll concentrations. The climatological distribution of solar radiation at the base of the deepest monthly climatological mixed layer for the world oceans is determined. This enables the identification of regions where solar penetration is a significant portion of the mixed layer heat budget. Further, a nondimensional parameter which relates heat lost to the
deepest monthly mixed layer in the form of solar penetration to the quantity of solar radiation available for upper ocean heating is introduced.

2. Mixed layer radiant heating and solar penetration

A one dimensional heat budget for a mixed layer of depth $D_{ml}$ may be expressed as

$$\frac{DT_{ml}}{Dt} = \frac{-Q_{loss}}{\rho \ c_p \ D_{ml}} + RHR_{ml}$$ (1)

where $T_{ml}$ is the mean temperature of the mixed layer, $D(\cdot)/Dt$ is the time rate of change evaluated following its motion ($\partial(\cdot)/\partial t + u\partial(\cdot)/\partial x + v\partial(\cdot)/\partial y$), $Q_{loss}$ is the sum of the net longwave, sensible, and latent heat losses at the sea surface, and turbulent fluxes at the mixed layer base, $\rho$ is the density of seawater, $c_p$ is the specific heat of seawater, and $RHR_{ml}$ is the radiant heating rate of the mixed layer. The rate at which solar radiation heats the upper ocean mixed layer is defined as

$$RHR_{ml} = \frac{\overline{E}_n(0^*) - \overline{E}_n(D_{ml})}{\rho \ c_p \ D_{ml}}$$ (2)

where $\overline{E}_n(0^*)$ is the net flux of solar radiation just beneath the sea surface, $\overline{E}_n(D_{ml})$ is the net solar flux at the depth of the mixed layer, and the overbars indicate spectral integration over the entire solar spectrum (250 - 2500 nm).

Irradiance fluxes are most accurately characterized as a function of wavelength (Morel and Prieur, 1977; Smith and Baker, 1978; Baker and Smith, 1982; Siegel and Dickey, 1987a; 1987b; Morel, 1988; Sathyendranath and Platt, 1988; Morel and Antoine, 1994; Siegel et al., 1995). Accordingly, the total net irradiance at depth, $\overline{E}_n(z)$, is determined by integrating the net spectral irradiance, $E_n(z, \lambda)$, over the entire solar spectrum, or

$$\overline{E}_n(z) = \int_{\lambda_{\text{ext}}}^{\lambda_{\text{sol}}} E_n(z, \lambda) \ d\lambda = \int_{\lambda_{\text{ext}}}^{\lambda_{\text{sol}}} (E_d(z, \lambda) - E_u(z, \lambda)) \ d\lambda$$ (3)

where $E_d(z, \lambda)$ and $E_u(z, \lambda)$ are respectively, the downwelling and upwelling spectral irradiance fluxes (in units of W m$^{-2}$ nm$^{-1}$), and $\lambda_{\text{sol}}$ represents the limits of integration for the solar spectrum (250 to 2500 nm).
Two spectral parameters are commonly employed to characterize the upper ocean light field: the diffuse attenuation coefficient spectrum, $K_d(\lambda)$, and the irradiance reflectance spectrum, $R(\lambda)$. The diffuse attenuation coefficient spectrum quantifies the rate at which spectral downwelling irradiance decreases with depth and is defined by the Beer-Lambert relation,

$$E_d(z,\lambda) = E_d(0^+,\lambda) \exp(-K_d(\lambda)z)$$

where $z$ is depth (positive) within the mixed layer. The reflectance spectrum is defined as the ratio of the upwelling to downwelling spectral irradiance, or $R(\lambda) = \frac{E_u(z,\lambda)}{E_d(z,\lambda)}$.

Using the definitions above, the mixed layer radiant heating rate, $RHR_{ml}$, may be evaluated as

$$RHR_{ml} = \frac{1}{\rho c_p D_{ml}} \int_{\lambda_{sol}} E_d(0^+,\lambda)(1-R(\lambda))(1-\exp(-K_d(\lambda)D_{ml}))d\lambda$$

Assuming energy conservation across the air-sea interface (i.e., $E_n(0^+,\lambda) = E_n(0^-,\lambda)$), values of $E_d(0^-,\lambda)$ can be determined from $E_d(0^+,\lambda)$ using:

$$E_d(0^-,\lambda) = \frac{E_d(0^+,\lambda)(1-\alpha)}{(1-R(\lambda))}$$

where $\alpha$ is the sea surface albedo defined as the ratio of upwelling to downwelling irradiance just above the sea surface. It is assumed here that $\alpha$ is not wavelength dependent as limited evidence of spectral albedo structure exists (Katsaros et al., 1985; King et al., 1990; Morel and Antoine, 1994).

Substitution for $E_d(0^-,\lambda)$ in equation (5) results in a final expression for the mixed layer radiant heating rate, and the net radiative flux at the mixed layer base, $E_n(D_{ml})$.

$$E_n(D_{ml}) = (1-\alpha) \int_{\lambda_{sol}} E_d(0^+,\lambda)(\exp(-K_d(\lambda)D_{ml}))d\lambda$$

The fraction of available solar radiation converted to thermal energy within the mixed layer provides a useful parameter for addressing the role of radiation penetration on upper ocean heating. A nondimensional parameter, $\Pi$, gives this fraction in terms of either solar fluxes or radiant heating rates.
\[ \Pi = 1 - \frac{\int E_d(0^+, \lambda) \exp(-K_d(\lambda) D_{ml}) \, d\lambda}{\bar{E}_d(0^+)} = \frac{RHR_{ml}}{RHR_{max}} \]  

with \( RHR_{max} = \bar{E}_d(0^+)(1-\alpha) / \rho c_p D_{ml} \). If all available solar radiation is converted to heat within the mixed layer (i.e. no solar penetration exists), \( \Pi \) is equal to one. In regions where the water is clear and the mixed layer depth is shallow, values of \( \Pi \) can be significantly less than one.

3. Data and methods

a. Surface irradiance spectrum

Incident solar flux values can be parameterized in terms of cloud amount and type, both available globally from remotely sensed data (e.g. Gautier et al., 1980; Bishop and Rossow, 1991). Here, data from the International Satellite Cloud Climatology Project (ISCCP) C1 data set and the radiative transfer model of Gautier et al. (1980) are used to determine climatological mean values of solar radiation over the world oceans (Rossow and Schiffer, 1991; Wang, 1993). Radiance measurements from various weather satellites, both geostationary and polar orbiting, were used to construct the ISCCP C1 data set on a 2.5 degree spatial scale and a 3 hour time scale (Rossow and Schiffer, 1991). Three parameters from the ISCCP C1 data set (precipitable water, cloud fraction, and cloud radiance) were used to determine 3 hour global values of shortwave radiation incident at the Earth's surface by modeling the effect of the atmosphere and clouds on the top of the atmosphere solar irradiance (Gautier et al., 1980; Wang, 1993). The resulting values of \( \bar{E}_d(0^+) \) were compared with incident shortwave values computed by applying the method of Breon et al. (1994) to data from the Earth Radiation Budget Experiment (ERBE). The results from both satellite data sets were in good overall agreement with a difference between the two data sets found in regions observed with a large viewing angle by the geostationary satellites (Wang, 1993). For large viewing angles, radiance measured at the satellite comes from both cloud tops and cloud sides. Therefore, cloud fraction derived by a threshold method, as done by ISCCP, will be overestimated. To correct for this overestimation, an ad-hoc scheme was developed to simulate the effects of three dimensional clouds on radiation (Bates and Gautier, 1989; Breon, 1992; Wang, 1993). Corrected incident solar values were
regridded onto 1 degree squares and both monthly and annual climatological fields of $\overline{E_d}(0^+)$ were calculated. Comparisons with direct observations and other remotely sensed data suggests that the incident flux data used here is accurate to order 10 W m$^{-2}$ (Wang, 1993).

Spectral decomposition of the ISCCP $\overline{E_d}(0^+)$ values is best accomplished through use of a spectral atmospheric radiative transfer model. The irradiance spectrum just above the sea surface is equal to a modeled spectrum re-scaled by the satellite-derived incident solar irradiance ($\overline{E_d}(0^+)$). Here, the Tanre (1979) 5S radiative transfer model is used to compute a model spectrum for a solar zenith angle of 23 degrees and a clear sky maritime atmosphere ($H_2O$ vapor = 4.1 g cm$^{-2}$, $O_3$ = 0.247 atm-cm. maritime aerosol distribution). The spectral shape is mostly invariant to changing atmospheric characteristics and solar zenith angle. Only when atmospheric water vapor was substantially altered did the spectral shape of $E_d(0^+\lambda)$ change noticeably. A 75% decrease in atmospheric water vapor resulted in roughly a 2% decrease in the amount of energy in the visible wavebands relative to the entire solar spectrum.

b. Albedo

Values of sea surface albedo are dependent on solar zenith angle, atmospheric transmittance (direct or diffuse sunlight), sea surface roughness, and wavelength (Payne, 1972; Simpson and Paulson, 1979; Katsaros et al., 1985). The most comprehensive study of sea surface albedo to date was performed by Payne (1972), who determined broadband albedo values for a variety of sun angles and atmospheric conditions. Climatological values of albedo for the low and mid-latitude regions of the Atlantic were reported to be mostly 0.06. In accordance with the findings of Payne (1972), an albedo of 0.06 is used here. Spectral structure is ignored as a complete study of spectral albedo is not known to exist.

c. Diffuse attenuation coefficient spectrum

In-water solar decay is quantified using the spectral diffuse attenuation coefficient, $K_d(\lambda)$ (eq. 4). A variety of parameterizations for $K_d(\lambda)$ have been proposed (Smith and Baker, 1978; Baker and Smith, 1982; Austin and Petzold; 1986; Siegel and Dickey, 1987b; Morel, 1988; Sathyendranath and Platt; 1988). These parameterizations have been developed mainly for modeling marine primary productivity, thus generally

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limited to the 400 - 700 nm spectral range. For radiant heating applications, a larger portion of the solar spectrum must be considered.

The parameterization of Baker and Smith (1982) is the best suited for radiant heating applications due to its extended spectral range (300 - 745 nm). In their work, Baker and Smith partitioned measurements of spectral diffuse attenuation coefficients from a variety of locations into clear water, "chlorophyll- like" pigment, and colored dissolved organic matter (CDOM) components. Here, the chlorophyll component of \( K_d(\lambda) \) is computed using upper ocean chlorophyll concentrations derived from the Coastal Zone Color Scanner (CZCS) level 3 composite imagery (Feldman et al., 1989), and the CDOM component is neglected as CDOM attenuation is negligible at wavelengths > 450nm (Baker and Smith, 1982). The CZCS data were averaged to compute monthly and annual mean climatological chlorophyll concentration fields, regridded to one degree squares, and values of \( K_d(\lambda) \) determined for 5 nm increments over the 300 - 745 nm range.

Spectral diffuse attenuation coefficient curves for open ocean waters containing varying amounts of chlorophyll are shown in figure 1. Below 300 nm and above 700 nm solar radiation attenuation increases rapidly. Solar energy contained in wavelengths between 350 and 600 nm is attenuated at a much slower rate, enabling penetration to significant depths. Net irradiance spectra at various depths for an incident flux of 200 W m\(^{-2}\) and chlorophyll concentrations of 0.1 mg m\(^{-3}\) and 1.0 mg m\(^{-3}\) show the decrease in spectral irradiance values at depth with increased chlorophyll concentration (figure 2). The corresponding spectrally integrated (300 to 745 nm) solar fluxes are shown in figure 3. A decrease in chlorophyll concentration from 1.0 to 0.1 mg m\(^{-3}\) results in a tripling of the net solar flux at 10 m, a 10 fold increase at 20 m, and a 50 fold increase at 40 m. Changes in upper ocean chlorophyll concentration significantly alters the net solar flux at depth through changes in spectral attenuation values (\( K_d(\lambda) \)) in the visible spectral range.

d. Mixed layer depth

Climatological monthly, seasonal, and annual mean profiles of temperature and salinity on a one degree grid over all ocean basins are included in the NOAA World Ocean Atlas 1994, an update of the original Levitus (1982) atlas (Levitus and Boyer, 1994; Levitus et al., 1994). Mixed layer depths have been calculated from these
profiles using a density change from the surface value ($\sigma_z(\text{mld}) = \sigma_z(0) + \alpha(0)\Delta T$ where $\alpha$ is the coefficient of thermal expansion and $\Delta T = 0.5^\circ C$; Sprintall and Tomczak, 1992), a criteria which considers both temperature and salinity. The maximum monthly mixed layer depth at each grid point was selected for calculation of $E_n(D_{ml})$.

4. Model validation

To illustrate the validity of the solar penetration model described here, in situ optical and physical data from the western equatorial Pacific warm water pool have been utilized (Siegel et al., 1995). These data were collected between 21 December, 1992 and 19 January, 1993 near 2°S, 156°E as part of the Tropical Ocean Global Atmosphere - Coupled Ocean Atmosphere Response Experiment (TOGA-COARE). Profiles of downwelling and upwelling spectral irradiance in 13 wavebands (from 340 to 683 nm) and standard CTD variables were sampled to 200 m from the R/V John Vickers. Water samples were collected twice daily for determination of chlorophyll $a$ concentration by Turner fluorometry. Spectral values of irradiance just above the sea surface were measured in the same 13 wavebands using a spectroradiometer identical to the profiling instrument and total solar irradiance incident at the sea surface was measured using an Eppley PSP pyranometer.

Data from two casts (a clear-sky and a cloudy-sky cast) have been selected for comparison with modeled values (figures 4 and 5). In situ values of spectral downwelling irradiance just above the sea surface, and spectral net irradiance at 10, 20, 40, and 60 meters are indicated by Xs and connected by lines. The corresponding modeled downwelling irradiance spectrum above the surface, and the modeled net irradiance spectrum at depth are shown by dotted lines. The spectral shape of the clear-sky data agrees well with the modeled spectral irradiance curves (figure 4a). This indicates that the 5S radiative transfer model (Tanre, 1979) adequately colors the incident solar flux data. Within the water column, modeled net spectral irradiance values are less than in situ values in the red portion of the spectrum. The modeled net spectral irradiance overestimates in situ values in the blue and blue-green spectral regions at 40 and 60 meters (beneath the mixed layer depth). Overestimation of fluxes beneath the mixed layer is expected as the model does not attempt to resolve the deep
chlorophyll maximum beyond the mixed layer base (figure 4b). Modeled and measured spectra are in close agreement beneath the mixed layer (40 and 60 m) when the model was forced with depth dependent chlorophyll values (not shown).

To compare total solar flux values, modeled and in situ spectral irradiance have been integrated from 340 to 683 nm (figure 4b). The modeled incident surface solar flux value underestimates the measured incident solar flux value by 3%. Within the upper 20 m, modeled values underestimate in situ solar fluxes by between 3 and 5%, indicating a model error of less than 2%, relative to the surface value. At depths of 40 and 60 m, well beneath the mixed layer, modeled values significantly overestimate in situ irradiance values due to the unresolved deep chlorophyll maximum. Model error was reduced to an overestimation of 5% beneath the mixed layer when driven with depth dependent chlorophyll values.

For the cloudy-sky comparison, modeled incident spectral irradiance values underestimate measured values, most notably in the blue and blue-green spectral regions (figure 5a). Within the water column, model underestimation of spectral net irradiance appears in the red portion of the spectrum (as was found with the clear sky comparison) and in the blue spectral region (figure 5a). The modeled incident solar flux value (after spectral integration) underestimates the in situ incident value by 14%. Within the upper 40 m, modeled net solar flux values underestimate in situ values by 15%, and at 60 m model underestimation is 8% (figure 5b). The error difference at 60 m is due to the deep chlorophyll maximum which exists just beyond 45 m (figure 5b). Relative to the surface flux bias, modeled values of net irradiance are accurate to within 3% in the mixed layer.

Nearly all of the model error for the cloudy-sky case appears in the spectral decomposition of the incident solar flux. The fact that model underestimation remains mostly constant with depth within the mixed layer suggests that the model accurately handles surface albedo and in-water solar decay. Model underestimation is most pronounced in the blue spectral region. Compared with the clear-sky case, it appears that clouds preferentially attenuate outside the blue and blue-green spectral regions (e.g. Spinhirne and Green, 1978). A detailed examination of cloud color utilizing this data set is presently underway.
The limited comparison of modeled and in situ data is not a thorough validation of this global solar radiation penetration model. Nevertheless, this comparison suggests that the solar flux within the mixed layer can be modeled to within a 5% error using a single mixed layer chlorophyll concentration. Such an error translates to a daily averaged flux value typically less than 1 W m\(^{-2}\) at a depth of 40 m. Significantly different conditions give rise to roughly the same in-water solar flux errors relative to incident surface values and point out the dual role that clouds can play in moderating solar fluxes. While clouds certainly reduce the net solar flux at depth by decreasing the solar flux incident at the sea surface, it also appears clouds may enhance solar penetration (relative to the surface flux) by shifting energy into the blue and blue-green spectral regions which penetrate to the greatest depths.

5. Results
a. Model inputs

Climatological annual mean values of \(E_d(0^+)^{\text{th}}\) derived from ISCCP data show the largest incident solar fluxes exist in the equatorial regions where clouds are sparse (figure 6a). In the Indian Ocean the regions of largest incident solar radiation lie just northwest of Australia and along the western boundary of the Arabian Sea. The Atlantic is characterized by a zonally uniform distribution of incident irradiance straddling the equator, with values decreasing with increasing latitude. In the Pacific, the region of greatest surface irradiance lies along the equator, between 150° W and 110° W longitude. Values quickly decrease just to the north of the equator, due to clouds associated with the Inter-Tropical Convergence Zone (ITCZ). South of the equator in the Pacific, values of surface irradiance decrease faster towards the west, due to the South Pacific Convergence Zone.

Climatological annual mean values of near-surface chlorophyll concentration derived from CZCS imagery are illustrated in figure 6b. The clearest ocean waters exist in the center of the major subtropical gyres where the convergence of surface Ekman transport and corresponding downwelling prevents upward nutrient fluxes. Regions of relatively chlorophyll rich waters exist along the equator, associated with the equatorial divergence and upwelling of nutrient rich waters (e.g., Barber and Chavez, 1991).
Polewards of the subtropical gyres, chlorophyll concentration increases with latitude. Coastal regions are characterized by relatively large values of chlorophyll concentration due to the occurrence of coastal upwelling, and possible input of terrigenous materials.

Monthly mean climatological mixed layer depths were computed using data from the NOAA World Ocean Atlas 1994 (Levitus and Boyer, 1994; Levitus et al., 1994) and a composite of the monthly maximum mixed layer depths was created (figure 6c). Using maximum monthly mixed layer depth ensures that thermal energy lost to the mixed layer in the form of solar penetration is lost on annual time scales, and is not merely trapped within seasonal pycnocline or barrier layer waters which are subsequently entrained into the mixed layer on seasonal time scales. The shallowest monthly maximum mixed layer depths are found in the tropics where values are mostly less than 50 m. In the central equatorial Pacific, maximum monthly mixed layer values extend to 100 m. Outside the tropics, maximum monthly mixed layer depths are mostly greater than 100 m. Mixed layer depths are generally at a maximum during winter months (figure not shown).

b. Penetrative solar fluxes

Climatological values of solar radiation penetrating the base of the deepest monthly mixed layer have been computed using the present solar penetration model (eq. 7) and are shown in figure 7a. Relatively large incident solar flux values, clear water, and shallow maximum mixed layer depths cause solar radiation penetrating the permanent pycnocline to be largest in the tropics. In the eastern tropical Pacific, solar penetration ranges from 15 to more than 30 W m⁻². Solar penetration values range from 10 to 20 W m⁻² in the equatorial Indian ocean, the Atlantic ocean, and in the western Pacific warm water pool region. In the central equatorial Pacific and parts of the equatorial Atlantic, relatively deep mixed layers are responsible for reducing solar fluxes penetrating the deepest monthly mixed layer depths to less than 7 W m⁻². In mid and high latitude regions the deepest of monthly mixed layer depths generally exceeds 100 m which prevents penetrating solar fluxes from exceeding more than a few W m⁻².

Solar penetration may be conveniently accounted for by evaluating values of \( \Pi \), which describe the fraction of available solar energy converted to thermal energy within the mixed layer (eq. 8). The values of \( \Pi \) presented here are climatological and specific to the attenuation coefficients and mixed layer depths used in their calculation. Regions
where values of $\Pi$ are low correspond to areas with large penetrative fluxes (figure 7b). Values of $\Pi$ are the smallest in the eastern equatorial Pacific where they range from 0.80 to 0.93, suggesting that as much as 20% of the incident solar flux is lost to penetration. In the equatorial Indian, equatorial Atlantic, and western equatorial Pacific, values of $\Pi$ are mostly in the 0.90 to 0.95 range. In the mid and high latitude regions where there is little penetration beyond the deepest monthly mixed layer depth $\Pi$ is mostly one, indicating complete attenuation of the incident solar flux above the deepest annual mixed layer depth.

Global estimates of climatological net air-sea heat flux values from Esbensen and Kushnir (1981) are illustrated in figure 7c for comparison with penetrating solar fluxes (figure 7a). Annual climatological values of the net air-sea heat exchange in tropical regions range from near zero along zonal bands at 20°N and 20°S to more than 60 W m$^{-2}$ in the eastern equatorial Pacific and Indian oceans. Comparison of the net air-sea flux with solar penetration suggests that the latter can be a large fraction of the net air-sea heat exchange. In zonal bands just off the equator, penetrative solar fluxes are about 10 W m$^{-2}$ while net air-sea heat fluxes are near zero. In the western and eastern equatorial Pacific, penetrative solar fluxes are between a quarter and half as large as net air-sea heat exchange values. This suggests that small errors in the penetrative component of solar radiation can have a large impact on estimates of mixed layer heating (cooling) and thus SST values. Solar penetration must not be neglected when computing mixed layer thermal evolution, particularly in the tropics where the net air-sea heat flux and penetrating solar flux are of the same order.

6. Discussion

*Evaluation of model performance*

This study introduces and applies a hybrid parameterization of in-water solar fluxes using remotely sensed data to identify regions where penetrative solar fluxes significantly effect mixed layer radiant heating rates. An obvious source of uncertainty in this solar penetration parameterization is the assumption that solar radiation in wavelengths outside the resolved spectral range (300-745 nm) does not contribute to mixed layer penetration. Attenuation coefficients for the clearest natural waters are
known to increase rapidly with increasing wavelength above 600 nm, and with
decreasing wavelength below 300 nm (Smith and Baker, 1981). Thus, energy outside
the evaluated portion of the solar spectrum will be quickly converted to thermal energy
after passing through the air-sea interface and will penetrate only the shallowest mixed
layers (< 10 m). To quantify any potential error associated with the limited spectral
range, a complete solar spectrum (250 - 2500 nm) was created with a radiative transfer
model and diffuse attenuation coefficients in the ultraviolet and near infrared spectral
regions were approximated using the pure water attenuation coefficient (Smith and
Baker, 1981) at 300 nm for wavelengths lower, and the value at 745 nm for
wavelengths higher. For an incident solar flux of 200 W m$^{-2}$, typical of mean
conditions over much of the ocean (figure 6a), the maximum solar flux outside the 300
- 745 nm range is 7.1 W m$^{-2}$ at 5 m, and 2.8 W m$^{-2}$ at 10 m, which is 3.5% and 1.4%
of the total solar radiation incident at the sea surface, respectively. These flux
estimates are extreme upper bounds as pure water values of K$_d$(λ) at 300 and 745 nm
severely underestimate actual K$_d$(λ) values outside the evaluated spectral range (figure
1). For mixed layer depths greater than 10 m, the penetrative solar flux can be
accurately determined by resolving only the 300 - 745 nm spectral region. However, to
adequately calculate solar fluxes closer to the surface, a greater portion of the solar
spectrum must be incorporated.

Knowledge of surface albedo over the global oceans is limited. For the
climatological calculations performed here, no explicit wavelength, latitude, solar zenith
angle, or atmospheric transmittance dependence on albedo is incorporated. It has been
suggested that albedo values for the ultraviolet and visible wavelengths are slightly
greater than for longer wavelengths (Katsaros et al., 1985; King et al. ,990).
However, no comprehensive study of these effects has been made. Small variations in
albedo will have little effect on deep penetrating solar fluxes in most regions. For
example, if the climatological albedo for mid to high latitude regions was changed from
0.06 to an extreme upper bound of 0.16 (Payne, 1972), a decrease in solar penetration
of no more than 1 W m$^{-2}$ would result (considering an incident solar flux no greater
than 200 W m$^{-2}$ and a Π value of 0.95). In tropical regions, a 1 W m$^{-2}$ decrease in
solar penetration is commensurate with an albedo increase of 0.04 (for an incident solar
flux of 250 W m$^{-2}$ and a Π value of 0.90). Thus, for the purpose of computing solar
fluxes which penetrate the permanent pycnocline, it is unlikely that use of a spatially uniform, spectrally independent albedo introduces a significant source of error.

Clouds have the ability to regulate the transmission of in-water solar fluxes by altering the ratio of diffuse to direct light and hence the radiance distribution at the sea surface. A decrease in atmospheric transmittance (increase in the diffuse to direct ratio) tends to smooth the diurnal cycle of the effective solar zenith angle potentially influencing both albedo and $K_d(\lambda)$. Atmospheric transmittance was considered by Payne (1972) in his determination of climatological albedo values. $K_d(\lambda)$ values for the penetrating wavelengths have been shown to be nearly invariant to typical illumination changes (Siegel and Dickey, 1987a; Gordon, 1989). Even if $K_d(\lambda)$ can be altered by the incident radiance distribution, a hypothetical 10% increase in $K_d(\lambda)$ values for a high solar zenith angle would result in less than a 1 W m$^{-2}$ decrease in the solar flux penetrating a relatively clear 50 meter mixed layer (for an incident solar flux of 50 W m$^{-2}$). Therefore, atmospheric influences on $\alpha$ and $K_d(\lambda)$ will have little to no effect on the penetrative solar flux values computed here.

Perhaps more important to solar penetration is the ability of clouds to alter the shape of the incident solar spectrum. Comparison of the modeled clear sky spectral decomposition used here with a measured cloudy sky irradiance spectrum shows that the amount of energy in the deep penetrating spectral wavebands (relative to total solar energy) can be at least as much as 14% greater in the presence of clouds (figure 5a). This suggests that spectral decomposition of the incident flux field is not uniform in space and time, but rather a function of atmospheric transmittance which is not considered here. A decrease in atmospheric transmittance will have the net effect of decreasing solar penetration as the overall decrease in the incident solar flux will be greater than the relative spectral enhancement in the deep penetrating wavebands. However, the relative energy shift into shorter wavelengths associated with decreased atmospheric transmission can effectively enhance the role of penetration by reducing $\Pi$ values. It is estimated that a 10 % increase in energy in the deep penetrating spectral region (relative to the broadband flux) can account for a decreases in $\Pi$ of about 0.01 for a mixed layer of 30 meters. Use of a clear sky spectral decomposition effectively puts a lower bound on the amount of energy in the deep penetrating spectral region (relative to the entire solar spectrum), and hence the fluxes of solar penetration reported

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here. Ideally, extent and type of cloud cover should be incorporated into the spectral decomposition of incident solar flux values for computation of solar penetration.

*Monthly evolution of mixed layer depth and solar penetration*

Computation of penetrative solar fluxes using maximum monthly mixed layer depth provides a lower bound on solar radiation penetration. This lower bound is extreme in mid to high latitude regions where large seasonal variations in mixed layer depth occur. Use of annual mean mixed layer depth results in penetrative solar flux values about 10 W m\(^{-2}\) greater for tropical regions, and as much as 25 W m\(^{-2}\) greater in mid and high latitude regions. While these larger fluxes do not completely penetrate the permanent pycnocline, they can result in thermal energy being trapped in seasonal pycnocline waters on monthly time scales before being entrained and made available for atmospheric exchange. A time lag in the availability of thermal energy lost to the mixed layer via solar penetration and subsequently entrained will affect climatic timescale air-sea interactions.

The roles of the annual solar cycle, seasonal mixed layer depth changes and phytoplankton blooms on the seasonal evolution of solar radiation penetration are illustrated using data from the sites of Ocean Weather Stations (OWS) November (N: 140°W, 30°N) and Papa (P: 145°W, 50°N) in the north Pacific Ocean (figure 8). As expected, \(\overline{E_d}(0^+)\) values are at a minimum during winter months, peak during summer months, and are a function of latitude. Incident solar flux values may change by more than 100 W m\(^{-2}\) during a semi-annual period and alone can be responsible for a factor of two change in solar penetration (figure 8). Such changes are evident throughout the mid and high latitudes. In tropical regions, the ITCZ shifts only a few degrees in latitude during the course of a year resulting in incident solar flux changes of less than 50 W m\(^{-2}\).

Spring phytoplankton blooms are evident at both stations N and P, and a fall bloom, characteristic of high latitudes, can be seen at OWS P (figure 8). In sub-arctic regions, CZCS estimates of chlorophyll concentration increase from \(-0.25\) mg m\(^{-3}\) during winter months to more than \(1.0\) mg m\(^{-3}\) during the spring bloom. This quadrupling in chlorophyll concentration results in a near 80% decrease (10 to 2 W m\(^{-2}\)) in the solar flux at 25 m, the annual mean mixed layer depth for that region \((\overline{E_d}(0^+) = 150\) W m\(^{-2}\)). The largest seasonal chlorophyll signal for tropical regions is
found in the Atlantic, where concentrations increase from near 0.1 mg m\(^{-3}\) during winter to 0.2 mg m\(^{-3}\) during spring. In contrast to high latitudes, this chlorophyll doubling corresponds to roughly a 40% decrease (24 to 14 W m\(^{-2}\)) in the quantity of solar radiation penetrating a typical mixed layer of 30 m (\(\bar{E}_d(0^+) = 240\) W m\(^{-2}\)).

The most pronounced seasonal signal occurs in mixed layer depth in mid and high latitude regions where significant shoaling generally begins during April, a minimum depth is reached about August, and deepening occurs throughout the remainder of the year (figure 8; Martin, 1985). Mixed layer depths range from less than 30 m during spring and summer months, to well over 100 m during fall and winter months in most mid and high latitude regions. This deepening results in near-complete elimination of solar penetration. For example, using annual mean incident irradiance and upper ocean chlorophyll concentration values typical of the mid to high latitudes (\(\bar{E}_d(0^+) = 150\) W m\(^{-2}\), chl=0.4 mg m\(^{-3}\)) such a mixed layer deepening causes the penetrating solar flux to decrease from 4 W m\(^{-2}\) to almost zero. Mixed layer heating rates are quite sensitive to changes in mixed layer depth. While this seasonal deepening has caused only a 4 W m\(^{-2}\) change in the penetrating solar flux, it has caused the mixed layer radiant heating rate to change dramatically, from \(-3^\circ\)C month\(^{-1}\) to less than 1 °C month\(^{-1}\), by increasing the volume of water to be warmed. By contrast, seasonal changes in mixed layer depth are less than 10 meters in the tropics.

At OWS N, monthly values of solar penetration range from near 1 W m\(^{-2}\) during the winter months to more than 30 W m\(^{-2}\) during summer (figure 8a). A relatively deep mixed layer and low incident solar flux keeps penetration to a minimum from January through March. Penetrating solar fluxes remain small in April as enhanced penetration associated with an increased incident flux and a shoaling mixed layer is offset by increased chlorophyll concentrations. Relatively large incident solar flux values, shallow mixed layers and relatively low chlorophyll concentrations make for maximum monthly penetrative fluxes of more than 30 W m\(^{-2}\) during June and July. A chlorophyll concentration increase (fall bloom) is primarily responsible for reduced penetration in August. From August through December, the mixed layer deepens and the incident solar flux decreases, resulting in a solar penetration reduction from 25 to 2 W m\(^{-2}\). Much of the summertime penetration is converted to thermal energy within seasonal pycnocline waters (figure 8a), as relatively little solar flux exists at the base of the
permanent pycnocline (figure 7a). Thermal energy deposited within seasonal pycnocline waters is entrained beginning in August. Thus, thermal energy associated with mixed layer penetration occurring from May through July is unavailable for atmospheric exchange until at least August and is not completely entrained until the following February. This indicates that thermal energy associated with seasonal solar penetration may become trapped in seasonal pycnocline waters, unavailable for atmospheric exchange, for up to nine months.

At OWS P, a relatively deep mixed layer and reduced incident solar flux keeps penetration to a minimum from January through April (figure 8b). The mixed layer shallows in May, however increased attenuation associated with the spring bloom prevents a penetration increase. Relatively shallow mixed layers, clear waters, and large incident flux values make for maximum penetrative solar fluxes from June through August. Throughout the remainder of the year, a deepening mixed layer, fall phytoplankton bloom, and decreasing incident solar flux all contribute to little or no solar penetration. Thermal energy associated with penetrating solar fluxes existing from June through September is trapped within seasonal thermocline waters until completely entrained the following January. Although solar penetration and the period for which associated thermal energy is trapped from atmospheric exchange are less extreme at OWS P, the annual cycle of penetration in high latitude regions may still influence air-sea heat exchange and should not be overlooked.

The possibility of a coupling between mixed layer depth and solar penetration on seasonal timescales exists, whereby solar penetration can influence the seasonal mixed layer evolution just as mixed layer depth effects solar penetration. When surface cooling terms nearly balance or exceed the mixed layer radiant heating rate, and solar penetration adds heat to the top of the seasonal pycnocline, the density gradient at the mixed layer base will decrease, thereby enhancing entrainment. Knowledge of all terms in the mixed layer heat budget is necessary for quantification of this process (e.g. Charlock, 1982). The seasonal cycle of solar penetration and the trapping of thermal energy within seasonal pycnocline waters must be considered for proper modeling of air-sea interaction processes and mixed layer evolution.

7. Conclusions
A hybrid parameterization for the determination of climatological in-water solar fluxes has been presented and applied to assess the regional distribution of solar penetration and its significance in the determination of ocean mixed layer thermal climate. Maximum monthly mixed layer depth values were used in the calculation of penetrative solar fluxes to ensure that thermal energy associated with solar penetration is not subsequently introduced into the mixed layer through entrainment of seasonal pycnocline waters. Solar penetration can be a significant portion of the upper ocean mixed layer heat budget on annual scales in the tropics and must be considered seasonally in the mid and high latitude regions. Annual climatological values of solar penetration for tropical regions can reach 40 W m$^{-2}$, resulting in overestimates of the mixed layer heating rate of 0.33 °C month$^{-1}$ if not accounted for. Beyond the tropics, thermal energy associated with penetrating solar fluxes upwards of 40 W m$^{-2}$ can be deposited within the seasonal pycnocline, trapped from atmospheric exchange for months. Errors in mixed layer radiant heating rates due to the improper treatment of solar penetration can propagate to subsequent rates of air-sea heat exchange. For coupled air-sea modeling efforts, consideration of solar penetration is necessary.

A nondimensional parameter, $\Pi$, has been introduced to quantify the fraction of available solar energy that is converted to thermal energy within the mixed layer. In tropical regions, $\Pi$ values are as small as 0.85, indicating that 15% of the total solar input will be lost from mixed layer heating through solar penetration on annual time scales. In sub-tropical and sub-polar regions, $\Pi$ values can be significant on monthly to seasonal time scales.

There has been recent discussion regarding use of the mixed layer heat budget for the western equatorial Pacific warm water pool to “deduce the effect of clouds on net solar radiation at the sea surface” (Ramanathan et al., 1995). Ramanathan and collaborators attributed the lack of balance in the annual warm pool mixed layer heat budget to an anomalous absorption of solar radiation by clouds. However, in their calculations, no consideration for solar penetration was made. Values of solar penetration computed here using the deepest monthly average mixed layer depth indicate that for the western Pacific warm pool region (140° to 170°E, 10°S to 10°N) the mean penetrative solar flux is 12 W m$^{-2}$. By utilizing a mixed layer heat budget
without a solar penetration term, Ramanathan et al. may have overstated their anomalous absorption estimate by about 40%.

Global annual values of $E_n(D_{ml})$ and $\Pi$ presented here are the first of their kind. These estimates may be improved upon by considering regional and seasonal analyses. Plans currently exist for a 20 year time series of remotely sensed ocean color measurements (Abbott et al., 1994). Near real-time ocean color imagery and remotely sensed cloud data can be used with the model presented here to provide further solar penetration information ($\overline{E}_n(z)$) to ocean circulation models. Accurate knowledge of mixed layer heat content, a quantity related to the seasonal evolution of radiant heating rate and solar penetration, will contribute to an improved understanding of the coupling between atmospheric forcing, upper ocean physical and biological processes.

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Figure List

Figure 1. Values of the diffuse attenuation coefficient spectrum, \(K_d(\lambda)\), for open ocean waters with varying chlorophyll concentrations (0.01, 0.1, 0.2, 0.5, 1.0, and 2.0 mg m\(^{-3}\)) following Baker and Smith (1982).

Figure 2. Spectral net irradiance at various depths (just beneath the surface, 10, 20, 40, 60, 80, and 100 meters). Values are based on an incident solar flux of 200 W m\(^{-2}\) and a) 0.1 mg m\(^{-3}\), b) 1.0 mg m\(^{-3}\) of chlorophyll \(a\) concentration.

Figure 3. Vertical profiles of net irradiance for an incident solar flux of 200 W m\(^{-2}\) and chlorophyll \(a\) concentrations ranging from 0.01 to 2.0 mg m\(^{-3}\) after spectral integration from 300 to 750 nm.

Figure 4. a) Comparison of in situ (Xs connected by lines) and modeled (dotted lines) values of spectral downwelling irradiance just above the sea surface, and spectral net irradiance at 10, 20, 40, and 60 meters. Cast recorded at 1315 local time on 14 January, 1993 during a period of clear skies, light winds, 21° solar zenith angle, and total incident solar flux of 965 W m\(^{-2}\). b) Comparison of in situ and modeled irradiance data after spectral integration from 340 to 683 nm. Sigma-theta derived from in situ CTD data. Water sampling indicates chlorophyll \(a\) concentrations are constant at \(~0.12\) mg m\(^{-3}\) in the upper 20 m, increase to \(~0.15\) mg m\(^{-3}\) at 40 m, and to \(~0.40\) mg m\(^{-3}\) at 60 m. Mixed layer depth is \(~20\) m.

Figure 5. As in figure 5, except comparing data recorded at 1148 local time on 28 December, 1992 under cloudy skies, relatively strong winds, 23° solar zenith angle, and total incident solar flux of 541 W m\(^{-2}\). Chlorophyll concentration is constant at \(~0.08\) mg m\(^{-3}\) in the upper 40 m and increases to 0.12 mg m\(^{-3}\) at 60 m. The mixed layer extends below 60 m.

Figure 6. Climatological estimates of a) downwelling irradiance at the sea surface derived from ISCCP data (W m\(^{-2}\)), b) upper ocean chlorophyll concentration derived
from CZCS imagery (mg m\(^{-3}\)), and c) maximum monthly mixed layer depth computed from temperature and salinity profiles and a sigma-t criterion (m).

**Figure 7a.** Modeled climatological values of the net solar flux at the base of the deepest monthly mixed layer (W m\(^{-2}\)). Values correspond to solar fluxes entering the permanent pycnocline. Largest values exist where the deepest monthly mixed layer and chlorophyll concentration are low, and the incident solar flux is high.

**Figure 7b.** Climatological estimates of \(\Pi\), the nondimensional mixed layer radiant heating parameter illustrating the importance of solar radiation penetration relative to the available surface flux. Values of \(\Pi\) indicate the fraction of incident solar radiation which is converted to thermal energy above the permanent pycnocline.

**Figure 7c.** Climatological values of the net air-sea heat flux from the Esbensen and Kushnir (1981) ocean atlas (W m\(^{-2}\)). A comparison of the net air-sea heat flux and the penetrative solar flux (Fig. 7a) indicates the importance of solar radiation penetration.

**Figure 8.** Timeseries of mean monthly values for the incident solar flux (W m\(^{-2}\)), upper ocean chlorophyll concentration (mg m\(^{-3}\)), mixed layer depth (m), and calculated penetrative solar flux (W m\(^{-2}\)) for a) OWS November (140°W, 30°N and b) OWS Papa (145°W, 50°N).
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Figure 1
Figure 3
Figure 4
Figure 5
Figure 8
Chapter #2

Radiant Heating of the Western Equatorial Pacific during TOGA-COARE

Abstract

Optical, physical and biological data collected in the western Pacific Warm Water Pool as part of the Tropical Ocean Global Atmosphere - Coupled Ocean Atmosphere Response Experiment (TOGA-COARE) are used to assess variations in the transmission of solar radiation through the water column, to investigate factors which regulate in-water solar transmission, and to examine the sensitivity of upper ocean thermal processes to solar transmission parameterizations. Solar transmission within the upper ocean mixed layer, below 10 m, can be accurately parameterized (within 10%) using a single exponential profile. Variations in the in-water transmission profile are explained primarily by estimates of the upper ocean chlorophyll concentration ($r^2=0.83$). Mixed layer chlorophyll concentration influences attenuation of the in-water light field. Clouds play a secondary role by altering the relative shape of the incident solar spectrum.

A comparison of solar transmission parameterizations indicates that use of a Jerlov type parameterization for the WWP can lead to a 15 W m$^{-2}$ error in the solar flux at 30 m (based on a climatological surface irradiance of 220 W m$^{-2}$). Application of a one-dimensional ocean mixed layer model gives a mean SST error of 0.15 C when the model is forced with a typical Jerlov type solar transmission profile. Instantaneous SST differences nearly reach 1.0 C. An increase in solar flux divergence for a mixed layer results in enhanced stratification and mixed layer shoaling. As mixed layer depth decreases, the quantity of solar radiation penetrating beyond the mixed layer increases, resulting in a destabilization of thermocline waters and a mixed layer depth increase. This feedback mechanism helps regulate both mixed layer depth and the solar irradiance lost to the mixed layer through penetration.
1. Introduction

Global circulation models are highly sensitive to variations in sea-surface temperature (SST), particularly in regions of high SST such as the western equatorial Pacific warm water pool (WWP; Gent 1991; Rasmusson and Carpenter 1982; Webster and Lukas 1992). Additionally, atmospheric convection rates and the local wind field are largely dependent upon SST for tropical oceans (Graham and Barnett 1987). A recent study of cloud radiative forcing has relied heavily on WWP heat budget accuracy (Ramanathan et al. 1995). Such dependencies suggest a need for proper quantification of upper ocean thermal processes for the Warm Pool (e.g. Webster and Lukas 1992).

Air-sea fluxes regulate mixed layer heat content in regions characterized by light winds, weak advection, and a shallow mixed layer, such as the western equatorial Pacific (McPhaden and Hayes 1991). Climatological estimates of the net air-sea heat flux for the WWP range from 10 to 70 W m\(^{-2}\) (Esbensen and Kushnir 1981; Weare et al. 1981; Reed 1985; Godfrey and Lindstrom 1989). An air-sea flux error of only 10 W m\(^{-2}\) can alter the temperature prediction for a 20 m mixed layer by nearly 1 C over a three month period. A primary goal of the Tropical Ocean Global Atmosphere - Coupled Ocean Atmosphere Response Experiment (TOGA-COARE) is an improved understanding of processes in the coupled ocean-atmosphere thermal system leading to the ultimate goal of enhancing climate modeling efforts (Webster and Lukas 1992).

Radiant heating is the largest term in the upper ocean heat budget of the western equatorial Pacific Ocean and is unique to the air-sea flux balance. Unlike other fluxes which act solely at boundaries, solar radiation penetrates the air-sea interface, and through its vertical divergence, directly heats water well beneath the ocean surface. It is insufficient to consider only the incident solar flux when addressing mixed layer thermal processes. Rather, it is necessary to consider the solar flux divergence, and the fraction of the incident solar flux which passes beyond the mixed layer base. This penetrating solar irradiance is lost to mixed layer heating and subsequent direct sea-air exchange processes (e.g. Lewis et al. 1990; Hayes et al. 1991; Sathyendranath et al. 1991; Siegel et al. 1995a; Anderson et al. 1996; Ohlmann et al. 1996).

Here, data collected as part of TOGA-COARE are used to assess variations in solar radiation transmission and to evaluate its role in ocean radiant heating. In section 2.
solar transmission and related parameters are formally defined. The COARE dataset is described in section 3. Data is presented and used to develop a mean solar transmission parameterization for the western equatorial Pacific in section 4. Relationships between transmission model parameters and measurable physical and biological quantities are developed in section 5. The COARE solar transmission parameterization is compared with other transmission parameterizations in section 6. The sensitivity of upper ocean evolution to solar transmission is examined with a one dimensional upper ocean mixed layer model in section 7. Results are summarized in section 8.

2. Parameters for quantification of solar transmission

The spectral diffuse attenuation coefficient ($K_d(t,z,\lambda)$) is typically used to quantify the vertical decay of downwelling plane irradiance and is defined as

$$K_d(t,z,\lambda) = -\frac{d(\ln E_d(t,z,\lambda))}{dz} (1)$$

where $E_d(t,z,\lambda)$ is the downwelling spectral irradiance dependent on time, depth (positive downwards) and wavelength. The diffuse attenuation coefficient can be used to relate downwelling spectral irradiance just beneath the surface (0') with that at depth (z) following

$$E_d(t,z,\lambda) = E_d(t,0',\lambda) \exp(-\int_{0'}^{z} K_d(t,z',\lambda) \, dz'). (2)$$

Formally, $K_d(t,z,\lambda)$ is defined as an apparent optical property, that is, dependent on both the radiance distribution and the optical properties of the water column (Mobley 1994). However, $K_d(t,z,\lambda)$ has been shown to be “quasi-inherent” or nearly invariant to changing radiance distributions (Baker and Smith 1980; Siegel and Dickey 1987b; Gordon 1989).

The diffuse attenuation coefficient spectrum is an important quantity in the computation of spectral solar transmission, $Tr(t,z,\lambda)$, which is the fraction of spectral irradiance incident at the sea surface which exists at depth z, or

$$Tr(t,z,\lambda) = \frac{E_n(t,z,\lambda)}{E_d(t,0^+,\lambda)} (3)$$
where the net flux of spectral irradiance at depth, \( E_n(t, z, \lambda) \), is defined as the difference between downwelling (\( E_d(t, z, \lambda) \)) and upwelling (\( E_u(t, z, \lambda) \)) spectral irradiance. After introducing the irradiance reflectance spectrum used to relate upwelling and downwelling irradiance (\( R(t, z, \lambda) = E_u(t, z, \lambda) / E_d(t, z, \lambda) \)), and defining surface reflectance, \( R_s(t, \lambda) \), as the fraction of the incident solar flux reflected by the sea surface, the spectral transmission function may be written as

\[
\text{Tr}(t, z, \lambda) = (1 - R(t, z, \lambda))(1 - R_s(t, \lambda)) \exp(-\int_0^z K_d(t, z', \lambda) \, dz')
\]

Sea-surface albedo is defined as the ratio of upward to downward irradiance just above the air-sea interface. Upward irradiance in the albedo definition is composed of light reflected by the sea-surface and reflected light from below which passes through the interface (Payne 1972). Assuming \( R(t, z, \lambda) \) is independent of depth (valid for a homogenous upper ocean mixed layer), albedo, \( \alpha(t, \lambda) \), can be approximated as the sum of \( R(t, 0^+, \lambda) \) and \( R_s(t, \lambda) \), minus their product, and spectral transmission can be written as

\[
\text{Tr}(t, z, \lambda) \equiv (1 - \alpha(t, \lambda)) \exp(-\int_0^z K_d(t, z', \lambda) \, dz')
\]

The total (spectrally integrated) transmission, \( \overline{\text{Tr}}(t, z) \), relating the total net solar flux at depth to the total solar flux impinging on the sea surface, may be directly obtained from irradiance data after spectral integration (\( \overline{\text{Tr}}(t, z) = \overline{E_n(t, z)} / \overline{E_d(t, 0^+)} \)) with overbars indicating integration over \( \lambda \) or calculated from surface irradiance, albedo, and the diffuse attenuation coefficient spectrum as

\[
\overline{\text{Tr}}(t, z) \equiv \frac{\int \overline{E_d(t, 0^+, \lambda)} (1 - \alpha(t, \lambda)) \exp(-\int_0^z K_d(t, z', \lambda) \, dz') \, d\lambda}{\overline{E_d(t, 0^+)}}.
\]

By assuming \( K_d(t, z, \lambda) \) is independent of depth (valid for the mixed layer) and integrating the numerator, the solar transmission function can be approximated as

\[
\overline{\text{Tr}}(t, z) \equiv \overline{A(t)} \exp(-\overline{K(t)} z)
\]

where \( \overline{A(t)} \) represents the fraction of the total incident irradiance in the resolved spectral region and \( \overline{K(t)} \) indicates the manner in which irradiance in the resolved region is
attenuated with depth (Tyler 1958, Siegel and Dickey 1987a). The spectral range resolved here (340 - 683 nm) allows total transmission to be determined for depths greater than 10 meters. For shallower depths, a greater portion of the solar spectrum must be included (e.g. Siegel et al. 1995a; Ohlmann et al. 1996). Solar transmission variations within the top few meters are presently being examined using a full spectral radiative transfer model.

3. Data and methods

Optical, physical and biological measurements were made aboard the R/V John Vickers from December 21, 1992 to January 19, 1993 in the vicinity of 2.08°S, 156.25°E as part of the TOGA-COARE intensive observation period (IOP: November 1, 1992 through February 28, 1993). Data were collected with a bio-optical profiling system (BOPS: Smith et al., 1984) at the rate of three 200 meter profiles per hour, with intermittent breaks for water sampling, vessel re-positioning, foul weather and instrument maintenance. More than 1500 casts were performed. The BOPS contained a spectroradiometer which measured downwelling and upwelling spectral irradiance in 13 discrete wavebands between 340 and 683 nm (MER-2040; Biospherical Instruments, San Diego, CA), a chlorophyll fluorometer (SeaTech, Corvallis, OR), a beam transmissometer (660nm, 25cm: SeaTech, Corvallis, OR) and a standard CTD system (SeaBird, Bellevue, WA). A similar spectroradiometer and an Eppley PSP pyranometer (Eppley Instruments, Newport, RI) were mounted on the ship’s mast to record incident spectral irradiance and the total incident radiation respectively. Surface measurements were made only during periods when the BOPS was deployed. A continuous timeseries of incident solar radiation data from which mean values can be calculated was measured at the TOGA-COARE central mooring (IMET: 1.75°S, 156.00°E; Weller and Anderson 1996). Seawater samples were collected twice daily (at local noon and midnight) and analyzed for chlorophyll a and phaeopigment concentrations using both standard fluorometric and high performance liquid chromatography techniques (Bidigare 1991). A more detailed discussion of these TOGA-COARE measurements is given in Siegel et al. (1995a).

Data processing was carried out following procedures developed for optical data by Siegel et al. (1995b). Continuous cast data were averaged into 1 m depth bins with
corresponding surface irradiance data. Spectral diffuse attenuation coefficient values, \( K_d(t,z,\lambda) \), were calculated at each bin depth by computing the slope of a least squares regression line fit to binned spectral irradiance data over a 10 m depth window (eq. 1). All binned data and the calculated spectral diffuse attenuation coefficients were then subsampled at 5 meter depth increments (5, 10, 15,...) resulting in 30 day timeseries at each subsampled depth with temporally corresponding surface values. Spectral irradiance data at each 5 m bin depth and corresponding surface values were then integrated over the resolved spectral region (340 - 683 nm). Finally, \( \overline{Tr}(t,z) \) was calculated as the ratio of spectrally integrated irradiance at depth to total surface irradiance for depths greater than 10 m (Ohlmann et al. 1996).

4. Results

4a. Solar transmission

A time-depth series of daily averaged solar transmission for the 30 day sampling period (figure 1) illustrates a high degree of solar transmission variability for the WWP. For the shallower depths, relative changes in daily mean \( \overline{Tr}(t,z) \) values are small, but can correspond to relatively large changes in absolute solar flux values. In contrast, large relative changes in transmission at the deeper depths result in relatively small changes in the net solar flux. For example, daily averaged values of \( \overline{Tr}(t,z) \) at 15 and 50 m change by 40 and 200 % respectively over the sampling period. These relative differences in transmission correspond to absolute solar flux variability of 13 and 9 W m\(^{-2}\) (15 and 50 m) based on a climatological incident irradiance of 220 W m\(^{-2}\) (Ohlmann et al 1996). For a typical WWP mixed layer depth (30 m; Lukas and Lindstrom 1991), values of solar transmission range from 0.07 (January 4, 1993) to 0.13 (December 21, 1992), corresponding to typical penetrative solar fluxes of 15 to 29 W m\(^{-2}\).

Individual values of \( \overline{Tr}(t,z) \) are used to determine the cruise mean transmission profile (figure 2). Following equation 7, model parameters were fit in a least squares sense using a single exponential regression over the 10 to 50 m depth range. Solar fluxes exist outside the resolved spectral region above 10 meters. Decreased solar transmission due to a deep chlorophyll maximum, mostly beneath 50 m, can skew a linear fit. Based on 30 days of sampling, the mean solar transmission profile is
\[ \overline{\text{Tr}}(z) = 0.355 \exp(-0.045z) \]

and standard error estimates for the model parameters \( A(t) \) and \( K(t) \) are 0.070 and 0.009 respectively. This single exponential parameterization does an excellent job of representing mean transmission values. Parameterized values of \( \overline{\text{Tr}}(z) \) explain more than 99.8% of the variance in cruise mean transmission values computed for each depth (10 to 50 m) by averaging. This shows that a single exponential function is adequate for modeling solar transmission over the mixed layer. Depth dependence in the attenuation parameter \( (K(t)) \) must be included for proper modeling of transmission below the mixed layer base where light attenuation can be a strong function of depth (Siegel and Dickey 1987a, Ohlmann et al. 1996).

4b. Surface irradiance

The primary effect on in-water solar fluxes comes from regulation of solar radiation reaching the sea surface by clouds (figure 3). Daily mean incident solar irradiance values for the sampling period range from 47 to 270 W m\(^{-2} \). Variations in the incident radiation will, to a large extent, propagate to depth. However, changes in spectral composition of the incident irradiance influence solar transmission and prevent this from occurring exactly (Ohlmann et al. 1996, Siegel et al. submitted). The ratio of incident downwelling spectral irradiance \( (\overline{E}_d(t,0^+,\lambda)) \) to total incident irradiance \( (\overline{E}_d(t,0^+)) \), plotted as a function of wavelength and irradiance, illustrates how spectral composition is related to incident irradiance (figure 4). For small values of \( \overline{E}_d(t,0^+) \), the surface irradiance has a spectral peak between 450 and 500 nm, and irradiance decreases slowly with respect to wavelength outside this band. As \( \overline{E}_d(t,0^+) \) increases, a relative shift in spectral composition flattens the spectral structure. Simply stated, a decrease in total solar radiation reaching the sea surface is accompanied by a relative enhancement of spectral irradiance in the blue-green region (Siegel et al. submitted). The spectral irradiance fractions shown in figure 4 are cast averages computed only between 1000 and 1400 local time. Solar zenith angles are less than 40° during this time interval, with the minimum incident clear sky irradiance exceeding 750 W m\(^{-2} \). By displaying only this subset of spectral irradiance fractions, effects of solar zenith angle on the spectral irradiance to total irradiance ratio are minimized. Contours of this ratio
for other isolated time periods (solar zenith angle ranges) indicate the same general pattern.

4c. **Subsurface parameters (chlorophyll concentration and spectral attenuation)**

Vertical changes in the subsurface light field are a function of absorption and scattering properties which, in turn, are related to type and amount of biomass and detrital materials present in the water column (Jerlov 1976; Smith and Baker 1981; Kirk 1994). A time-depth series of upper ocean chlorophyll a concentration is shown in figure 5. The first two weeks of the sampling period are characterized by strong westerly winds which create an anomalously deep mixed layer (Smyth et al. 1996a; Wijesekera and Gregg 1996; Cronin and McPhaden 1997). During this time, mixed layer chlorophyll concentration is near 0.1 mg m\(^{-3}\), and concentration within the deep chlorophyll maximum is \(~0.3\) mg m\(^{-3}\). When the westerly wind event ends (January 4, 1993), the mixed layer shoals and mixed layer chlorophyll concentration increases to \(~0.2\) mg m\(^{-3}\). Chlorophyll concentration within the deep chlorophyll maximum increases to more than 0.4 mg m\(^{-3}\). The size and magnitude of the chlorophyll increase is consistent with advection, however there is no spatial data to confirm this (e.g. Cronin and McPhaden 1997). The chlorophyll concentrations reported here are within the range of values previously observed for the Warm Pool region (Dandonneau 1992; Radenac and Rodier 1996).

**In situ** chlorophyll fluorescence signals provide a continuous measure of relative chlorophyll a concentration changes (Dickey and Siegel 1993, Kiefer and Reynolds 1992). Chlorophyll fluorescence exhibits a time-depth evolution quite similar to that of chlorophyll concentration (not shown). A diurnal signal in fluorescence exists due to phytoplankton physiological processes related to photoquenching, making it difficult to properly determine the daily cycle in chlorophyll concentration from fluorescence data (e.g. Kiefer and Reynolds 1992). However, daily averaged values of fluorescence are representative of relative variations in chlorophyll pigment biomass (Dickey and Siegel 1993). A linear regression between daily averaged values of depth integrated (surface to 50 m) chlorophyll fluorescence and chlorophyll concentration gives a correlation coefficient of 0.74. This indicates a strong relationship considering chlorophyll data
come from twice-a-day samples at discrete depths (surface, 20, 40 and 60 meters), and fluorescence is determined from ~25 profiles made each day.

The attenuation of solar radiation within the water column is roughly exponential with depth as a function of wavelength. Attenuation can be partitioned into clear water, chlorophyll, and colored dissolved organic material (CDOM) components (Baker and Smith 1982, Siegel and Michaels, 1996). Clear water is most transparent in the blue-green spectral region with attenuation increasing dramatically with increasing wavelength in the near-infrared spectral region. In contrast, the CDOM attenuation component decreases exponentially with increasing wavelength. The attenuation coefficient for chlorophyll generally decreases with increasing wavelength, with peaks near 440 and 680 nm (Kirk 1994). The relatively clear waters of the WWP are most transparent to solar energy in the blue-green spectral region. Little solar energy exists beyond the upper few meters for wavelengths greater than ~620 nm.

Time-depth series of daily averaged spectral diffuse attenuation coefficients ($K_d(t,z,\lambda)$) at three wavelengths (380, 490, and 565 nm) are shown in figure 6. Values of $K_d(t,z,380)$ for the first half of the timeseries are near 0.04 m$^{-1}$ in the upper 50 meters and increase to 0.10 m$^{-1}$ near 90 meters (figure 6a). On January 4, 1993, just after the westerly wind burst event, $K_d(t,z,380)$ values increased to near 0.06 m$^{-1}$ in the upper 80 meters and decreased to 0.08 m$^{-1}$ below this depth. These changes suggest a redistribution of attenuating material within the water column following the wind event (figure 6a). The $K_d(t,z,380)$ distribution is well correlated with that of chlorophyll (figure 5) for the first half of the sampling period. However, the deep chlorophyll maximum present near 50 meters after January 4th is not well represented by attenuation at 380 nm. This suggests the presence of a CDOM attenuation component (Siegel and Michaels 1996).

Values of the diffuse attenuation coefficient at 490 nm (figure 6b) range from 0.02 to 0.06 m$^{-1}$ and display a pattern similar to chlorophyll concentration. Attenuation at 490 nm is generally 50 to 75% that at 380 nm as decreases in attenuation by both clear water and CDOM components are overcome by increased attenuation due to chlorophyll (Kirk 1994). On January 4, attenuation at 490 nm increases in the upper 80 m and decreases below 80 m (figure 6b). Values of the diffuse attenuation coefficient at 565 nm (figure 6c) range from 0.07 to 0.09 m$^{-1}$. In contrast to the shorter wavelengths,
$K_d(t,z,565)$ variability is slight as the chlorophyll and CDOM attenuation components are small relative to the clear water component. As a result, agreement between $K_d(t,z,565)$ and the chlorophyll distribution is poor. Neither the chlorophyll increase on January 4th nor the subsurface chlorophyll maximum are readily apparent in the time-depth series of diffuse attenuation data at 565 nm (figure 6c).

Timeseries of $K_d(t,z,\lambda)$ for the resolved spectral region (340 to 683 nm) at fixed depths of 20 and 80 meters illustrate the spectral dependency of diffuse attenuation along with the typical range of variability encountered for mixed layer (20 m: figure 7a) and pycnocline (80 m: figure 7b) waters. Values of $K_d(t,20,\lambda)$ are relatively large and constant at wavelengths above 550 nm (figure 7a). These values are within 20% of the diffuse attenuation coefficients for clear natural water given by Smith and Baker (1981). The region of greatest transparency exists near 480 nm where attenuation ranges from 0.03 to 0.05 m$^{-1}$. Variations in attenuation near this wavelength are due to variations in chlorophyll concentration as this is near the chlorophyll $a$ absorption peak. Temporal variability is greatest at 340 nm. Attenuation at this wavelength ranges from 0.06 to 0.10 m$^{-1}$ presumably due to variations in CDOM and chlorophyll (Kirk 1994; Siegel and Michaels 1996). The response in spectral attenuation to the chlorophyll bloom on January 4th is evident for wavelengths up to ~550 nm (figure 7a).

Values of $K_d(t,80,\lambda)$ are nearly constant and similar to those at 20 m for wavelengths between 525 and 575 nm (figure 7b). Attenuation at 80 m is at a minimum near 480 nm. Between 400 and 450 nm, $K_d(t,80,\lambda)$ ranges from 0.05 to 0.09 m$^{-1}$ presumably due to movement of the deep chlorophyll maximum across the 80 meter isopleth (figure 5). Variations at 340 nm are slight compared to those present at 20 m. Both $K_d(t,80,\lambda)$ for the 400-450 nm range and chlorophyll concentration values at 80 m are near double those at 20 meters. The greatest changes in $K_d(t,80,\lambda)$ occur around January 2nd, two days prior to the large mixed layer chlorophyll increase (figure 7b). This is likely due to the passing of the deep chlorophyll maximum through the 80 m isopleth.

5. Variations in solar transmission
The previous examination of surface and subsurface parameters illustrates variations in spectral composition of the incident irradiance and the spectral diffuse attenuation coefficients. The shape of the incident solar spectrum is influenced by cloud type and amount, and the spectral diffuse attenuation coefficient is a strong function of the type and quantity of attenuating materials in the water column. Together, these factors are responsible for variations in the transmission of solar radiation through the upper ocean. In this section, timeseries of the $A(t)$ and $K(t)$ parameters from the simple exponential model for $\overline{T_r}(t,z)$ are presented (eq. 7). These parameters, which represent solar transmission variations, are then compared with coincident physical and biological parameters.

Timeseries of daily averaged solar transmission at three depths (20, 40 and 60 m) (computed as $\overline{T_r}(t,z) = \overline{E_n}(t,z) / \overline{E_d}(t,0^\circ)$) are compared with transmission values computed from single exponential profiles determined empirically by day (eq. 7) in figure 8. Data and modeled values are within 5% at 20 and 40 meters and within 10% at 60 meters indicating variations in transmission on mixed layer depth scales can be successfully explained as a single exponential with two model parameters.

The physical and biological factors which can alter the $A(t)$ and $K(t)$ parameters in equation 7 include upper ocean biomass concentration and atmospheric transmittance. Clearly chlorophyll biomass plays a primary role in regulating the attenuation of irradiance within the water column (Siegel et al. 1995). Atmospheric transmittance influences the relative spectral composition of the incident irradiance and its angular distribution (total solar attenuation depends on both the spectral distribution of energy, as attenuation is a strong function of wavelength, and the mean cosine of the light field). Atmospheric transmittance is a function of cloud properties, solar zenith angle, and atmospheric constituents such as water vapor and aerosols. Atmospheric transmittance is described here in terms of a cloud index, defined as one minus the ratio of measured incident irradiance to modeled clear sky incident irradiance (e.g. Gautier et al. 1980; Siegel et al. submitted).

Timeseries of model parameters and presumably related independent parameters are shown in figure 9. The daily averaged value of depth integrated (0 - 50 m) chlorophyll fluorescence is relatively constant near ~5 mg m$^{-2}$ until January 4th, increases to more than ~12 mg m$^{-2}$, and then generally declines (figure 9). The fluorescence and $K(t)$
parameter timeseries are in close agreement as expected. Daily averaged values of the incident solar irradiance range from 50 to 270 W m\(^{-2}\) (figure 9). Prior to January 9th, overcast days are common, whereas the remainder of the sampling period is characterized by incident irradiance values mostly greater than the cruise mean (189 W m\(^{-2}\)). A cloud index is computed using these incident solar flux values and clear sky values from the DISORT atmospheric radiative transfer model (Stamnes et al. 1988; Siegel et al. submitted). The daily averaged cloud index ranges from less than 0.1 to more than 0.8, indicating clouds reduce the clear sky incident solar flux by between 10 and 80% (figure 9). The ratio of incident irradiance in the resolved spectral region to the total broadband value ranges from near 0.50 to more than 0.60 indicating that between 50 and 60% of the incident flux is contained in the 340 to 683 nm spectral region (figure 9). There is no obvious correlation between solar flux, cloud index or incident irradiance ratio and model parameters.

Relationships among parameters may be better elucidated through examination of a correlation matrix (Table 1a). As expected, the most significant relationship exists between chlorophyll fluorescence (chlorophyll biomass) and the model attenuation parameter (K(t); eq. 7). The correlation coefficient indicates that upper ocean chlorophyll explains 83% of the variability in K(t). Statistically significant (at the 95% level) correlations between A(t) and K(t), and A(t) and fluorescence are most likely an artifact of the resolved spectral region and curve fitting. Single exponential fits are performed over the 10 to 50 m depth range, and A(t) is essentially the surface intercept. Thus, increases in A(t) can accompany increases in slope due to increased attenuation (K(t) or fluorescence). The statistically significant relationship between cloud index and fraction of the total incident irradiance in the visible spectral region indicates that changes in relative spectral composition can be explained through a cloud index (eg. Siegel et al. in review).

To address second order relationships, the data were high pass filtered to remove variability associated with the chlorophyll biomass increase on January 4th. The ad-hoc filtering scheme computes mean values for both the pre-chlorophyll increase (period 1; December 21 - January 3) and post-chlorophyll increase (period 2: January 4 - January 19) periods and removes the appropriate mean from data in the respective time periods. A correlation matrix for the residuals (Table 1b) shows a statistically
significant correlations (95% significance level) between K(t) and fluorescence, and between A(t) and the ratio of visible to total irradiance. This reiterates the importance of upper ocean chlorophyll concentration on transmission, and suggests that spectral composition of the incident irradiance does contribute to variations in solar transmission on mixed layer depth scales and may be important during periods of relatively constant chlorophyll concentration. The relationship between cloud index and the ratio of visible to total irradiance was discussed above. Fluorescence yields covary with cloud index as photoquenching is less likely in reduced light conditions (Kiefer 1973).

6. Discussion

6a. Comparison of solar transmission parameterizations

The total net solar flux at depth (\(\overline{E}_n(t, z)\)) can be simply modeled by combining the transmission parameterization (eq. 7) and total incident irradiance as

\[
\overline{E}_n(t, z) = \overline{E}_d(t, 0^+) A(t) \exp(K(t) z)
\] (9)

The mean incident irradiance value computed from IMET mooring pyronometer data for the 30 day sampling period is 189 W m\(^{-2}\) (figure 3), which is roughly 20% less than climatological estimates for the region (Esbensen and Kushnir 1981; Weare et al. 1981; Ohlmann et al. 1996). When combined with the cruise mean transmission parameterization (eq. 8), in-water solar fluxes are 27, 17, 11, and 7 W m\(^{-2}\) at 20, 30, 40 and 50 m respectively. The solar flux which exists at a typical mixed layer depth in the WWP region is significant, and is important in closing mixed layer heat budgets and modeling upper ocean evolution.

In-water solar fluxes are typically parameterized in terms of Jerlov water type, a subjective integer index used to represent upper ocean water clarity (Jerlov 1976). The western Pacific WWP has been classified as Jerlov type II ocean water by Jerlov (1976) and as Jerlov type IB by Simonot and Le Treut (1986). Both of these classifications are based upon optical or Secchi disk data which are more than 20 years old. A recent study by Brainerd and Gregg (1997) suggests the WWP is clearer than Jerlov type I water, and Anderson et al. (1996) assume a Jerlov type IA classification, fairly consistent with our observations. Even with only five open ocean water type
indices (I, IA, IB, II, and III) there is significant disagreement on Jerlov water type for the WWP.

Several solar transmission parameterizations based on Jerlov water type exist (e.g. Paulson and Simpson 1977; Zaneveld and Spinrad 1980; Woods et al. 1984). More recently a set of solar transmission models which parameterize in-water solar fluxes in terms of attenuating material concentrations have been developed (Siegel and Dickey 1987a; Morel 1988, Morel and Antoine 1994; Ohlmann et al. 1996). A comparison among solar transmission parameterizations is illustrated in figure 10. This comparison includes the Paulson and Simpson (1977) double exponential model which is often used in radiant heating applications due to its computational simplicity (despite relying on a Jerlov index) and the Ohlmann et al. (1996) parameterization which has a high degree of spectral resolution (5 nm) and is based on chlorophyll concentration.

The Paulson and Simpson (1977) parameterization for Jerlov type IA water (PS77-IA) shows the closest agreement with the single exponential TOGA-COARE parameterization presented here (eq. 8; figure 10). Values are within 10% between 5 and 35 meters, with the difference increasing to ~20% at 60 meters. The PS77-IA parameterization underestimates in-water solar fluxes as determined from the mean COARE profile by less than 2 W m⁻² at depths between 5 and 70 m (based on a climatological surface irradiance of 220 W m⁻²). Agreements between the mean COARE transmission profile (eq. 8) and the Paulson and Simpson (1977) parameterization for Jerlov water types IB and II (PS77-IB and PS77-II, respectively) are much worse. The PS77-IB parameterization underestimates COARE transmission by 20% at 10 m, and more than 50% at 50 m. These relative errors correspond to absolute flux differences of 9 W m⁻² at 10 m, and 4.5 W m⁻² at 50 m (surface irradiance of 220 W m⁻²). Values of transmission modeled using the PS77-II parameterization underestimate transmission determined from the mean COARE profile by 50%, and 83% (25, and 7 W m⁻²) at 10, and 50 m respectively. The Ohlmann et al. (1996) model underestimates solar transmission determined from the mean COARE profile (eq. 8) by less than 4% between 5 and 25 m, and overestimates transmission at 50 m by 25%. These relative differences correspond to absolute in-water solar flux differences of less than 2.5 W m⁻² for all depths beyond 5 m.
This comparison of solar transmission parameterizations clearly illustrates that significant errors in modeled in-water solar fluxes can arise due to an incorrect choice of Jerlov water type for a particular region. More accurate estimates of in-water solar fluxes require consideration of continuous biological and physical variables which influence the transmission of solar radiation through the water column.

6b. Sensitivity of upper ocean thermal structure to solar transmission

The sensitivity of upper ocean heat content to variations in solar transmission has been established (Denman 1973; Simpson and Dickey 1981; Dickey and Simpson 1983; Martin 1985). Previous studies have addressed idealized cases of dramatically altered attenuation coefficients (corresponding to Jerlov water types) in various one-dimensional mixed layer models. Here, the Price, Weller and Pinkel (PWP: 1986) mixed layer model is used to examine effects of variations in solar transmission on upper ocean thermal structure in the WWP during COARE. Evolution of SST, mixed layer depth, and solar radiation penetrating the mixed layer base are examined using transmission parameterizations determined empirically from the COARE optics data presented here (eq. 7) and the PS77-IB model. The PS77-IB profile represents a "typical" Jerlov water type parameterization for the western equatorial Pacific (Simonot and Le Treut 1986). The PWP model was recently used with TOGA-COARE flux data to assess the role of three-dimensional processes (Cronin and McPhaden 1997) and local precipitation (Anderson et al. 1996) on the Warm Pool heat budget.

The PWP model is one-dimensional and characterizes the mixed layer as a "slab". Mixing occurs through static and dynamic instabilities. The model was originally developed to resolve diurnal cycling in the upper ocean; however, it has performed adequately when integrated over significantly longer time periods (e.g. Anderson et al. 1996, Cronin and McPhaden 1997). The PWP model used here includes both buoyancy and thermal effects of rain (Anderson et al., 1996) and supports daily variations in solar transmission. All model runs were initialized with temperature, salinity and current profiles sampled at the beginning of the simulation period (December 22, 1992) and forced with hourly surface flux averages available from the IMET mooring dataset (Weller and Anderson 1996). The time-step and depth resolution are one hour and one meter respectively. Results are essentially the same
when the model is forced with 15 minute flux data and run with a corresponding time-
step.

The model was initially run with daily averaged solar transmission profiles. This
presumably results in the most accurate solar radiant energy forcing of upper ocean
evolution. For comparison, the model was run using the mean transmission profile for
the pre-chlorophyll bloom period (period 1, Dec. 21 - Jan. 3 : $A(p1)=0.34$,
$K(p2)=0.036 \text{ m}^{-1}$), the mean profile for the post-chlorophyll bloom period (period 2,
Jan. 4 - Jan. 19; $A(p2)=0.37$, $K(p2)=0.052 \text{ m}^{-1}$), and the PS77-IB profile
($A(IB)=0.33$, $K(IB)=0.059 \text{ m}^{-1}$). In all cases, attenuation of energy in the near
infrared wavebands is described by an e-folding value of 1.67 m$^{-1}$ following the PS77-
IA transmission parameterization which most closely matches the COARE profile
(figure 10).

Timeseries of SST from model simulations and direct measurement at the IMET
mooring are shown in figure 11a. Modeled temperatures are relatively insensitive to
variations in solar transmission during period 1, due to the deep mixed layer during this
time (Anderson et al 1996; Smyth et al 1996b, Cronin and McPhaden 1997). Little
solar radiation penetrates deep mixed layer depths regardless of the solar transmission
parameterization used. Differences among modeled SST values determined using the
four solar transmission parameterizations reach only 0.10 C during period 1 (figure
11b-d). Differences in mean values of modeled SST computed for period 1 are less
than 0.02 C.

Modeled temperature shows a significantly greater sensitivity to solar transmission
during quiescent conditions (more typical of the WWP). Differences among modeled
SST values reach 0.83 C during period 2 (figure 11b-d). The greatest difference is
with SST computed using the PS77-IB parameterization (figure 11d). Differences
between period 2 mean SST values computed from model runs using the three solar
transmission parameterizations determined from TOGA-COARE data are less than 0.05
C. The difference between period 2 mean SST computed with daily solar transmission
profiles and the PS77-IB parameterization reaches 0.15 C. This indicates that even a
typical Jerlov water type can result in substantial errors in SST.

Timeseries of mixed layer depth, computed from CTD data and modeled density
profiles are shown in (figure 12a). Mixed layer depth has been computed using a
density criteria of 0.01 kg m\(^{-3}\) from the surface (Peters and Gregg 1987; Anderson et al 1996; Smyth et al. 1996b). In general, the PWP model does a good job of reproducing mixed layer depth. Differences between mixed layer values modeled using the four solar transmission profiles are mostly less than 5 m (figure 12b-d). The mean mixed layer depth for the 30 day period as determined from CTD data is 20.4 m whereas mean modeled mixed layer values are 21.6, 21.8, 21.3 and 20.6 m when the model is forced with daily transmission profiles, the period 1 mean profile, the period 2 mean profile, and the PS77-IB parameterization respectively. These results suggest that increased solar transmission results in increased mixed layer depth. Presumably, increased thermocline heating (solar penetration) aids in the entrainment process by destabilizing thermocline water as previously suggested (e.g. Martin 1985; Anderson et al 1996; Ohlmann et al. 1996; Schneider et al 1996).

Effects of solar transmission variations on the diurnal cycle of mixed layer depth is examined over a four day timeseries (December 27 - 31) extracted from the full record. This period was selected to be sufficiently short as to focus on daily evolution and to be representative of typical warm pool mixed layer depth (30 m; Lukas and Lindstrom 1991). Model outputs give mean values for solar penetration through the mixed layer of 33 W m\(^{-2}\) and 26 W m\(^{-2}\) for the integration period when forced with daily transmission profiles and the PS77-IB profile respectively. The greatest instantaneous difference in penetration between the two parameterizations is more than 30 W m\(^{-2}\). The increased solar penetration with the daily solar transmission profiles results in a mean mixed layer depth increase of 2.3 m compared with the PS77-IB transmission profile. Mixed layer depth differences come during afternoon deepening and last until the beginning of morning stratification (figure 12a).

The cumulative quantity of heat lost to the mixed layer in the form of solar penetration is illustrated in figure 13a. For determination of heat loss from observations, mixed layer depth was calculated for each cast and the \textit{in-situ} net solar flux determined for that depth. Cumulative heat loss from model outputs was computed similarly at each time-step as a function of the incident solar flux, solar transmission and mixed layer depth. Cumulative heat lost over the 28 day period was 155 MJ m\(^{-2}\) (64 W m\(^{-2}\) on average) when directly determined from data. Modeled values of cumulative heat loss are 117, 112, 113, and 124 MJ m\(^{-2}\) (48.4, 46.3, 46.7,
and 51.3 W m\(^{-2}\) on average) when computed using daily transmission profiles, the period 1 mean profile, the period 2 mean profile, and the PS77-IB transmission parameterization respectively. The difference of roughly 40 MJ m\(^{-2}\) (16.5 W m\(^{-2}\) on average) between measured and modeled values is due occasional shallow mixed layer depths appearing in the data.

Cumulative heat loss curves show a change in slope near January 6th (figure 13a). Over the December 21st to January 6th period, the four solar transmission parameterizations give a difference in penetrative heat loss of 7 MJ m\(^{-2}\) (2.9 W m\(^{-2}\) on average; figure 13b-d). During this time, solar penetration is a strong function of attenuation. That is, penetrative heat loss is least when using the PS77-IB solar transmission parameterization for which the attenuation value is greatest, and penetrative heat loss is greatest when using the period 1 profile for which the attenuation value is least (figure 13b-d). This is to be expected for relatively deep mixed layers.

From January 6th to 19th, the difference in heat lost to the mixed layer due to solar penetration among the various solar penetration parameterizations reaches 15 MJ m\(^{-2}\) (6.2 W m\(^{-2}\) on average; figure 13d). This difference is of the order of the one-dimensional heat budget residual calculated for the WWP by Smyth et al (1996b). After January 6th, solar penetration does not appear as a clear function of attenuation. Rather, the cumulative heat loss difference between the daily and period 1 (daily and PS771B) parameterizations becomes increasingly negative (increasingly positive) despite attenuation being less (greater) for the period 1 (PS771B) parameterization (figure 13b,d). This points to the role of the A(t) parameter and the influence of solar penetration on mixed layer depth. For shallow mixed layers, alteration of the solar transmission profile so that transmission is reduced results in a decrease in solar penetration. This increases stratification, and decreases mixed layer depth. The shallower mixed layer results in an increase in the penetrative solar flux through the mixed layer base despite the transmission decrease. Such a feedback mechanism suggests that solar radiation penetration is self-regulating through its effect on mixed layer depth.

By using a hypothetical solar transmission profile which combines the model coefficient, A(p1), and the attenuation coefficient, K(p2), the change in penetrative heat
loss due to the change in the A(t) parameter alone can be isolated. This cumulative heat loss difference is as large as 10 MJ m\(^{-2}\) (4.1 W m\(^{-2}\) on average) over the January 6th to 19th period. The effect of the A(t) parameter decreases significantly as mixed layer depth increases.

This mixed layer modeling study illustrates that even small changes in the solar transmission parameterization (within the range observed during TOGA-COARE) can significantly alter mixed layer heat storage and thermal energy exchange rates. The solar transmission parameterization influences mixed layer heat content through regulation of the solar flux divergence and mixed layer depth. Selection of a solar transmission parameterization corresponding to a discrete Jerlov water type is inadequate. Rather, solar transmission should be based upon optical characteristics of the upper ocean and atmospheric conditions.

7. Conclusions

In-water optical data collected during TOGA-COARE are used to empirically determine a mean solar radiation transmission profile and to characterize solar radiation transmission variability. A single exponential function is sufficient for modeling in-water solar fluxes beyond 10 m. Daily mean solar transmission values determined from empirical single exponential functions are within 10% of directly measured values over the upper 60 m (figure 8). Our COARE data indicates that between 15 and 30 W m\(^{-2}\) penetrate a typical WWP mixed layer of 30 m (based on a climatological surface irradiance of 220 W m\(^{-2}\)). Instantaneous values of solar penetration are often much larger. By failing to properly resolve in-water solar fluxes, modeling efforts can lead to erroneous upper ocean evolution.

The majority of solar transmission parameterizations used today are based on the discrete Jerlov water type index. Solar transmission variations illustrated here cannot be fully resolved with a discrete index. A comparison of solar transmission parameterizations indicates that use of a Jerlov type parameterization for the WWP can lead to errors in the solar flux at 30 m of 15 W m\(^{-2}\) (incident irradiance of 220 W m\(^{-2}\); figure 10). Solar transmission parameterizations based upon continuous measurable parameters must be developed. To first order, variations in solar transmission can be

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explained in terms of upper ocean chlorophyll biomass concentration. While this is not a new finding, it is restated in effort to improve solar transmission parameterizations. Variations in attenuation are most pronounced between 340 and 550 nm where chlorophyll has an absorption peak and CDOM absorption is still relatively large. Thus, variations in the model attenuation parameter, K(t), are primarily due to variations in this spectral range. Clouds can also influence solar transmission by altering the relative composition of the incident irradiance. Clouds are of secondary importance, but can be significant when considering penetration at the shallowest mixed layer depths during periods of relatively constant chlorophyll concentration. A next step is to express the A(t) and K(t) parameters in terms of chlorophyll concentration and a cloud index. This work is underway.

A one dimensional mixed layer model is forced with solar transmission profiles determined from the data presented here, and a profile corresponding to a likely Jerlov water type for the Warm Pool region. Upper ocean evolution shows little dependence on the solar transmission parameterization during the first half of the simulation period when the mixed layer is anomalously deep. During conditions more typical of the WWP (second half of the simulation period), use of the Jerlov based solar transmission parameterization results in temperature differences which reach 0.83 C, compared to use of transmission profiles determined empirically by day (figure 11). Solar transmission can influence mixed layer depth as well as temperature. A change from empirically determined transmission profiles to a Jerlov water type profile results in a near 8% decrease in mean mixed layer depth (figure 12). This shallower mixed layer allows for a solar penetration increase. Through its effect on mixed layer depth, solar penetration is somewhat self-regulating. A decrease in solar transmission can ultimately result in an increase in the solar flux passing through the mixed layer base following increased stratification and mixed layer shallowing.

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References


Figure List

Figure 1. Daily averaged solar transmission, $\overline{\text{Tr}}(t,z) = \overline{E_n}(t,z) / \overline{E_d}(t,0^*)$, calculated from in-situ solar fluxes recorded during TOGA-COARE.

Figure 2. Solar transmission, $\overline{\text{Tr}}(t,z) = \overline{E_n}(t,z) / \overline{E_d}(t,0^*)$, from in-situ solar fluxes recorded during TOGA-COARE. Individual cast data has been binned by meter and extracted at 5 m depth increments for display. The solid line shows the best fit (in a least squares sense) solar transmission profile. Dashed lines are at plus/minus one standard deviation. Standard error estimates are 0.070 and 0.009 for the $A(t)$ and $K(t)$ parameters respectively (eq. 7). Transmission variability at each depth corresponds to the temporal variability shown in figure 1.

Figure 3. Total solar irradiance reaching the sea surface, $\overline{E_d}(t,0^*)$, from data recorded at the TOGA-COARE central mooring (IMET buoy). Two hour averages are plotted.

Figure 4. Ratio of the downwelling spectral irradiance to total downwelling irradiance at the sea surface ($E_d(t,0^*,\lambda)/\overline{E_d}(t,0^*)$) as a function of wavelength and total irradiance. Only ratios determined between 1000 and 1400 local time (solar zenith angle < 40°) are shown to reduce solar zenith angle dependence. Contours are in units of $10^5$ m$^{-1}$. As total solar radiation increases, relatively less energy is contained in the visible wavebands (relatively more in the near-infrared wavebands) for which oligotrophic ocean water is most transparent.

Figure 5. Time-depth evolution of chlorophyll concentration (mg m$^{-3}$) determined from water samples collected twice daily using standard fluorometric techniques. A near doubling of chlorophyll concentration was observed on January 4th which altered the attenuation of in-water solar fluxes.

Figure 6. Spectral diffuse attenuation coefficient contours (eq. 1: m$^{-1}$) at a) 380 nm, b) 490 nm, and c) 565 nm. Values are computed by cast over a 10 m moving window, and daily averaged values plotted.
Figure 7. Spectral diffuse attenuation coefficient contours (eq. 1: m⁻¹) at a) 20 m and b) 80 m. Values are computed by cast over a 10 m moving window, and daily averaged values plotted. The absence of light beyond 575 nm at 80 m prevents the display of attenuation data.

Figure 8. Measured and modeled values of daily averaged solar transmission at 20, 40, and 60 meters. Measured values are computed from $\overline{\bar{T}}_r(t,z) = \overline{E_d}(t,z) / \overline{E_d}(t,0^+)$. Modeled values are computed using single exponential profiles fit to daily data (eq. 7).

Figure 9. Daily values of the solar transmission model parameters (eq. 7), total incident solar irradiance (W m⁻²), depth integrated (0 - 50 m) equivalent chlorophyll concentration (mg m⁻²), cloud index, and ratio of incident irradiance in the resolved spectral region (340 - 683 nm) to total incident irradiance. Total incident irradiance is from IMET data. Integrated chlorophyll concentration values are computed from depth integrated fluorescence and the regression equation for integrated chlorophyll and fluorescence (chl = 0.33*fluo + 0.81; $r^2$=0.74). Cloud index is defined as one minus the ratio of measured incident irradiance to modeled clear sky irradiance.

Figure 10. Comparison among solar transmission profiles as determined from the Paulson and Simpson (1977) parameterization for Jerlov water types IA, IB, and II, the full spectral Ohlmann et al. (1996) parameterization, and the TOGA-COARE cruise mean profile presented here (eq. 8). Transmission and net flux at depth based on an incident irradiance of 220 W m⁻² are shown in figure 10a. Percent difference from the TOGA-COARE parameterization is shown in figure 10b.

Figure 11. Measured and modeled SST, and differences in SST when modeled using various solar transmission parameterizations (see text). a) In-situ values are from IMET data and modeled temperature at 1 m as given by the PWP mixed layer model when forced with daily solar transmission profiles from TOGA-COARE. b-d) Temperature (at 1 m) difference between values modeled using daily solar transmission profiles from TOGA-COARE and values modeled using the part 1 mean profile (b), the part 2 mean profile (c), and the Paulson and Simpson (1977) Jerlov type 1B parameterization (d).
Figure 12. Same as figure 11 for mixed layer depth, defined as that depth where the surface density is exceeded by at least 0.01 kg m$^{-3}$. Mixed layer depth was directly determined from CTD profiles recorded aboard the R/V John Vickers during TOGA-COARE. Modeled mixed layer depth was computed from temperature and salinity outputs.

Figure 13. Measured and modeled values of cumulative heat lost to the mixed layer in the form of solar penetration, and differences in cumulative heat loss when modeled using various solar transmission parameterizations. a) In-situ values from the TOGA-COARE optical data and modeled values given by the PWP mixed layer model when forced with TOGA-COARE daily solar transmission profiles. b-d) Cumulative heat loss differences between values modeled using daily solar transmission profiles from TOGA-COARE and values modeled using the part 1 mean profile (b), the part 2 mean profile (c), and the Paulson and Simpson (1977) Jerlov type 1B parameterization (d).
Table 1. Correlation coefficients for model parameters and independent physical/biological quantities for a) daily mean values, and b) residuals after removal of the pre-chlorophyll increase and post chlorophyll increase temporal means from their respective periods. Bold indicates significance at the 95% confidence level.

**Correlation for daily mean values**

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**Correlation for daily residuals**
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Figure 2

\[ Tr(z) = 0.355 \times \exp(0.045z) \]
Figure 9
Figure 10
Figure 11
Figure 12
Figure 13
Chapter #3

Optical Influences on Ocean Radiant Heating in the Near-Surface Layer

Abstract

Modeled irradiance profiles are used to quantify variations in the transmission of solar radiation through the near-surface layer of the ocean, and to determine physical and biological factors which can be used to explain these variations. Results from a modified version of the HYDROLIGHT radiative transfer model indicate that net irradiance values at 0.01 and 5.0 m can vary by 25 and 37 W m$^{-2}$ respectively, based on a climatological surface irradiance of 220 W m$^{-2}$. Such variations in solar transmission are due to chlorophyll biomass concentration, clouds, and solar zenith angle. Chlorophyll influences solar attenuation in the visible wavebands, but has little effect on transmission within the uppermost meter where the quantity of near-infrared energy is substantial. Clouds serve to increase solar transmission in the upper few meters by causing a greater portion of the incident irradiance to exist in the deep-penetrating, visible wavebands. Increases in solar zenith angle result in decreased solar transmission through changes in surface albedo and the mean cosine of the incident irradiance. Model results are used to develop a strawman radiant heating parameterization for the near-surface layer. This empirical model is expressed as the sum of four exponential terms with model parameters defined as linear combinations of chlorophyll concentration, cloud amount, and solar zenith angle. The model improves the accuracy of irradiance values at depth by more than 10 W m$^{-2}$ when compared to existing parameterizations which are completely invariant or depend only on chlorophyll concentration.
1. Introduction

The thermal and dynamical evolution of the upper ocean is sensitive to the vertical distribution of the solar radiation available for ocean radiant heating (Denman 1973, Simpson and Dickey 1981, Charlock 1982, Kantha and Clayson 1994, Schneider et al 1996, Brainerd and Gregg 1997, Ohlmann et al. 1997). For example, Simpson and Dickey (1981) reported a 0.5 C change in mixed layer temperature over a 24 hour period when solar attenuation was altered from that corresponding to Jerlov type I to that for Jerlov type II water (Jerlov 1976). This suggests the need for models which accurately represent solar radiation attenuation and its spatial and temporal variations (e.g. Ohlmann et al. 1997). Radiant heating parameterizations generally describe solar transmission as a sum of vertically decaying exponential functions, with coefficients for each term determined from Jerlov's water type index (Kraus 1972, Paulson and Simpson 1977, Zaneveld and Spinrad 1980, Paulson and Simpson 1981, Woods et al. 1984). As these parameterizations have evolved, issue has been taken with both the number of exponential terms, and the relationship between Jerlov water type and attenuation values for each term.

Although radiant heating is a continuous quantity and Jerlov water type is a subjective, discrete, integer index, models which relate the transmission of solar radiation to the continuous, measurable, physical and biological quantities on which it depends have only recently been developed (e.g. Morel 1988, Morel and Antoine 1994, Ohlmann et al. 1996). The latest suite of radiant heating models express attenuation parameters for the visible spectral wavebands as a function of chlorophyll concentration (Morel and Antoine 1994, Ohlmann et al. 1996). This is a marked improvement over use of Jerlov water type. Outside the visible wavebands (~400 - 700 nm) which contain roughly half the sun's energy, chlorophyll does not influence solar attenuation (Baker and Smith 1982, Morel 1988). Thus, existing upper ocean radiant heating parameterizations neglect variations within the upper few meters where energy in the near-infrared wavebands is significant.

Some radiant heating parameterizations altogether neglect the near-infrared wavebands which are completely attenuated within the upper few meters. This is satisfactory if concern is with solar radiation penetration on mixed layer depth scales. Other parameterizations attempt to resolve the entire near-infrared spectral region with
only one or two exponential terms. No radiant heating parameterizations currently
address variations outside of those due to changes in Jerlov water type or chlorophyll
biomass concentration. Here, modeled profiles of in-water irradiance are used to
address variations in the transmission of solar radiation over millimeter to meter depth
scales, the scales primarily associated with solar energy in the red and near-infrared
spectral regions. The purpose of this study is to quantify variations in radiant heating
rates for the near-surface layers, and to determine which, if any, physical and biological
factors can be used to explain these variations. Such a study is a necessary step toward
the development of improved near-surface radiant heating parameterizations.

Accurate irradiance measurements for the upper few meters of the ocean do not
exist. However, irradiance profiles can be generated using in-water radiative transfer
models such as HYDROLIGHT (Mobley 1994). HYDROLIGHT computes spectral
radiance profiles and derived quantities such as downwelling spectral irradiance,
upwelling spectral irradiance, spectrally integrated irradiances, and attenuation
coefficients. These parameters are sufficient for determining spectral and spectrally
integrated solar transmission and ultimately radiant heating rates. Transmission values
enable the ocean's influence on in-water solar fluxes to be isolated. Spectral quantities
must be considered for a complete understanding of ocean optical processes and radiant
heating rate variations.

HYDROLIGHT results are used to investigate the roles of chlorophyll, cloud
cover, solar zenith angle, and wind speed perturbations on solar transmission within
the upper few meters. The quantity of light attenuating materials in the upper ocean,
generally inferred from chlorophyll a concentration in open ocean waters, has been
shown to be the primary regulator of in-water solar transmission (Smith and Baker
1978, Siegel and Dickey 1987, Morel 1988, Lewis et al. 1990, Siegel et al. 1995,
Ohlmann et al. 1997). However, these studies are concerned with applications in bio-
optics and radiant heating on mixed layer depth scales where only the deep penetrating
visible wavebands are of interest. The effect of chlorophyll biomass on solar
transmission within the top meter, where a significant portion of solar energy exists
outside the visible wavebands, is not known.

In addition to regulating the quantity of solar radiation reaching the sea surface,
clouds play a role in shaping the relative spectral composition of the incident irradiance
(Nann and Riordan 1991, Ohlmann et al. 1996, Siegel et al. in review). By increasing atmospheric path-length, clouds preferentially attenuate in the red and near-infrared spectral regions. This alters the shape of the incident solar spectrum, putting a relative greater portion of the available solar energy in the ultraviolet and blue-green wavebands. With relatively less energy in the quickly attenuated red and near-infrared wavebands, and relatively more energy in the deep penetrating spectral region, changes in total transmission and the depth distribution of radiant heating occur. A recent study by Siegel et al. (in review) used spectral and total incident irradiance data from the western equatorial Pacific, the SBDART model (Ricchiazi et al. in review), and the Morel and Antoine (1994) in-water solar transmission parameterization to show that the radiant heating rate for the upper 10 cm of the ocean, normalized by the total incident flux, can decrease by 50% in the presence of clouds.

Solar zenith angle can affect solar transmission through the sea-surface albedo and the mean cosine of the in-water irradiance. A dependence of the diffuse attenuation coefficient and diffuse reflectance on sun angle has been illustrated using Monte Carlo simulations (Kirk 1984, Gordon 1989). Wind forcing of the sea surface has been shown to affect in-water radiative transfer through modification of the surface albedo in a few isolated instances (Payne 1972, Simpson and Paulson 1979, Katsaros et al. 1985, Priesendorfer and Mobley 1986). Once identified, the influential parameters can be incorporated into a solar transmission parameterization which resolves the entire solar spectrum and radiant heating variations within the top few meters of the ocean. This will ultimately lead to a parameterization which properly resolves solar transmission variations in the near-surface layer, thus enhancing upper ocean and air-sea exchange modeling efforts.

2) A Modified In-water Radiative Transfer Model

2a) Model Description

HYDROLIGHT directionally discretizes the one dimensional monochromatic radiative transfer equation

\[ \mu \frac{dL(\tau, \xi, \lambda)}{d\tau} = -L(\tau, \xi, \lambda) + \omega_0(\tau; \lambda) \int L(\tau', \xi', \lambda) \hat{\beta}(\tau; \xi' \rightarrow \xi, \lambda) d\Omega(\xi') + S(\tau; \xi, \lambda) \]  

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where \( L(\tau, \xi; \lambda) \) is spectral radiance at optical depth \( \tau \), direction \( \xi \), and wavelength \( \lambda \). 
\( \omega_0(\tau, \lambda) \) is the single scattering albedo defined as the scattering to attenuation ratio, 
\( \hat{\beta}(\tau, \xi \rightarrow \xi'; \lambda) \) is the scattering phase function, and \( S(\tau, \xi; \lambda) \) describes any internal radiance sources (Mobley and Preisendorfer 1988, Mobley 1989, Mobley 1994). The set of all directions, \( \Xi \), is partitioned into a finite set of quadrilateral regions, or quads, enabling \( L(\tau, \xi; \lambda) \) and \( \hat{\beta}(\tau, \xi \rightarrow \xi'; \lambda) \) to be replaced by \( L(\tau; u, v; \lambda) \) and \( \hat{\beta}(\tau; r, s \rightarrow u, v; \lambda) \) respectively, in Equation 1. For this study, 24 lines of constant \( \phi \) (0 ≤ \( \phi \) ≤ 2\( \pi \); 15° intervals) and 20 lines of constant \( \Omega \) (0 ≤ \( \Omega \) ≤ \( \pi \); 10° intervals with 2 polar caps) are used. Such a partitioning scheme is adequate for resolving changes in solar zenith angle and for the introduction of diffuse light due to clouds.

HYDROLIGHT uses invariant imbedding techniques to transform equations for \( L(\tau; u, v; \lambda) \) into a set of Riccati differential equations governing transmittance and reflectance of the radiance field as a function of depth (Mobley and Preisendorfer 1988, Mobley 1989, Mobley 1994). The differential equations are solved for given sea-surface and bottom boundary conditions using a Runge-Kutta method. The ultimate result of HYDROLIGHT is the spectral radiance distribution \( \langle L(\tau; u, v; \lambda) \rangle \) as a function of depth. Integration of radiance over direction gives spectral irradiance as a function of depth. Integration of spectral irradiance over wavelength gives total solar flux at depth.

Inputs to HYDROLIGHT include absorption and scattering properties of the water column, wind speed for determining surface roughness, and the radiance distribution incident at the sea surface. Total absorption and scattering are determined by summing absorption and scattering for the various constituents of the water body of interest. Here, pure water absorption and scattering values from Smith and Baker (1981) are used. The second component of total absorption, that due to particulates, is determined from chlorophyll concentration following the chlorophyll specific absorption parameterization of Morel (1991). This absorption parameterization takes into consideration absorption due to dissolved substances which presumably co-vary with chlorophyll (Prieur and Sathyendranath 1981, Morel 1991). The second component of scattering (due to particulates) is also parameterized in terms of chlorophyll concentration (Gordon and Morel 1983).

The total scattering phase function used by HYDROLIGHT is expressed as
\[ \hat{\beta} = \frac{b_w}{b} \hat{\beta}_w + \frac{b_p}{b} \hat{\beta}_p \]  

(2)

where \( \hat{\beta}_w \) and \( \hat{\beta}_p \) are the pure water and particulate phase functions respectively, \( b_w \) and \( b_p \) are the pure water and particulate scattering coefficients respectively, and \( b \) is the total scattering coefficient. The pure water scattering phase function is from Einstein-Smoluchowski theory (Morel 1994). An average Perzold (1972) phase function is used for characterization of scattering by particulates (Mobley et al. 1993).

HYDROLIGHT simulates the air-sea interface by generating a grid of wave facets from wind speed (Preisendorfer and Mobley 1986, Mobley 1994). The wave slope information and the incident angle of incoming photons are then combined with Snell's law and Fresnel reflectance to track photons at the interface using Monte-Carlo ray tracing. This scheme considers both multiple scattering by wave facets and the possibility of shielding by waves. For the bottom boundary, HYDROLIGHT assumes an infinitely thick homogeneous layer of water with the same optical properties as the water at the maximum depth of interest. Bi-directional radiance reflectance is computed for this layer, and is applied as a bottom boundary condition.

The incident radiance distribution provided to HYDROLIGHT is determined with the Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) model for the entire solar spectrum (250 - 2500 nm; Ricchiazi et al. in review). SBDART combines Mie scattering code to resolve plane parallel cloud reflectance, low resolution band models developed for LOWTRAN 7 to resolve molecular absorption (Pierluissi and Marogoudakis 1986), a standard aerosol model, a Rayleigh scattering component, and the discrete ordinates radiative transfer model of Stamnes et al. (1988). Irradiance values from SBDART have been found to be in good agreement with surface irradiance measurements (Ricchiazi et al. in review).

For this study, HYDROLIGHT was modified to resolve the entire solar spectrum (250 - 2500 nm; the original HYDROLIGHT code works only in the 350 - 700 nm wavelength domain). This requires that absorption and scattering coefficients for the ultra-violet and near-infrared spectral regions be accounted for. Smith and Baker (1981) give clear water absorption and scattering coefficients for the 250 - 800 nm range. Beyond 800 nm, absorption due to pure water is from the imaginary part of the
index of refraction for water obtained from Hale and Querry (1973). The pure water absorption spectrum incorporated with HYDROLIGHT is illustrated in figure 1a. Absorption values determined from the index of refraction for water agree well with the Smith and Baker (1981) coefficients near 800 nm where the two absorption curves join. Scattering coefficients for clear water in the ultra-violet and near-infrared spectral regions are determined using a $\lambda^{-4.32}$ law (Mobley 1994). The pure water scattering spectra used by HYDROLIGHT is illustrated in figure 1b. Scattering coefficients also agree well near 800 nm where they join the Smith and Baker (1981) values. The reference wavelength was arbitrarily chosen to be 600 nm. Changes in the reference wavelength have little affect on the modeled pure water scattering spectra.

Absorption and scattering properties of particulates in the ultra-violet and near-infrared spectral regions are not well known. Particulate absorption in the ultra-violet is determined by linearly extrapolating the chlorophyll-specific absorption spectra of Prieur and Sathyendranath (1981) down to 250 nm. For wavelengths greater than 700 nm, particulate absorption is set to zero (Prieur and Sathyendranath 1981, Morel 1991). Particulate absorption is of little consequence in the red and near-infrared spectral regions where absorption is dominated by pure water. For example, at 700 nm, the largest wavelength for which results were reported by Prieur and Sathyendranath (1981), particulate absorption associated with a chlorophyll concentration of 0.3 mg m$^{-3}$ is 0.010 m$^{-1}$, whereas the clear water absorption value is 0.650 m$^{-1}$ (Smith and Baker 1981). At 750 nm, clear water absorption has increased to 2.47 m$^{-1}$ and particulate absorption is presumably less than 0.010 m$^{-1}$. The particulate absorption spectra used by HYDROLIGHT for various chlorophyll biomass concentrations are shown in figure 1c.

Particulate scattering for the visible wavebands is determined in HYDROLIGHT with the Gordon and Morel (1983) parameterization. A $\lambda^{-1}$ relationship is used to extend particulate scattering values to the ultra-violet and near-infrared spectral regions. Particulate scattering spectra for various chlorophyll values are shown in figure 1d.

2b) Model Comparisons
Comparisons among HYDROLIGHT results, the Paulson and Simpson (1981) solar transmission profile, and the Morel and Antoine (1994) profile indicate the expected range of modeled solar transmission values as a function of depth, and illustrate differences in the way in-water solar fluxes are handled. The profiles selected for comparison are from the only solar transmission parameterizations which resolve the entire solar spectrum. The Paulson and Simpson (1981) model partitions the total solar spectrum into nine wavebands and relies on exponential curves fit to irradiance values measured in pure water (Defant 1961). The parameterization does not allow for any variation in solar transmission. The Morel and Antoine (1994) model focuses on solar transmission variations associated with changes in upper ocean biomass concentrations. The model has 5 nm resolution up to 750 nm, and between 25 and 200 nm resolution out to 2500 nm. Solar attenuation for the ultra-violet and visible spectral regions are determined from summing pure water and particulate components (Morel 1988). Attenuation beyond 750 nm is derived from the imaginary part of the index of refraction for pure water (Hale and Querry 1973) as in HYDROLIGHT.

The Paulson and Simpson (1981) and Morel and Antoine (1994) parameterizations are strictly concerned with the transmission of irradiance within the water column. They eliminate dependence on the air-sea interface by defining solar transmission as

\[ \text{Tr}(z,t) = \frac{E_x(z,t)}{E_x(0^*,t)} \]

(3)

where \( E_x(z,t) \) is the total downwelling solar flux (\( E_d(z,t) \); Paulson and Simpson 1981) or total net solar flux (\( E_n(z,t) \); Morel and Antoine 1994) at depth \( z \) and time \( t \), and \( E_x(0^*,t) \) is the total downwelling (Paulson and Simpson 1981) or total net (Morel and Antoine 1994) solar flux just beneath the surface. Given \( E_x(0^*,t) \) and \( \text{Tr}(z,t) \) values, solar fluxes at depth (\( E_x(z,t) \)) are easily computed. An alternative definition for solar transmission which relies upon \( E_d(0^*,t) \), the downwelling solar flux incident at the sea surface, rather than \( E_x(0^*,t) \) which is not directly measurable, is preferred and will be used later.

A comparison of solar transmission profiles for upper ocean chlorophyll concentrations of 0.03 and 0.30 mg m\(^{-3}\) is shown in figure 2. Each profile is based on the same incident radiance distribution (SBDART: clear sky, 10° solar zenith angle).
Irradiance values just beneath the sea surface, used by the Paulson and Simpson (1981) and Morel and Antoine (1994) parameterizations, are from HYDROLIGHT results. The HYDROLIGHT transmission profile uses total net solar flux. Compared with the Paulson and Simpson (1981) profile, the HYDROLIGHT transmission values are mostly greater despite a chlorophyll attenuation dependency. The difference increases from less than 1% at 0.5 mm to near 10% at 10 cm. Based on a daily average incident solar flux of 220 W m⁻² (Ohlmann et al. 1996), a 10% change in the solar flux at 10 cm corresponds to an absolute flux difference of near 15 W m⁻². The transmission difference due to use of total net solar irradiance, versus total downwelling irradiance, should be small (Gordon 1992).

Differences between the Morel and Antoine (1994) and HYDROLIGHT solar transmission profiles are within 2% in the upper 10 cm and increase to near 20% at 20 m (chl. = 0.3 mg m⁻³). Close agreement within the upper meter is expected as both models rely on the pure water refractive index for determination of attenuation in the near-infrared spectral region (Hale and Query 1973). The decrease in agreement with depth is likely due to the difference in computing attenuation from chlorophyll concentration. When solar transmission profiles are computed for an upper ocean chlorophyll concentration of 0.3 mg m⁻³, the difference remains less than 2% in the upper 10 cm, but increases to 25% at 20 m. A 10% change in the solar flux at 5 m corresponds to an absolute difference of near 6 W m⁻² (for a surface irradiance of 220 W m⁻²).

2c) Model Runs for a Near Surface Radiant Heating Analysis

HYDROLIGHT was initially run with incident radiance for a clear sky and a 10° solar zenith angle, an upper ocean chlorophyll concentration of 0.03 mg m⁻³, and an ocean surface roughened with a 2 m s⁻¹ wind. This case is typical of clear midday conditions in the western equatorial Pacific region and, for ease of understanding, is subsequently referred to as the base case. For comparison, chlorophyll concentration was changed to 0.3 and 3.0 mg m⁻³, solar zenith angle was changed to 35 and 60°, wind speed was changed to 5 and 10 m s⁻¹, and the sky was changed from clear to cloud indices of 0.20, 0.40, 0.60, and 0.90 (table 1). Cloud index is defined as one minus the ratio of the actual to clear sky surface irradiance, so that a cloud index of

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0.20 corresponds to a 20% reduction in the incident irradiance due to clouds. To arrive at the cloud indices used here, optical thickness of the cloud layer at 550 nm was set to 4, 10, 20, and 100 respectively in the SBDART model. HYDROLIGHT was run with 10 nm resolution in the visible spectral region (250 - 700 nm) and 50 nm resolution in the 700 - 2500 nm region. The use of 50 nm resolution for the near-infrared bands resulted in more than a 50% decrease in computing time and less than a 1% change in integrated transmission values compared to complete 10 nm resolution.

The rate at which solar radiation heats an upper ocean layer of thickness \( z \) is

\[
\frac{dT}{dt} = \text{RHR}(z) = \frac{E_n(0^+, t) - E_n(z, t)}{\rho c_p z}
\]

where \( T \) is temperature, \( t \) is time, \( E_n(0^+, t) \) is the total (spectrally integrated) net flux of solar radiation just beneath the sea surface, \( E_n(z, t) \) is the total net solar flux at the base of the layer (depth \( z \)), \( \rho \) is the density of seawater, and \( c_p \) is the specific heat of seawater. Values of \( E_n(0^+, t) \) and \( E_n(z, t) \) can be parameterized from the total incident surface flux \( E_d(0^+, t) \) through solar transmission, defined here as

\[
\text{Tr}(z, t) = \frac{E_n(z, t)}{E_d(0^+, t)}
\]

Use of \( E_d(0^+, t) \) in the transmission definition is desirable as it is easily measured and can be computed from remotely sensed data (e.g. Gautier et al. 1980, Bishop and Rossow 1991, Weller and Anderson 1996). In addition, this definition contains effects of the air-sea interface, or sea-surface albedo (\( \alpha \)), which must be quantified for determining \( E_n(0^+, t) \) in the radiant heating expression (eq. 4). Sea-surface albedo is defined as the ratio of upwelling to downwelling irradiance just above the sea-surface or

\[
\alpha(t) = \frac{E_u(0^+, t)}{E_d(0^+, t)}
\]

Using the net irradiance definition, \( E_n(z, t) = E_d(z, t) - E_u(z, t) \), and conservation of energy across the interface, \( E_n(0^+, t) = E_n(0^+, t) \) albedo can be written as

\[
\alpha(t) = 1 - \frac{E_u(0^+, t)}{E_d(0^+, t)}
\]
and calculated from HYDROLIGHT results. Surface albedo relates to transmission as

$$\alpha(t) = 1 - \text{Tr}(0^+, t).$$  \hspace{1cm} (8)

Although total solar fluxes are of interest for radiant heating calculations, spectral values of solar transmission and sea-surface albedo must be addressed for a complete understanding of radiant heating rate variations. Definitions for spectral transmission and albedo follow those of equations 5 through 8, with the addition of wavelength ($\lambda$) dependence. Specifically, spectral transmission is defined as

$$\text{Tr}(z, t, \lambda) = \frac{E_n(z, t, \lambda)}{E_d(0^+, t, \lambda)}$$  \hspace{1cm} (9)

and spectral albedo follows from eq. 8 as

$$\alpha(t, \lambda) = 1 - \text{Tr}(0^+, t, \lambda).$$  \hspace{1cm} (10)

Both solar transmission and sea-surface albedo have pronounced spectral signatures. A transmission change for the deep penetrating visible wavebands will have different effects on radiant heating rates than a similar transmission change for the near-infrared wavebands which are completely attenuated in the upper few meters. Likewise, an albedo change for the visible wavebands will influence radiant heating rates over greater depths than a corresponding albedo change in the near-infrared. And, since the total incident irradiance contains relatively more energy in the visible than in longer wavebands, an albedo increase for visible wavebands will give a greater reduction in total net irradiance just beneath the sea-surface than a corresponding albedo increase in the near-infrared spectral region.

3) Results

3a) The Chlorophyll Concentration Influence

Total solar transmission, $\text{Tr}(z, t)$, determined from HYDROLIGHT for the base case (table 1: chl = 0.03 mg m$^{-3}$) and changes in solar transmission associated with chlorophyll perturbations (chl = 0.3 and 3.0 mg m$^{-3}$) are shown in figure 3. A ten-fold increase in chlorophyll concentration results in less than a 0.01 change in solar
transmission for the upper meter. A 0.01 transmission change corresponds to a solar flux change of 2.2 W m\(^{-2}\), using a climatological surface irradiance of 220 W m\(^{-2}\). This same chlorophyll increase is responsible for a decrease in solar transmission of more than 0.05 at depths beyond 10 m. A chlorophyll increase to 3.0 mg m\(^{-3}\) results in a solar transmission decreases of less than 0.01 within the upper 10 cm, and more than 0.1 beyond 3 m, corresponding to total solar fluxes of less than 2.2 and more than 22 W m\(^{-2}\) respectively (figure 3).

Transmission spectra for the base case and changes in spectral transmission resulting from chlorophyll concentration increases are illustrated in figure 4. Chlorophyll concentration affects solar transmission only in the ultraviolet and visible wavebands. The depth dependence in solar transmission changes is due to the spectral shapes of both pure water and particulate attenuation curves. Phytoplankton biomass and associated products attenuate solar radiation primarily in the visible spectral region, with absorption peaks near 440 and 680 nm (Kirk 1994). As the solar spectrum narrows and becomes increasingly peaked in the blue-green wavebands with depth (figure 4a), the importance of chlorophyll concentration in regulating in-water solar fluxes increases (figures 4b,c).

Spectral values of sea-surface albedo computed from HYDROLIGHT for the base, chl = 0.3, and chl = 3.0 mg m\(^{-3}\) cases are shown in figure 5a. Spectral albedo peaks above 0.06 at ~400 nm and is mostly near 0.025 beyond 700 nm for the three chlorophyll cases. After increasing chlorophyll concentration two orders of magnitude, spectral albedo increases by less than 0.02 near 550 nm and remains unchanged beyond 750 nm. Spectral albedo decreases slightly between 425 and 450 nm presumably due to the chlorophyll absorption peak in this spectral range. The hundred-fold increase in chlorophyll concentration causes only a 0.016 increase in total albedo, which corresponds to a 3.6 W m\(^{-2}\) decrease in energy available for upper ocean heating (for an incident irradiance of 220 W m\(^{-2}\)). And, changes in available energy are in the visible wavebands which have relatively small impacts on radiant heating rates as visible energy is distributed over relatively large depths. The bio-optical scattering parameterization and particle phase function called upon by HYDROLIGHT can influence spectral and total albedo. However, resulting albedo changes are not significant.
In summary, upper ocean chlorophyll concentration plays a significant role in regulation of solar transmission and radiant heating rates on mixed layer depth scales (cf. Lewis et al. 1990, Morel and Antoine 1994, Siegel et al. 1995, Ohlmann et al. 1996). However, variations in chlorophyll concentration are of little importance when addressing radiant heating for the upper meter, as a significant amount of solar energy exists beyond the chlorophyll sensitive visible wavebands at the shallowest depths. Chlorophyll effects on sea surface albedo are relatively small, and occur only in the visible wavebands.

3b) The Cloud Influence

Changes in total solar transmission from the base case associated with perturbations in cloud amount (cloud index = 0.2, 0.4, 0.6, and 0.9) are shown in figure 6. Total solar transmission increases with cloud index beyond the upper few millimeters as a greater portion of the incident irradiance exists in the deep penetrating blue-green wavebands. Transmission differences increase with depth to near 50 cm and then decrease, indicating a decrease in spectral irradiance (relative to the total irradiance) with e-folding scales of O(10 cm). The total solar transmission difference between the clear sky and 0.20 cloud index cases is negligible (< 0.005) beyond a few millimeters. The transmission difference for the 0.40 cloud index case increases to a maximum of 0.025 near 50 cm, and then decreases to zero at 15 m. For 0.60 and 0.90 cloud indices, solar transmission differences from the base case reach more than 0.05 and 0.14 respectively at 10 cm. Solar transmission changes of 0.05 and 0.14 correspond to total solar flux changes of 11 and 31 W m$^{-2}$ respectively (incident irradiance of 220 W m$^{-2}$).

Radiant heating rate changes due to variations in clouds are more difficult to quantify than transmission. Overall, clouds reduce the incident solar flux and thus the total heat input. However, relative variations in the shape of the incident spectrum and the light field geometry (flux normalized radiance distribution) result in radiant heating rate changes which may not directly correspond to the incident irradiance reduction. Figure 7 shows contours of relative radiant heating rate differences from the clear sky case as a function of cloud index and depth. In the presence of purely white clouds and an invariant albedo, the reduction in radiant heating rate would correspond to the incident solar flux reduction, or cloud index. In this case, contour lines in figure 7
would follow lines of constant cloud index. However, the contour lines indicate that radiant heating rate reductions due to clouds are greater than reductions in the incident irradiance. For a cloud index of 0.40 (a 40% reduction in the total incident solar flux due to clouds), the radiant heating rate decrease for the upper 1 cm is near 55%, and the decrease in the upper 10 cm is greater than 50% (figure 7). In the most extreme cloud index case considered here, a 90% reduction in the incident flux due to clouds results in more than a 95% decrease in the radiant heating rate for the upper 10 cm.

Geometric composition of the incident radiance distribution influences the mean cosine of the light field, sea-surface albedo, and ultimately solar transmission. Total sea-surface albedo values for various cloud indices are given in table 1. Albedo is at a minimum (0.036) for the clear sky low solar zenith angle case and generally increases with increasing clouds. Once the light field becomes sufficiently diffuse (cloud index near 0.40), albedo remains relatively constant near 0.059. Albedo is at a maximum for clear sky high solar zenith angle cases and generally decreases with increasing clouds (not shown; e.g. Payne 1972, Katsaros et al 1985). For low solar zenith angles, solar transmission in the upper few meters increases with cloud index despite the corresponding albedo increase (table 1).

Values of incident spectral irradiance normalized by total irradiance for the base case, and differences in this quantity between the base and cloudy sky cases are shown in figure 8. As cloud index increases, a greater portion of the incident energy (relative to the total irradiance) exists in the deep penetrating visible spectral region (figure 8). For the clear sky case, 65% of the total incident irradiance exists in the 250 to 800 nm spectral region. This value increases to 70 and 83% for the 0.40 and 0.90 cloud index cases respectively.

Figure 5b shows spectral values of sea-surface albedo for the base and perturbed cloud index cases. Spectral albedo for the base case increases with wavelength to a maximum of 0.065 near 400 nm, and mostly fluctuates between 0.02 and 0.03 beyond 600 nm. Changes in spectral albedo with clouds are greater in the red and near-infrared wavebands than in the visible spectral region. Changes in total albedo with clouds come from a combination of changes in spectral albedo values and changes in relative spectral composition of the incident irradiance. It is estimated that total albedo increases
roughly 5% when there is a 10% increase in visible energy relative to the total solar flux.

Clouds influence the mean cosine of both the incident and sub-surface light fields through their light scattering properties. Spectral transmission values mostly decrease with clouds for small solar zenith angles and increase with clouds for high solar zenith angles. Figure 9 shows spectral transmission differences between clear and cloudy sky cases just beneath the sea-surface, at 0.1, 1.0, and 5 m. The spectral transmission difference just beneath the sea-surface is synonymous with the role of albedo (eq. 8). The remaining difference in spectral transmission is due to the angular distribution of the light field. In all cases, surface albedo is responsible for nearly half of the change in transmission. The wavelength of greatest transmission difference decreases with depth, as the solar spectrum narrows. Transmission differences are mostly constant for cloud indices greater than 0.40 as the light field has become sufficiently diffuse.

In summary, clouds reduce the amount of solar energy incident at the sea-surface, thereby reducing total ocean radiant heating. Clouds influence total solar transmission by preferentially attenuating in the longer wavebands, thereby altering the relative composition of the incident irradiance, and by changing the surface radiance distribution through increased diffuse light. Certainly the net reduction in incident irradiance is the primary process in regulating ocean radiant heating. However, changes in the incident radiance distribution can cause an albedo increase, and corresponding transmission decrease, of more than 0.02. With relatively more energy in the deep penetrating visible wavebands under cloudy skies, a 40% reduction in incident irradiance results in more than a 45% decrease in the radiant heating rate for the upper meter.

3c) The Solar Zenith Angle Influence

The role of solar zenith angle (θ) on transmission is similar to that of clouds. An increase in θ results in a net decrease in the incident irradiance, a change in the mean cosine of the incident light field, and a recoloring of the incident spectrum. However, effects of θ on transmission are generally small compared to effects of clouds.
HYDROLIGHT model results show that $\theta$ changes have a significant affect on relative radiant heating rates only in the top centimeter of the ocean under clear skies.

Differences in total solar transmission from the clear sky base case associated with solar zenith angle perturbations ($\theta = 35^\circ$, $60^\circ$, and $60^\circ$ with cloud index = 0.4) are shown in figure 10. An increase in $\theta$ from 10 to $35^\circ$ results in a solar transmission decrease of less than 0.01 over the entire depth range. The transmission differences between the base and $60^\circ$ $\theta$ case are near 0.055 within the upper meter and then decrease with depth. Transmission differences are due primarily to sea-surface albedo. Increasing the solar zenith angle from 10 to $60^\circ$ results in a total albedo increase of 0.051 (table 1). The solar transmission change associated with a $\theta$ increase from 10 to $60^\circ$ corresponds to a decrease in the in-water solar flux of $\sim 13$ W m$^{-2}$ (based on a climatological incident irradiance of 220 W m$^{-2}$).

The primary effect of $\theta$ on upper ocean radiant heating is through regulation of the total incident irradiance. Therefore, radiant heating rates relative to the surface flux must be compared to assess the role of $\theta$ on radiant heating (as with clouds). Figure 11 shows relative decreases in incident solar flux from the base case (solid lines) and relative radiant heating rate differences (dotted lines) for various solar zenith angle cases. Increases in $\theta$ from the base case to 35 and $60^\circ$ result in relative decreases in the incident solar flux of 20 and 59% respectively. If the incident radiance distribution relative to the total irradiance was independent of $\theta$, reductions in radiant heating would correspond to reductions in total incident irradiance. However, when $\theta$ is increased to 35 and $60^\circ$ there are only 19% and 58% reductions respectively in radiant heating rates for the top 1 cm. When solar zenith angle is $60^\circ$ and clouds are added (0.40 cloud index) there is an 81% decrease in the incident solar flux and an 85% reduction in the radiant heating rate for a 1 cm layer.

Differences in spectral composition of the incident irradiance, relative to the total flux, for various solar zenith angles are shown in figure 12. When solar zenith angle is increased from 10 to $35^\circ$ the only significant changes exist in narrow atmospheric water vapor absorption bands which contain little energy. For the $60^\circ$ $\theta$ case there is an additional decrease in energy near 600 nm and additional increases in the near-infrared. For the $60^\circ$ solar zenith angle case, 64% of the incident energy exists between 250 and 800 nm, compared with 65% for the base case. This spectral shift is opposite that for
clouds, resulting in relative increased radiant heating in the uppermost layers. Changes in spectral composition with solar zenith angle are significantly smaller than those associated with clouds.

Spectral albedo for the various solar zenith angle cases is shown in figure 5c. Differences in spectral albedo are slightly greater in the red and near-infrared wavebands than in the visible. This decreases the amount of energy available to heat the upper few centimeters of the ocean relative to the total heat input. Additionally, spectral changes in albedo more than offset radiant heating increases caused by spectral enhancement of the near-infrared region. Differences in spectral transmission values between the base case and various solar zenith angle cases just beneath the surface, at 0.1, 1, and 5 m are shown in figure 13. By comparing spectral transmission differences just beneath the surface \((z = 0')\) with those at depth, the role of surface albedo relative to mean cosine of the in-water light field can be determined. With the exception of spectral narrowing, differences in spectral transmission remain relatively constant with depth. This further suggests that effects of \(\theta\) on transmission are conveyed through albedo.

In summary, solar zenith angle largely influences the total incident solar flux, and thus upper ocean radiant heating rates. Changes in total irradiance are accompanied by changes in the radiance distribution and changes in spectral composition relative to the total flux. Variations in relative spectral composition with increasing solar zenith angle are opposite those for increasing clouds, and small by comparison. Surface albedo is the process primarily responsible for transmission changes with \(\theta\). Total albedo increases by more than 0.05 when solar zenith angle is increased from 10 to 60° (clear sky). Radiant heating rate variations relative to the incident irradiance are significant only within the upper few centimeters and are small compared to changes with clouds.

3d) The Wind Speed Influence

Differences in total solar transmission between the base case and the 10 m s\(^{-1}\) wind speed case for both 10 and 60° solar zenith angles are given in figure 14. Differences for the 10° \(\theta\) cases are less than 0.0025 at all depths. Transmission differences between the 2 and 10 m s\(^{-1}\) cases for a 60° \(\theta\) (difference between the dotted and dashed
lines in figure 14) are much larger, but still less than 0.01, corresponding to an absolute solar flux difference of no more than 2.0 W m\(^{-2}\) (\(E_\theta(0^\circ) = 220\) W m\(^{-2}\)). Total transmission differences are due almost entirely to albedo. Table 1 shows a transmission difference of 0.002 between the base case and the 10 m s\(^{-1}\) wind speed case at 10 cm, and a corresponding 0.002 difference in albedo between the cases. Similarly, the transmission difference between the 60° θ, 2 m s\(^{-1}\) and the 60° θ, 10 m s\(^{-1}\) cases is 0.005 and the corresponding difference in albedo is 0.007.

Spectral sea-surface albedo values for the various wind speed and solar zenith angle cases are shown in figure 5d. Spectral albedo values decrease slightly with increased wind speed for both the 10 and 60° solar zenith angles. Decreases are greatest in the visible spectral region making for little variation in total solar transmission in the near-surface region. Differences in spectral transmission values at 0°, 0.1, 1, and 5 m are shown in figure 15. For a 10° θ, transmission differences peak near 300 nm and decrease with depth. Transmission curves display the same general shape for all depths considered. At 60°, spectral solar transmission differences change significantly with depth. Spectral transmission is greater for the 10 m s\(^{-1}\) case between 250 and 1000 nm just beneath the sea surface. At all other depths, spectral transmission for the 10 m s\(^{-1}\) case is less near 700 nm. This is due to changes in the subsurface radiance field caused by variations in the incident angle of photons relative to the sea-surface slope (e.g. Preisendorfer and Mobley 1986).

In summary, changes in wind speed give rise to variations in the air-sea interface which influence solar transmission within the water column, primarily through sea-surface albedo. Effects of wind speed on solar transmission are greatest during periods of clear skies and large solar zenith angles. However, even under these conditions, wind speed effects are small compared to those of chlorophyll, cloud amount, and solar zenith angle.

4) Discussion

4a) Parameterizing Solar Transmission
Solar transmission profiles for the base case, a chlorophyll concentration of 3.0 mg m\(^{-3}\), a cloud index of 0.60, and a solar zenith angle of 60° illustrate the expected range of transmission variation for the upper 20 meters (figure 16). These transmission values correspond to ranges in total solar fluxes of 25 and 37 W m\(^{-2}\) at 0.01 and 5.0 m respectively, based on an incident irradiance of 220 W m\(^{-2}\). These values are sufficiently large that near-surface variations in solar transmission must be accounted for in upper ocean models.

A primary goal of this study is to identify parameters which can significantly influence ocean radiant heating rates in the near-surface region. These parameters include chlorophyll, clouds, and in the absence of clouds, solar zenith angle. This is not a complete modeling effort in search of the perfect solar transmission parameterization. However, given the current state of full spectral transmission parameterizations, it is worthwhile to use information from the HYDROLIGHT model runs to develop a preliminary parameterization which resolves variations in transmission within the near-surface region where near-infrared radiation is important. The parameterization is of the form

\[
Tr = \sum_{i=1}^{4} a_i \times \exp(-k_i \times z) \tag{11}
\]

with depth (z) positive downwards, and the \(a_i\) and \(k_i\) coefficients defined in terms of chlorophyll, cloud index and solar zenith angle.

This parameterization is developed using nonlinear curve fitting and multivariate analyses in the following manner. A gradient-expansion type algorithm is used to fit exponential curves in the form of equation 11 to each of the 13 transmission profiles generated with HYDROLIGHT results. Exponential fits with four terms explain more than 99% of the variance in each of the profiles. Multivariate least squares analysis is then used to write the fit parameters as linear combinations of the independent parameters (chlorophyll concentration, cloud index, and solar zenith angle; Table 2). Results are essentially the same (differences less than 1%) with and without inclusion of wind speed dependence. Linear fits explain more than 95% of the variance in the model exponents (\(k_i\)'s in eq. 11) and more than 92% of the variance in the coefficients (\(a_i\)'s in eq. 11).
Parameterized values of the $a_i$'s and $k_i$'s (eq. 11) for each of the HYDROLIGHT runs are given in Table 3. The model parameters can be roughly interpreted as representing the fraction of the total surface irradiance ($a_i$'s) and corresponding e-folding depth ($1/k_i$'s). Model results illustrate how the attenuation parameters ($k_i$'s) increase significantly with increased chlorophyll biomass concentration. As cloud index increases, the $a_1$ and $a_2$ parameters increase while the $a_3$ and $a_4$ parameters decrease. This is consistent with the shift in energy to the deep penetrating wavebands discussed above. The $k_i$'s increase slightly with solar zenith angle as the cosine of the light field increases. There is an overall decrease in $a_i$ values with solar zenith angle as increased albedo reduces the fraction of the surface irradiance which passes through the air-sea interface. There is no wind speed dependence.

The parameterization is successful in reproducing transmission profiles generated by HYDROLIGHT. Differences between solar transmission from HYDROLIGHT results and parameterized profiles for the four cases considered above are shown in figure 16b. Transmission differences are mostly within 4%, or less than 0.02. In comparison, transmission differences of more than 0.05 arise when cloud index and solar zenith angle dependencies are ignored (figures 6, 10). Differences between HYDROLIGHT generated transmission and parameterized transmission values correspond to solar flux differences of mostly less than 5 W m$^{-2}$, based on a climatological incident irradiance of 220 W m$^{-2}$. This is a significant improvement over solar transmission parameterizations which prohibit variations within the near-surface region or are based on the subjective Jerlov water type index (Jerlov 1976).

4b) **Sea-Surface Albedo**

Sea-surface albedo studies are generally confined to the air-sea interface and fail to consider solar transmission. Similarly, transmission studies often neglect the role of the air-sea interface (e.g. Morel and Antoine 1994). The HYDROLIGHT results presented here are sufficient for addressing both sea-surface and in-water effects on transmission. In-water effects are isolated by considering the difference between transmission just beneath the surface and that at depth (Tr(0) - Tr(z)). This quantity is subsequently referred to as the “in-water effect”. Influences on solar transmission occurring beneath the air-sea interface which corresponding to perturbations in
independent variables are determined by computing the in-water effect as a function of depth for the 13 HYDROLIGHT cases and subtracting the in-water effect for the base case. Air-sea interface effects on transmission corresponding to perturbations in independent parameters are isolated by subtracting the base case albedo value from albedo values computed for the remainder of the cases (Tr(0') - Tr(0',base case)).

The quantities mentioned above are given in table 4 as in-water effect differences at 10 cm and 5 m, and albedo differences. In general, perturbations in the independent parameters considered here give transmission differences for which in-water and sea-surface effects are of the same order. In-water effects dominate for relatively large chlorophyll and cloud index changes (table 4). Transmission variations associated with large changes in solar zenith angle are primarily due to sea-surface effects, or albedo (table 4). These calculations suggest that regardless of whether sea-surface effects are built into the definition of solar transmission or are handled separately, they can be equally important as in-water effects, and must be properly considered for accurate determination of near-surface transmission.

Sea-surface albedo values have been given as a function of atmospheric transmittance and solar zenith angle by Payne (1972). In his study, Payne defined atmospheric transmittance as the ratio of downward irradiance incident at the sea-surface to irradiance at the top of the atmosphere. HYDROLIGHT and SBDART model results have been used to compute albedo values which can be directly compared with Payne’s results. Overall, values agree well. The greatest albedo difference, 0.009, occurs for the 3.00 mg m⁻³ chlorophyll case for which HYDROLIGHT overestimates Payne’s albedo value by 29%, and for the 0.20 cloud index case where the overestimate is 21%. Values agree to within 10% for all other cases. A 10% change in albedo corresponds to an absolute solar flux change of less than 2 W m⁻², based on a climatological surface irradiance of 220 W m⁻² and an albedo of 0.06 (a widely used value from the Payne study).

The greatest discrepancy with Payne’s (1972) findings is with regard to the quantity of light scattered back from below the sea-surface, or emergent irradiance. Payne indicates emergent irradiance comprises no more than 15% of the total upward irradiance just above the sea surface. HYDROLIGHT results suggest this value can reach 51%. The contribution of emergent irradiance to the total upward flux just above
the sea surface for each of the HYDROLIGHT model runs is given in table 5. These values are based on the assumption that all upwelling photons impinging upon the sea-air interface pass through it, and therefore represent upper bounds.

Knowledge of the spectral structure of albedo is necessary for accurate determination of near-surface solar transmission. The visible portion of the total incident irradiance contains a relatively large portion of the energy, and has a relatively large e-folding value. Therefore, an albedo change in the visible region will have different effects on transmission than a similar change in the near-infrared spectral region. Albedo values for the visible and near-infrared spectral regions are given in table 5. For increased chlorophyll concentration, visible albedo increases by nearly 15% while near-infrared albedo increases by less than 1%. This supports the insensitivity of solar transmission to chlorophyll concentration changes in the upper meter. In contrast, visible albedo increases by 47% and near-infrared albedo increases by nearly 100% when the cloud index is increased to 0.90.

The similar dependencies among transmission and albedo supports the inclusion of albedo within the solar transmission parameterizations. Once included, albedo can easily be backed out (eq. 8). Comparison of albedo values determined directly from HYDROLIGHT results and from the simple parameterization presented here (eq. 8, 11; table 2) gives differences (HYDROLIGHT - parameterized) between -0.02 and 0.015. Parameterized albedo values underestimate directly determined values for all cloud cover cases (\(c_l = 0.2, 0.4, 0.6, \text{ and } 0.9\)) and overestimate for the 35° solar zenith angle case and the increased wind speed cases (\(w_s = 5 \text{ and } 10 \text{ m s}^{-1}\)). Such an albedo parameterization is a significant improvement over use of a single invariant albedo value. The albedo parameterization can be improved upon by increasing the number of HYDROLIGHT cases used in the empirical determination of model parameters. This work is under way.

5) Conclusions

Results from the HYDROLIGHT radiative transfer model indicate that in-water solar fluxes can vary by as much as 40 W m\(^{-2}\) at depth within the upper few meters of the ocean (based on an incident irradiance of 220 W m\(^{-2}\)). Such variations in solar
transmission are due primarily to upper ocean chlorophyll concentration, cloud cover and solar zenith angle. Chlorophyll concentration effects the attenuation of solar radiation within the visible spectral wavebands. Chlorophyll concentration has little influence on solar transmission values within the top meter where near-infrared energy is substantial. Clouds alter the effective solar zenith angle and the spectral shape relative to the total irradiance. Overall, clouds increase solar transmission in the upper few meters by causing a relatively greater portion of the solar energy to exist in the deep penetrating visible wavebands. Total transmission at 10 cm changes by 0.04 when a clear sky is changed to a 0.60 cloud index sky. The influence of solar zenith angle is greatest under clear skies. A change in solar zenith angle from 10 to 60° results in a transmission decrease in the near-surface layer of ~0.05 due solely to the increase in sea-surface albedo.

The few parameterizations which adequately resolve solar transmission within the uppermost layer of the ocean are invariant or depend solely on chlorophyll concentration. A simple empirical parameterization has been presented which defines solar transmission in terms of upper ocean chlorophyll concentration, cloud index and solar zenith angle. Transmission accuracy increases by more than 0.05, compared to use of an invariant transmission parameterization, when HYDROLIGHT generated transmission profiles are reproduced with the parameterization presented here. This corresponds to an improvement in the net solar flux at depth by more than 10 W m⁻² (based on 220 W m⁻² at the sea-surface). This study shows that chlorophyll, clouds, and solar zenith angle must all be considered for the proper parameterization of radiant heating within the top few meters of the ocean. Improvements among relationships between independent and model parameters can be made through more comprehensive radiative transfer modeling studies, or in-situ optical experiments.

Solar transmission has been defined here as the ratio of the net solar flux at depth to the total downwelling solar flux incident at the sea-surface (eq. 5) and thus includes the role of sea-surface albedo. This definition is convenient as it enables determination of \( E_d(z,t) \) from \( E_d(0^*,t) \), a readily available parameter which can be directly measured with a high degree of accuracy or parameterized from remotely sensed cloud data. By defining solar transmission in this way, sea-surface albedo can be parameterized in terms of cloud cover and solar zenith angle, on which albedo has been found to
depend. Comparison of parameterized albedo values with those computed from HYDROLIGHT results gives differences up to 50%, or less than 0.02 in absolute terms. The albedo parameterization presented here is a significant improvement over use of an invariant albedo, often near 0.05. Changes in sea-surface albedo alone can be responsible for transmission variations in the upper few meters of 0.05, or more than 10 W m\(^{-2}\) (for an incident irradiance of 220 W m\(^{-2}\)). The transmission of solar radiation within the upper few meters of the ocean cannot be accurately parameterized with an invariant profile or by relying solely on Jerlov’s subjective water type index.

Acknowledgements. Curt Mobley graciously provided HYDROLIGHT and Catherine Gautier kindly supplied SBDART. Discussions with Bill O’Hirok and Paul Ricchiazi were extremely helpful in guiding this study. Support has been provided by the National Science Foundation (OCE-91-10556).
References


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Figure List

Figure 1. Spectral absorption and scattering values used by HYDROLIGHT. a) Pure water absorption from Smith and Baker (1981; 250 - 800 nm) and the imaginary part of the index of refraction (800 - 2500 nm). b) Particulate absorption for chlorophyll values of 0.03, 0.3, and 3.0 mg m\(^{-3}\) from Morel (1991). Beyond 700 nm particulate absorption is set to zero. c) Pure water scattering from Smith and Baker (1981; 250 - 800 nm) and extrapolated using a \(\lambda^{-1.32}\) relationship (Mobley 1994; 800 - 2500). d) Particulate scattering for 0.03, 0.3, and 3.0 mg m\(^{-3}\) chlorophyll from Gordon and Morel (1983).

Figure 2. Solar transmission profiles from HYDROLIGHT. the Morel and Antoine (1994) full spectral parameterization, and the Defant (1961) 9 waveband parameterization for a) 0.03 mg m\(^{-3}\) chlorophyll and b) 0.3 mg m\(^{-3}\) chlorophyll. The Defant (1961) parameterization is independent of chlorophyll.

Figure 3. a) Solar transmission for chl = 0.03 mg m\(^{-3}\), clear sky, \(\theta = 10^\circ\), wind speed = 2 m s\(^{-1}\) and b) solar transmission differences corresponding to increased chlorophyll concentrations (0.3 and 3.0 mg m\(^{-3}\)).

Figure 4. a) Spectral solar transmission for the base case at 0, 0.1, 1, and 5 m. b - c) Spectral solar transmission differences between the base case and cases for increased chlorophyll concentrations (0.3 (b) and 3.0 mg m\(^{-3}\) (c)).

Figure 5. Spectral albedo for a) the base case and perturbed chlorophyll concentration (chl = 0.03, 0.3, and 3.0 mg m\(^{-3}\)). b) the base case and perturbed cloud index (ci = 0.0, 0.2, 0.4, 0.6, and 0.9). c) the base case and perturbed solar zenith angle (\(\theta = 10^\circ, 35^\circ, 60^\circ,\) and \(60^\circ\) with a 0.40 cloud index). d) the base case and perturbed wind speed (ws = 2 and 10 m s\(^{-1}\) at 10 and 60\(^\circ\)  \(\theta\)).

Figure 6. a) Solar transmission for the base case (chl = 0.03 mg m\(^{-3}\), clear sky, \(\theta = 10^\circ\), wind speed = 2 m s\(^{-1}\)) and b) solar transmission differences corresponding to increased cloud indices (ci = 0.2, 0.4, 0.6, and 0.9).
Figure 7. Contours of radiant heating rate changes relative to the change in incident irradiance, as a function of cloud index and depth. Decreases in radiant heating rates are always greater than reductions in the available energy (surface incident irradiance).

Figure 8. a) Incident spectral irradiance \( \left( E_{d}(0^{+},\lambda) \right) \) normalized by the total incident irradiance \( \left( E_{d}(0^{+}) \right) \) for the base case. This curve illustrates relative spectral composition of the incident solar flux. b) Relative differences in spectral composition between the clear sky base case and cases for varying cloud indices \( (ci = 0.2, 0.4, 0.6, 0.9) \). Cloud index increases are accompanied by a shift in energy from the red and near-infrared to shorter wavelengths.

Figure 9. Spectral transmission differences between the clear sky base case and cases of varying cloud indices \( (ci = 0.2, 0.4, 0.6, 0.9) \) at a) 0°, b) 0.1°, c) 1°, and d) 5 m.

Figure 10. a) Solar transmission for the base case \( (chl = 0.03 \text{ mg m}^{-3}, \text{ clear sky, } \theta = 10^\circ, \text{ wind speed } = 2 \text{ m s}^{-1}) \) and b) solar transmission differences corresponding to increased solar zenith angle \( (\theta = 35, 60\), and 60° with a 0.40 cloud index). \)

Figure 11. Relative decreases in the incident solar flux from the base case for various solar zenith angles \( (\theta = 35, 60\), and 60° with a 0.40 cloud index; solid lines) and relative decreases in associated radiant heating rates from the 10° \( \theta \) base case as a function of depth.

Figure 12. a) Incident spectral irradiance \( \left( E_{d}(0^{+},\lambda) \right) \) normalized by the total incident irradiance \( \left( E_{d}(0^{+}) \right) \) for the base case. This curve illustrates relative spectral composition of the incident solar flux. b) Differences in relative spectral composition between the clear sky base case and cases of varying solar zenith angles \( (\theta = 35, 60\), and 60° with a 0.04 cloud index). The largest spectral changes come with the addition of clouds.
Figure 13. Spectral transmission differences between the clear sky base case and cases of varying solar zenith angle (θ = 35, 60, and 60° with a 0.04 cloud index) at a) 0°, b) 0.1, c) 1, and d) 5 m.

Figure 14. a) Solar transmission for the base case (chl = 0.03 mg m⁻³, clear sky, θ = 10°, wind speed = 2 m s⁻¹) and b) solar transmission differences corresponding to varying wind speeds for low and high solar zenith angle cases (wind speed = 10 m s⁻¹ for 10° θ, and wind speed = 2 and 10 m s⁻¹ for 60° θ).

Figure 15. a,c) Spectral solar transmission for the base case (a) and the 60° θ case (c) at 0°, 0.1, 1, and 5 m. b,d) Spectral solar transmission differences between the base case and the 10 m s⁻¹ wind speed case (b), and between the 60° θ, 2 m s⁻¹ wind speed case and the 60° θ, 10 m s⁻¹ wind speed case (d).

Figure 16. a) Total solar transmission for the base case and cases of perturbed independent variables. These curves show the expected range of solar transmission values. b) Difference between transmission values from HYDROLIGHT results and the parameterization given here.
Table 1  Independent parameter values and solar transmission results for the HYDROLIGHT model runs.

<table>
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<tr>
<th>Case</th>
<th>Chl mg m(^{-3})</th>
<th>Cloud index</th>
<th>Zenith angle deg</th>
<th>Wind speed m s(^{-1})</th>
<th>Tr 0.1 m</th>
<th>Tr 5 m</th>
<th>Albedo</th>
</tr>
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<td>0</td>
<td>10</td>
<td>2</td>
<td>0.657</td>
<td>0.289</td>
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</tr>
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<td>0.289</td>
<td>0.035</td>
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<td>2</td>
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<td>0.298</td>
<td>0.059</td>
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Table 2  Linear regression coefficients for determination of the 8 model parameters used in the solar transmission parameterization presented here. Fit is of the form 
$y = C_1 \cdot \text{chl} + C_2 \cdot \text{ci} + C_3 \cdot \theta + C_4$ where chl is chlorophyll concentration in mg m$^{-3}$, ci is cloud index, and $\theta$ is solar zenith angle.

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<th></th>
<th>$C_1$</th>
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<th>$C_3$</th>
<th>$C_4$</th>
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<tr>
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<td>-9.31e-04</td>
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Table 3  Solar transmission model parameters (eq. 11) determined for each of the HYDROLOIGHT runs given in Table 1 using the linear regression coefficients from Table 2.

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<th>a₂</th>
<th>a₃</th>
<th>a₄</th>
<th>k₁</th>
<th>k₂</th>
<th>k₃</th>
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<td>0.226</td>
<td>0.201</td>
<td>0.154</td>
<td>0.063</td>
<td>0.693</td>
<td>12.01</td>
<td>653.4</td>
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<td>0.648</td>
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<td>0.603</td>
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<td>0.063</td>
<td>0.693</td>
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<td>653.4</td>
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<tr>
<td>case #11</td>
<td>0.374</td>
<td>0.226</td>
<td>0.201</td>
<td>0.154</td>
<td>0.063</td>
<td>0.693</td>
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<tr>
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<td>0.064</td>
<td>0.739</td>
<td>12.18</td>
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<tr>
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Table 4  Independent parameters for each of the HYDROLIGHT model runs, the sub-surface transmission difference from the base case at 10 cm and 5 m, and the albedo difference from the base case. Subsurface transmission differences indicate the increase or decrease in transmission divergence from the base case. This is computed as \([\text{Tr}(0',\text{case } j) - \text{Tr}(z,\text{case } j)] - [\text{Tr}(0',\text{base case}) - \text{Tr}(z,\text{base case})]\) with case \(j\) ranging from 2 to 13 and \(z = 0.1\) and 5 m.

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<th>ci</th>
<th>(\theta)</th>
<th>ws</th>
<th>in-water diff (z=0.1) m</th>
<th>in-water diff (z=5) m</th>
<th>albedo diff</th>
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Table 5  Independent parameters for each of the model runs, the maximum contribution of emergent irradiance, albedo for the visible spectral region, and albedo for the near-infrared spectral region.

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<th>ws</th>
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<th>$E_u(0')$</th>
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Figure 1
Figure 2
Figure 4
Figure 5
Figure 6
Figure 8
Figure 10
Figure 12
Figure 13
Figure 14
Figure 15
Figure 16