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The Hydrodynamics of a Bleaching Event: Implications for Management and Monitoring

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Abstract

This chapter examines the hydrodynamic conditions that are present during a coral bleaching event. Meteorological and climate parameters and influences are discussed. The physics of mixing and its influence on the horizontal and vertical variations of sea temperature are examined. A specialized hydrodynamic model for Palau is then presented as a case study to demonstrate the utility of these models for understanding spatial variations during bleaching events. This case study along with the other sections of this chapter provide the foundation for concluding that hydrodynamic modeling can provide us with a relatively accurate glimpse of the spatial variation of thermal stress and, therefore, what future stress events may hold for corals. Although the timing of a coral bleaching event is unknown and cannot be predicted with current technology, the relative patterns of sea surface temperature during individual bleaching events can be predicted using current modeling techniques. However, improvements in our understanding of coral physiology and higher spatial-resolution climate models are necessary before the full potential of these predictions can be utilized in management decisions.

Introduction

Coral bleaching is a generalized stress response by the coral-zooxanthellae symbiosis and is not necessarily related to any one stressor [Glynn, 1993]. To date, mass coral bleaching events have been correlated with thermal stress [e.g., Dennis and Wicklund, 1993; Drollet et al., 1994; Winter et al., 1998; Hoegh-Guldberg, 1999; McField, 1999; Berkelmans, 2002]. The physiological mechanism is that high temperatures damage the photosynthetic pathway, which leads to a breakdown of the photosynthetic process [Jones et al., 1998]. After the thermal threshold is surpassed, the normally robust photo system can be overwhelmed by significant amounts of light, eventually causing the formation of reactive oxygen molecules that eventually destabilizes the relationship between corals and their symbionts [Hoegh-Guldberg, 1999; Downs et al., 2002]. Therefore, although light is an important factor in the coral bleaching story, it is not normally a stressor until water temperatures have exceeded certain limits [Berkelmans, 2002].

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At present, coral bleaching conditions are monitored in near real-time using satellite-based Advanced Very High Resolution Radiometer data. The U.S. National Oceanic and Atmospheric Administration (NOAA) Coral Reef Watch (CRW) program produces half-weekly, 50-km resolution sea surface temperature (SST) and various derivative products with global coverage (see http://coralreefwatch.noaa.gov). These products provide reef managers and stakeholders with up-to-date information on their jurisdiction. However, physical characteristics of a reef-site that can enhance or mitigate thermal stress are identifiable and, as such, can be used to provide managers with a map defining which regions are more susceptible to thermal stress. This chapter will investigate the origin and spatial variation of the warm water that is known to be a major factor in coral bleaching.

Mass Coral Bleaching: Climate or Weather?

Links between El Niño and Bleaching

There is much talk about El Niño being the "cause" of the 1998 coral bleaching event. If this link exists then we would expect a correlation between El Niño and bleaching around the world. Arzayus and Skirving [2004] took the CRW satellite-derived Degree Heating Week (DHW) product and using the suggested value of DHW = 4 to indicate bleaching, they hindcast bleaching conditions back to 1985 for the entire world. A comparison was then made between the hindcast bleaching conditions and El Niño, La Niña and Neutral states (here referred as ENSO states) as defined by the National Centers for Environmental Prediction's Oceanic Niño Index. (http://www.cpc.noaa.gov/products/analysis_monitoring/ensostuff/ensoyears.shtml).

Arzayus and Skirving [2004] defined that a 50km-square region is correlated with an ENSO state if 70% or more of bleaching events occur during that ENSO state. They found that only 0.2% of bleaching events on reefs are correlated with an ENSO state (0.05% with El Niño, 0.14% with La Niña, and 0.01% with Neutral conditions). Although the effects of different ENSO states on bleaching severity were not examined, it was clearly shown that the onset of bleaching is not correlated with ENSO for the vast majority of world reefs. In fact, the variability of local weather conditions is greater than the climatological means which are used to characterize ENSO states.

Bleaching Weather

Skirving and Guinotte [2001] investigated the origin of the warm water that caused parts of the Great Barrier Reef (GBR) to bleach during 1998. They noted that a combination of low wind speed and neap tides was correlated with high SST. They also noted that during these warm periods there was another correlation between shallow bathymetry and relatively cooler SST.

These correlations led them to conclude that the warm water was a result of local heating from solar radiation in conditions where there was a lack of hydrodynamic mixing. The idea that SST anomalies leading to coral bleaching are mostly a result of local heating has since been supported by many field observations [Wilkinson, 1998; Wilkinson, 2000; Berkelmans et al., 2004; Bird et al., 2004; Skirving et al., 2004].

Very few mass coral bleaching events in the world are a result of advected warm water [Skirving, 2004]. Little to no wind, clear sunny skies and weak ocean currents characterize these events and, as such, local heating is the cause of almost all thermally-induced mass coral bleaching events. It would therefore be more accurate to describe mass coral bleaching as a weather phenomenon rather than the result of climate, as is currently popular. Climate is likely to modulate the frequency and intensity of these weather events, but more research is necessary before direct links between climate states [e.g., El Niño] and coral bleaching can be understood.

El Niño and Weather: A GBR Case Study

The record 1998 GBR bleaching event occurred during the intense 1998 El Niño event and, as a result, many scientists and managers believed that El Niño may be a key component to significant bleaching events on the GBR.

The 2002 coral bleaching event in the GBR was more significant than the 1998 bleaching event in every measurable aspect; SSTs were generally higher, bleaching was more extensive and there was higher mortality [Wilkinson, 2002]. The puzzling thing was that the El Niño did not begin until a number of months after the GBR bleached. This cast doubt on the causal link between El Niño and GBR bleaching events.

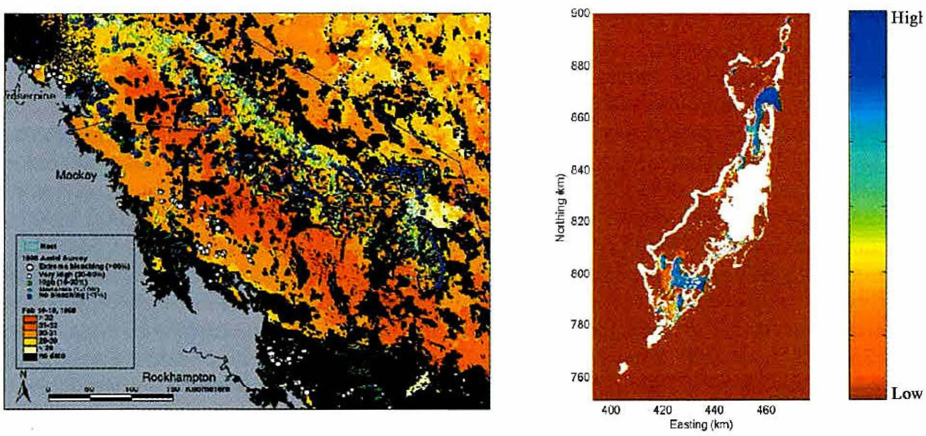
In 2003, the Coral Reef Watch team within NOAA/NESDIS used their DHW and HotSpot satellite products to examine the intensity and accumulated heat stress for the GBR during both bleaching events [Liu et al., 2003]. They noticed that each bleaching event was accompanied by a significant pool of anomalously warm water that covered thousands of square kilometers. The key difference between the 1998 and 2002 bleaching events was not in the intensity of the SST anomaly that caused each event, but in the proximity of the anomaly to the GBR [Liu et al., 2003]. The fact that the 1998 bleaching occurred during a strong El Niño, while the 2002 event occurred outside an El Niño, may be related to the position of the centre of the anomaly. Every significant El Niño event since 1985 has been accompanied by a significant SST anomaly situated off the east coast of Australia, well south of the GBR (see hindcast SST anomaly products on http://coral-reefwatch.noaa.gov/satellite/). The 2002 bleaching event had a smaller, less intense anomaly than that of 1998, but it was situated in the Coral Sea directly off the central GBR. It is this proximity that allowed it to have a far greater influence over local GBR conditions and hence the more significant levels of bleaching observed during 2002. [Liu et al., 2003]

Clearly short- and long-term effects are intertwined. There is a need to monitor local conditions in order to forecast coral bleaching events and also to expect some modulation of the frequency and intensity of bleaching events due to the changing climate.

Spatial Variability of SST During a Bleaching Event

During a bleaching event, spatial patterns of SST are quite complex and have a scale of hundreds to tens of thousands of meters. Plate 1a is taken from Skirving and Guinotte [2001] and is an SST image of the southern GBR during the 1998 bleaching event. It clearly shows the high complexity that existed in the spatial patterns of SST during this event. Skirving and Guinotte [2001] also point out that this bleaching event (like most others around the world) was characterized by bright sunny skies and very low winds. This is generally accepted as a crucial part of the formula for a thermally induced mass coral bleaching event.

One problem with this is that the spatial scales of bright sunny skies (i.e., no cloud) are much larger than the observed variability in SST. This means that we need to look for local



1a **1b**

Plate 1. (a) Average SST for 16-18th February for the Southern GBR region. Reefs and bleaching are also depicted. (from Skirving and Guinotte, 2001). (b) Map of thermal capacitance for Palau, ranging from low (red) to high (blue). (modified from Heron and Skirving, 2004)

conditions which intervene in the causal link between insolation and sea surface temperature and which impose the observed spatial variability in the SST. It has been suggested that, in the absence of wind, hydrodynamic mixing is the only mechanism that could create such a complex SST pattern [Skirving and Guinotte, 2001; Skirving, 2004; Skirving et al., 2004].

Solar energy is absorbed mainly within the top few meters of the water column and, without any vertical mixing, tends to form a stable stratified layer with the warmer water at the top. If there is no vertical mixing then the warm surface layer has the potential to cause coral bleaching. If there is vertical mixing then the temperature of the surface water is reduced and approaches the average temperature of the water column, and the SST condition for bleaching is less likely to occur.

There are four different mechanisms that can vertically mix the water column: wind, low frequency currents [e.g., East Australian Current, Gulf Stream, etc], high frequency currents (e.g., tides) and swell waves.

The effect of winds on the sea surface is to cause surface stress which in turn can form surface gravity waves and also drive surface currents. To a fairly good approximation, low amplitude surface gravity waves are linear and do not lose energy as they propagate. As the wind becomes stronger and the wave heights increase the non-linear effects grow and energy is lost at the surface in micro-breakers, whitecaps and breaking waves. Energy which is lost from the waves ends up as turbulent mixing in the water column. There is little research on the effects of low wind speeds on coral bleaching conditions, and in particular on the effects of current shear caused near the surface by these winds.

Low- and high-frequency currents on continental shelves experience frictional stress at the sea floor, and this is projected through the vertical column as turbulent kinetic energy and vertical mixing. There are also inherent instabilities in water flow when the current speeds are high. Since the water at the surface is warmer than the water below it, complete mixing in the vertical will result in a cooling at the surface and a warming at depth. Further to this, deeper locations would experience a larger reduction in the surface temperature by this mechanism. Therefore, during a bleaching event, there is a relationship between the patterns of SST and a combination of the depth of water and strength of the currents [Skirving et al., 2004; Heron and Skirving, 2004].

Swell can affect the local vertical mixing in the water column. Swell is defined as long wavelength gravity waves generated outside of the local area. Thus, it is quite possible to have swell when there are no local wind waves. The most dramatic effect of swell is on the exposed side of a reef where swell breaks. Some of the energy is transferred into an overflowing bore which carries water onto the reef flat, but most of it is dispersed into turbulence at the reef front. There is a zone at the outer edge of a reef on the exposed side where significant vertical mixing can occur even in the absence of winds and currents.

The Physics of Vertical Mixing

During the daylight hours there is generally a net flow of heat into the ocean. The major part of this flux is in the form of electromagnetic radiation which is absorbed in the upper layers of the ocean. The resulting warm layer may then be mixed down by dynamical processes driven by currents and waves. The outflow of heat through the sea surface dominates at night. The combination of these fluxes gives a net diurnal variation whose long-term mean sets the climatological mean for the sea surface temperature.

Insolation

The solar radiation spectrum at the bottom of the atmosphere peaks in the visible range of wavelengths and has absorption lines and bands due to the composition of the atmosphere. Figure 1 (upper panel) shows typical spectra for the incident solar radiation at the top of the atmosphere and at the sea surface. This a conceptual sketch and for accurate calculations we need to use detailed estimates which take into account the variations in the solar constant and the latitude of the reef. The atmospheric model for absorption in this case changes the total energy from 1353 W/m^2 at the top of the atmosphere [Thekaekara and Drummond, 1971] to 933 W/m^2 at the sea surface. To calculate the atmospheric absorption for a particular site an atmospheric model, such as MODTRAN [Bernstein et al., 1996], can be used. The main control on the energy arriving at the sea surface is exerted by the aerosol and water vapor content of the atmosphere. Therefore, the type of air mass above a coral reef will have considerable influence over the amount of surface solar radiation. A clear sky and low aerosol content, which occur in regions such as the Red Sea and parts of the GBR, will provide maximum insolation. This is evident in the extreme SSTs often recorded in parts of these regions.

Conversely, regions that experience dust storms, high humidity and cloudy conditions will experience lower amounts of insolation, with the same sun angles, resulting in lower SSTs.

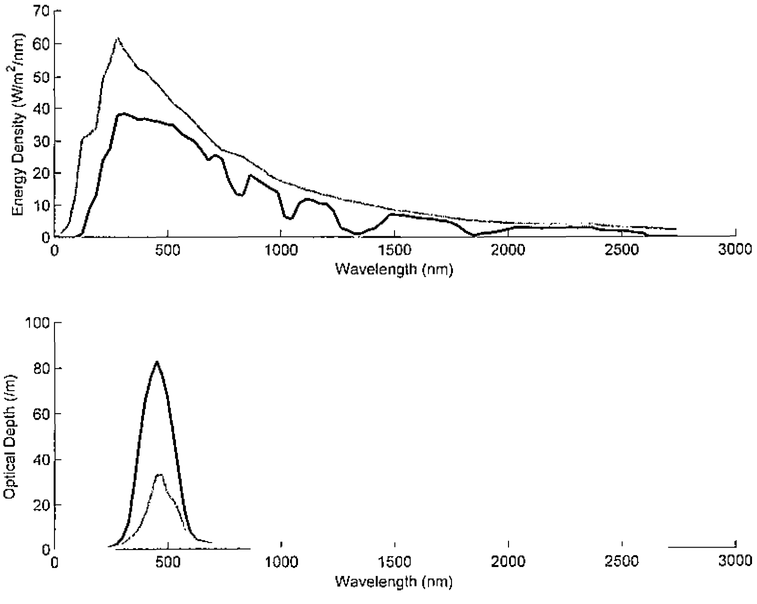


Figure 1. (Upper panel) Stylized graph of solar energy density at the top of the atmosphere (dark grey) and at the sea surface (black). (Lower panel) Stylized optical depth for case I water (black), case II water (dark grey) and case III water (light grey).

Absorption

For a single radiation wavelength and homogeneous water the amount of radiant energy absorbed, ΔI, in an interval of depth, Δz, is assumed to depend on the amount incident on that interval. This can be written as

$$\Delta I = -\alpha I \Delta z \tag{1}$$

where α is an absorption coefficient equal to the inverse of the optical depth, I is the incident energy, and z is the depth. Equation (1) leads directly to the expression for the radiant energy, $I(z)$, at depth z in the water column in terms of the energy I_0 incident at the surface.

$$I(z) = I_0 \exp(-\alpha z) \tag{2}$$

Equations (1) and (2) work for a single absorbing constituent in the water, and for an absorption coefficient which has no variation with depth. This is often not the case. Figure 1 (lower panel) shows how the optical depth ($1/\alpha$) varies with wavelength for three different types of water (case I – extremely pure ocean water, case II – turbid tropical-subtropical water, case III – mid-latitude water [Stewart, 2005; Jerlov, 1976]). This schematic illustrates the point that radiation in the visible band penetrates much further into the water than infra-red radiation which is absorbed very close to the surface, even in the clearest of case I waters. Under these conditions the absorption has to be calculated by summation across all wavelengths for each layer of water. To express this we recognize that the incident radiation varies with wavelength as shown in the upper panel of Figure 1

$$I_0 = \sum_{\lambda} E(\lambda)\Delta\lambda \tag{3}$$

where $E(\lambda)$ is the energy density at the sea surface. Because each wavelength has its own value of optical depth (I/α) we have to calculate the attenuation through the water column to depth z for each wavelength, and then add them up to find the total remaining energy at that depth.

$$I(z) = \sum_{z}\sum_{\lambda} E(\lambda)\exp(-\alpha z)\Delta\lambda\Delta z \tag{4}$$

This is the value plotted in Figure 2 for the insolation and optical depths shown in Figure 1. Note that the curves shown in Figure 2 are not exponential; they are the sum of exponentials with differing optical depths. The most striking feature of Figure 2 is that 90% of the insolation energy is absorbed above 2.32 m, 0.81 m and 0.35 m for case I, case II and case III water respectively. This is juxtaposed with the fact that case I water has an optical depth of up to 80 m at about 450 nm.

For the purpose of discussing the heating effect of insolation energy in the water column, the heating effect is restricted to the top (approx) 1 m in case II water. This clearly demonstrates the importance of vertical mixing if this heat energy is to be removed from the surface layer.

Equation (4) assumes that the optical depth of the water is constant through the water column. If there is any layering of the biomass or sediments then the optical depth (I/α) may vary with depth and this could be put into a modified form of Equation (4).

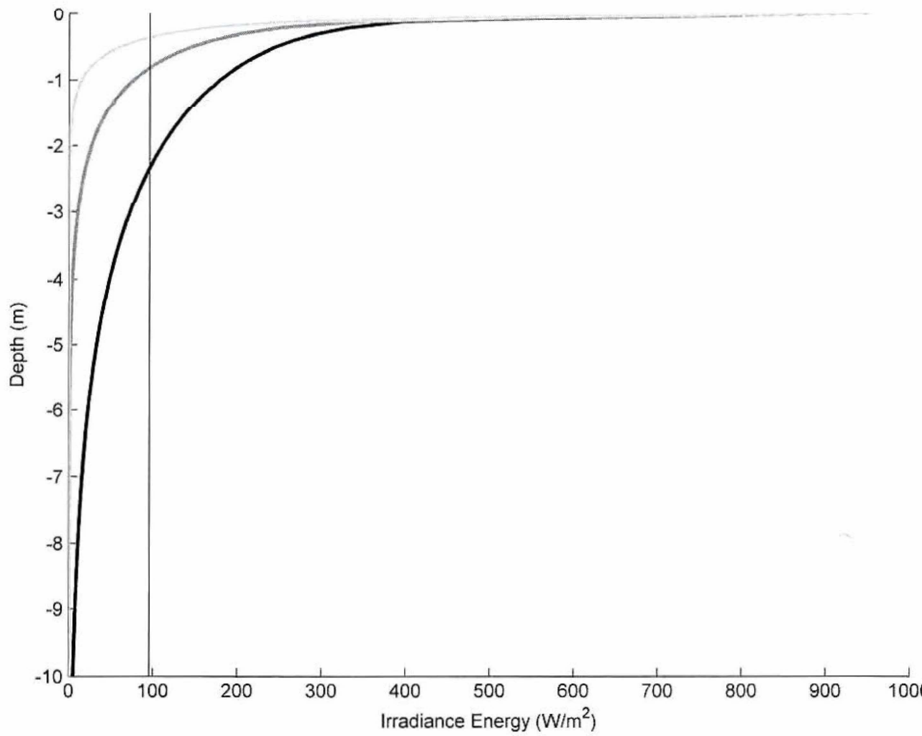


Figure 2. Total radiation energy, integrated over all wavelengths, decreases with depth for the case I water (black), case II water (dark grey) and case III water (light grey) data shown in Figure 1. The thin vertical line can be used to find the depths at which the energy density in the radiation has fallen to 10% of the value at the surface.

Dynamical mixing by currents

Ocean currents have a tendency to induce mixing under most conditions. In shallow water, where we are likely to encounter coral reefs, we can expect a boundary layer shear flow due to friction at the bottom of the water column. This is the basic response to tides and geostrophic forcing, to which we can superpose the effects of wind stress at the surface, stratification and wave-induced mixing. Note that the only concept of laminar flow is in the viscous layer at the bottom, and eddy diffusion prevails throughout the water column.

Mixing due to currents is driven by the vertical shear in the horizontal velocity of the water in the column and is carried out by eddies in the vertical plane. A commonly assumed model for the vertical eddy viscosity, N_z, is the linear model given by

$$N_z = ku_*(h - z), \tag{5}$$

which leads to the logarithmic bottom-friction layer,

$$u(z) = \frac{u_*}{k} \ln\left(\frac{h-z}{z_0}\right), \tag{6}$$

where z is the distance from the surface (positive downwards), h is the water depth, z_0 is the thickness of the viscous layer, u_* is the friction velocity and k is the von Karman constant. The vertical gradient in the horizontal flow has a shearing tendency which induces mixing.

Mixing of the vertical column due to bottom friction is strongest near the bottom where velocity shears are greatest. However with strong currents and shallow water this can impact on the mixing of the upper solar-heated layer. One important thing about this simple theory of mixing in the logarithmic boundary layer is that it gives us a conceptual reference frame for turbulent mixing in the water column when velocity shears are caused by other phenomena.

One such phenomenon is the formation of a horizontal eddy on the lee side of vertical obstructions (reefs, islands, etc.) [see Wolanski et al., 1984]. Such eddies are formed by large horizontal shears between the main flow and the shadow of the structure, where the horizontal turbulence scales are favorable (i.e., appropriate Reynolds number). As this mechanism requires significant currents, it is likely that the bottom-friction mixing associated with the currents will be substantial.

Mixing due to wind stress

Wind at the surface of the sea produces momentum transfer to the water, and hence a wind stress velocity at the surface. The velocity at the surface is transferred down through the column by eddy diffusion. If we assume that the vertical eddy viscosity, controlled by the stress at the surface, grows linearly with depth then we have a mathematical form similar to the bottom friction layer with

$$N_z' = ku_*'z, \tag{7}$$

where z is the distance from the water surface (positive downwards), and u_*' is the stress velocity at the surface.

The actual eddy viscosity in the water column is a combination of N_z' and N_z, and the velocity profile is a combination of the bottom boundary layer and the surface boundary layer. This leads to complications in numerical modeling of the currents and various schemes have been suggested for combining the eddy viscosity terms.

It is clear that the velocity shears induced by wind at the surface of the water have a significant role in the vertical mixing of the surface solar heated layer. Bleaching weather suggests little to no wind; in practice, low-speed winds generally exist and need to be considered.

Stratification

Stratification imposes an impediment to mixing due to the potential energy of the stratification. The Richardson number compares the potential energy (PE) of stratification and the turbulent kinetic energy (KE) and provides us with an index to measure the severity of the stratification. Under these conditions the turbulent kinetic energy works to erode the stratified layer.

Following the approach of Simpson and Hunter [1974], de Silva Samarasinghe [1989], and others, we consider the rate of loss of potential energy to be equal to some small fraction per second, σ, of the turbulent kinetic energy as

$$\frac{\partial}{\partial t}(PE) = -\sigma(KE), \tag{9}$$

which can be written as

$$\frac{\partial}{\partial t}\left\{\int_0^h gz(\rho-\bar{\rho})dz\right\} + \sigma\bar{\rho}\int_0^h N_z\left(\frac{\partial u}{\partial z}\right)^2 dz = 0 \tag{10}$$

per unit area of the water column, where h is the depth of the water column, g is gravitational acceleration, ρ is the density, $\bar{\rho}$ is the mean water density in the column, u is the horizontal velocity, and N_z is the vertical eddy viscosity. Simpson and Hunter [1974] found $\sigma = 0.0037\ s^{-1}$ in the Irish Sea; Hearn [1985] derived a similar value.

Equations 9 and 10 give us important insights into the stability of the upper layer in the water. When the sun heats the surface layer, the density of that layer is reduced, the potential energy (in Equation 9) is increased, and the water column is likely to be inherently stable. When PE dominates there is a high risk of coral bleaching. If the KE term in Equations 9 and 10 is large then mixing is enhanced and the density gradient is eroded. When KE dominates the solar heated water is mixed with deeper water and the risk of coral bleaching is reduced.

An elementary model for predicting the conditions for coral bleaching can use observed or estimated currents (u) and local water depth (h) for gauging the relative importance of the energy terms. This approach is to carry out a calibration (using previous data) of current speed against vertical mixing (or even against the mitigation of bleaching).

If there is wind stress driving currents in the warm surface layer then there is likely to be an enhanced current shear at the bottom boundary of the stratification which can assist mixing.

Another form of stratification is the thin surface layer which is evaporatively cooled by water vapor (latent heat) flux from the ocean to the atmosphere. This is a thin layer of the order of millimeters, with a regeneration time constant of several seconds if it is destroyed, for example, by a micro-breaker [Mobasheri, 1995]. This layer is unstable in the water column and promotes mixing. When we put this micro-layer mixing in the context of solar insolation on the order of a meter depth below the surface it is quickly lost in the scales of energy transfer and penetration depth. A more significant effect of the "skin layer" is that it is this layer which provides the infrared radiation used by satellite radiometers to measure the surface temperature. The skin layer reduces the brightness temperature by up to half a degree-Celsius (and perhaps more in tropical waters). This is not a random error, but is a variable offset in the measured temperature which depends on the nature of the skin layer.

The skin layer has little impact on coral bleaching because it is so thin that it does not contain much heat energy.

Mixing due to wave breaking

In the open ocean, most of the wave energy is conserved and not lost to mixing processes. It is only when the waves become non-linear that they lose energy to turbulence. It is the process of wave breaking that dominates the transfer of wave energy to mixing. At reef fronts the transfer is almost complete with only a remnant of wave energy being reflected back to the ocean, some of it transferring to a forward bore in the breaking wave, and a significant fraction going into turbulence at the breaker location. For a propagating surface gravity wave most of the energy is in the upper part of the water column. This is illustrated in Figure 3 where we show the depth profile of the horizontal surge velocity for a wave with 1 m amplitude (mean to crest) and 6 s period. This is a typical oceanic wind wave and the graph shows how the velocity decreases rapidly with depth. If there is any non-linearity or breaking then the associated energy becomes available for mixing.

The kinetic energy density for a wave with amplitude, a, angular frequency, ω and wavenumber, k is given by

$$KE(z)=\frac{1}{2}\rho(u^2+w^2), \tag{11}$$

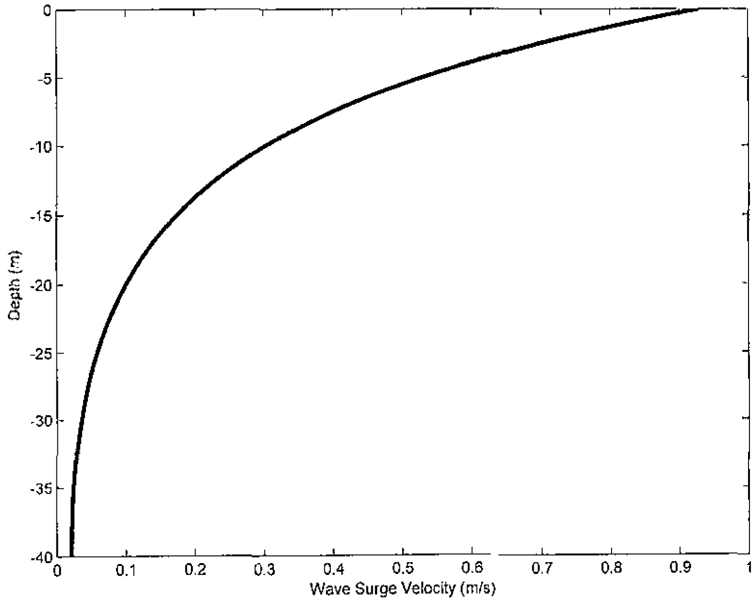


Figure 3. Horizontal surge velocity versus depth for a typical wind wave. Most of the wave energy is in the top few meters. The vertical surge velocity profile follows the same curve near the surface but departs and goes to zero at the bottom of the water column, set to 40 m here.

where u, and w are the instantaneous horizontal and vertical depth-dependent particle velocities:

$$u^2 = \frac{\pi a^2 g^2 k^2}{\omega^2} \frac{\cosh^2(k(h-z))}{\cosh^2(kh)} \tag{12}$$

$$w^2 = \frac{\pi a^2 g^2 k^2}{\omega^2} \frac{\sinh^2(k(h-z))}{\cosh^2(kh)} \tag{13}$$

where ρ is the density, g is gravitational acceleration, h is the depth of the water column and z is the (positive-downward) distance below the sea surface.

This wave energy is generally not available for mixing on shelf waters. However, when a wave encounters a reef front it loses most of its energy and provides a dominant mixing effect for the solar heated layer near the surface. This effect is so dominant that it is difficult to think there would be coral bleaching on the weather side of a reef except in very flat-calm conditions. Waves breaking on the reef front also send pulses of water forward across the reef flat. This pulsing bore is also well-mixed and we would expect mitigation of bleaching on the parts of the reef flat that are flushed with this water.

The physical processes of wave breaking on the reef front and the subsequent pulsing of water across the reef flat have a strong mitigating effect on coral bleaching.

Hydrodynamic Modeling for a Bleaching Event

The physical mechanisms that can influence bleaching are deterministic and, as such, can be modeled to predict the spatial variations in thermal stress. Bleaching weather conditions suggest maximum insolation and no significant wind-induced mixing. The inclusion of a parameter for swell is yet to be done. The effect of swell on mixing depends on the swell direction and the bathymetry of the reef and its surrounds. Swell waves are very effective mixers where they exist and can mix the water when they impinge on a reef. However, they are not capable of cooling an entire reef and will not be available for every reef. Swell may therefore contribute to the variability of coral bleaching on local scales; however, it is yet to be included in the hydrodynamic model presented here.

This leaves currents as the only mechanism which the model considers for altering spatial patterns of SST. The vertical temperature profile is determined by surface heat flux, dominated by solar radiation. Currents then mix this vertical profile via bottom friction and turbulent kinetic energy. The spatial pattern of mixing then modulates the vertical temperature profiles to create patterns of low to high SST during a bleaching event. To date, coral bleaching has occurred in regions of high SST, with the regions of cool water remaining relatively stress free. The case study described in the next section illustrates the success of linking hydrodynamic modeling of currents to the modulation of coral bleaching.

Case study - Palau heat stress model

During the latter half of 1998, Palau experienced unprecedented bleaching that resulted in significant mortality and the loss of significant proportions of one of the few remaining pristine coral reefs in the world [Wilkinson, 2002]. Prior to and since 1998, little to no coral bleaching has been observed. Figure 4 is a plot of accumulated heat stress at Palau as measured by the NOAA Coral Reef Watch DHW satellite product. A DHW value of 4 or more indicates significant bleaching [Liu et al., 2003; Skirving et al., 2006]. Note the DHW = 4 line in Figure 4; the 1998 accumulated stress easily surpassed that mark and is the only year to have done so since 1985. While the DHW product provides a large-scale description of coral bleaching events, it does not describe the smaller-scale variations of thermal stress. An understanding of these variations will lead to improved management.

The Nature Conservancy and the Palau Government joined forces to design and implement a Protected Areas Network (PAN) for Palau's coral reef ecosystem. They recognized bleaching as one of the major future threats to the Palau coral reef ecosystem. However, with only one poorly documented bleaching event, it is hard to understand the response of this ecosystem to coral bleaching and then build resilience to such events into the PAN.

At the same time, NOAA and the Australian Institute of Marine Science (AIMS) were collaborating on the development of hydrodynamic models to predict heat stress during a bleaching event. In 2003, it was decided to combine these efforts and for NOAA and AIMS to produce a heat stress model for Palau for use in the PAN.

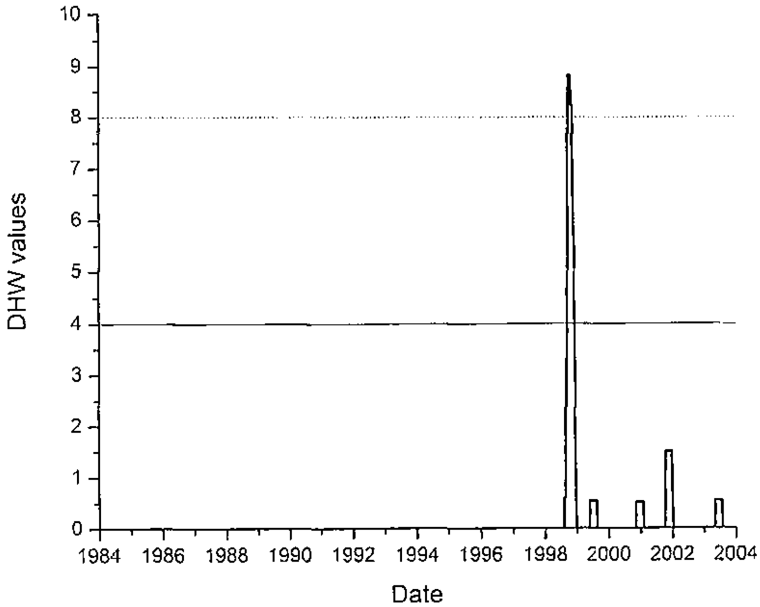


Figure 4. Accumulated thermal coral stress at Palau based on the Degree Heating Week product for the period 1985 to 2004.

For the model to be constructed, NOAA and AIMS needed [Skirving et al., 2005]:

1) The Palau bathymetry: Due to a lack of available data, NOAA derived the bathymetry from a combination of Landsat data and bathymetric transects taken with a depth sounder from a small boat. This produced a chart with 256.5 meter horizontal resolution and an rms error of approximately 1 meter vertically.
2) Low frequency currents: The Naval Research Laboratory [NRL] Layered Ocean Model (NLOM) and NOAA's Ocean Surface Current Analyses – Real time (OSCAR) were used to derive the seasonal low frequency currents around Palau.
3) High frequency currents: A combination of tide gauge data collected in and around Palau and a global tidal model was used to derive a model that could accurately predict the tides. The bathymetry and the tidal model were then used with the Princeton Ocean Model (POM) to build a model of tidally induced currents in and around Palau. Field data collected over a period of 5 months were used to calibrate and validate the output of this model.
4) Vertical temperature profile: This was derived by modeling a patch of water with a homogeneous temperature and applying a diurnal cycle of solar radiation and a constant, low wind of 2.6 m/s for a period of two weeks.

Simpson and Hunter [1974] provide the parameterization that was used to distinguish between stratified and well-mixed water by combining the currents with the bathymetric data. This information was then used in conjunction with the vertical temperature profile to determine the likely spatial distribution of sea surface temperature. To account for advection and tidal variation, while providing a static image for use in the PAN, the surface cooling due to mixing was accumulated over one tidal cycle (two spring-neap cycles). This accumulation parallels the DHW approach used by NOAA/CRW, with units of temperature-time.

This model is better described as a measure of thermal capacitance than as accumulated cooling. Thermal capacitance is the ratio of heat absorbed to the resultant temperature rise. Areas of low thermal capacitance will exhibit a larger increase in temperature for a given amount of heat input than areas of high thermal capacitance. Regions with complete vertical mixing (greater accumulated cooling) have high thermal capacitance, while stratified regions (lesser or no accumulated cooling) have low thermal capacitance.

The result of this is that the well-mixed regions in the Palau model represent regions of mild thermal climatology (i.e., less temperature variation and hence less thermal stress), whereas the stratified regions represent those areas that will experience the most extreme temperature range (i.e., greater thermal stress). Plate 1b is a thermal capacitance map derived from the Palau model; the blue regions have high capacitance (well-mixed, mild climate), while the red regions have low capacitance (stratified, variable climate).

A chart of this type can be extremely useful when designing a PAN. In general, most PANs are currently designed so as to provide protection to "representative bioregions". This means that as much as possible, every type of bioregion within the ecosystem of interest should be equally represented within the PAN. However, an ecosystem is not only made up of different species. It is also important to recognize that an ecosystem is made up of organisms that have unique physiological characteristics within each species. It is adaptation to the local climate (mild or variable) that will define these physiological characteristics.

When designing a PAN, it is relatively straight forward to map bioregions on the basis of species composition; however, the unique physiological properties within each species are not represented within these techniques. These physiological characteristics are likely grouped into areas that mimic the relative thermal capacitance through the region. As such, even without knowledge of the individual characteristics, it is possible to incorporate the spatial variation of physiology by means of the thermal capacitance.

When considering the ecosystem response to an individual thermal stress (bleaching) event, it is the regions of high thermal capacitance which will moderate the temperature rise, thereby experiencing lesser thermal stress. For this reason, high-capacitance areas should be selected for protection. It is important to note, however, that during extreme thermal stress events (such as was seen in Palau in 1998) even these areas may experience bleaching conditions.

When considering the sustainability of the ecosystem with respect to long-term climate change, it is the low thermal capacitance regions which should be protected. Low-capacitance regions experience variable climates exposing organisms to extreme temperatures and, thus, frequent periods of thermal stress. This exposure may lead to increased resilience to rising temperatures, via physiological adaptation, and aid their survival as climate change occurs. Representing equal proportions of high and low thermal capacitance areas is necessary and ideal for the short- and long-term protection of coral reefs.

Conclusion

Hydrodynamic modeling can provide us with a relatively accurate glimpse of what future stress events may hold for corals. However, for the full potential of the model to be realized and employed in management, additional work is required. Advancements in coral physiology to determine the response of organisms to thermal stress, including issues of acclimation and adaptation, will improve management during the existing climate regime. In addition to these, the inclusion of improved climate models (e.g., in spatial resolution and accuracy) would allow more accurate predictions of future bleaching events.

A careful examination of the facts surrounding the physical conditions during thermally-induced mass coral bleaching events leads to a few surprising conclusions:

1) Mass coral bleaching is a weather event and is not necessarily linked with climate.
2) In general, the vast majority of coral reefs around the world are not predisposed to bleaching during an ENSO.
3) Twice as many of the world's reefs bleached during the 1998-99 La Niña than during the 1997-98 El Niño.
4) Bleaching weather is characterized by cloudless (sunny) skies, low to no wind and low currents.
5) 90% of the sun's energy is absorbed in about the top 2 meters of the water column.
6) The SST patterns during a bleaching event are dominated by spatial variations in vertical mixing.
7) The hydrodynamic processes that cause mixing and hence create most of the SST patterns during a bleaching event are largely predictable.

In general, this means that although the timing of a coral bleaching event is unknown and cannot be predicted with current technology, the relative patterns of SST during the next bleaching event can be predicted using current techniques for hydrodynamic modeling. Hydrodynamic modeling, when combined with an improved knowledge of coral physiology, can go a long way to helping us understand the exact nature of mass coral bleaching, allowing for improved monitoring and predictions.

Implications for Management

Although the vertical profile of temperature can change from event to event, the mixing parameters change very little as the tidal and low frequency currents are cyclic and thus effectively predictable for any specific location. These parameters can be used to identify, using hydrodynamic modeling techniques, the variation of thermal capacitance across the region of interest. The result is that the SST pattern during a severe bleaching event is effectively static from one bleaching event to another, with the magnitude of temperature related to the input heat. Identification of these patterns allows a higher degree of management of coral reefs prior to and during the onset of thermal stress events.

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