Oceanic Heat Content Variability in the Eastern Pacific Ocean for Hurricane Intensity Forecasting

LYNN K. SHAY AND JODI K. BREWSTER
Division of Meteorology and Physical Oceanography, Rosenstiel School of Marine and Atmospheric Science, Miami, Florida

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ABSTRACT

Recent evidence supports the premise that the subsurface ocean structure plays an important role in modulating air–sea fluxes during hurricane passage, which in turn, affects intensity change. Given the generally sparse in situ data, it has been difficult to provide region-to-basin-wide estimates of isotherm depths and upper-ocean heat content (OHC). In this broader context, satellite-derived sea surface height anomalies (SSHAs) from multiple platforms carrying radar altimeters are blended, objectively analyzed, and combined with a hurricane-season climatology to estimate isotherm depths and OHC within the context of a reduced gravity model at 0.25° spatial intervals in the eastern Pacific Ocean where tropical cyclone intensity change occurs.

Measurements from the Eastern Pacific Investigation of Climate in 2001, long-term tropical ocean atmosphere mooring network, and volunteer observing ship deploying expendable bathythermograph (XBT) profilers are used to carefully evaluate satellite-based measurements of upper-ocean variability. Regression statistics reveal small biases with slopes of 0.8–0.9 between the subsurface measurements compared with isotherm depths (20° and 26°C), and OHC fields derived from objectively analyzed SSHA field. Root-mean-square differences in OHC range between 10 and 15 kJ cm\(^{-2}\) or roughly 10%–15% of the mean signals. Similar values are found for isotherm depth differences between in situ and inferred satellite-derived values. Blended daily values are used in the Statistical Hurricane Intensity Prediction Scheme (SHIPS) forecasts as are OHC estimates for the Atlantic Ocean basin. An equivalent OHC variable is introduced that incorporates the strength of the thermocline at the base of the oceanic mixed layer using a climatological stratification parameter \(\sqrt{N_{\max}/N_o}\), which seems better correlated to hurricane intensity change than just anomalies as observed in Hurricane Juliette in 2001.

1. Introduction and background

Based on deliberations of the Prospectus Development Team 5 tasked by the National Oceanic and Atmospheric Administration (NOAA) and the National Science Foundation (NSF), improving our understanding of hurricane intensity requires knowledge of the 1) atmospheric circulation, 2) inner-core and eyewall processes, and 3) upper-ocean circulation and ocean heat transport (Marks et al. 1998). While the oceanic energy source for hurricanes has largely been known for more than half of a century (Palmen 1948; Fisher 1958; Perlroth 1967; Leipper 1967), subsequent studies indicate that the maximum hurricane intensity is constrained by thermodynamic effects where the sea surface temperature (SST) is considered an important contributor (Miller 1958; Emanuel 1986).

Research on the SST response to hurricane passage has been largely focused on the “negative” feedback aspects of how a cooled upper ocean affects the atmosphere by decreasing air–sea fluxes as the oceanic and lower atmospheric boundary layer temperatures approach an equilibrium value (Chang and Anthes 1978). As the hurricane strengthens, winds induce more stress on the ocean mixed layer (OML) causing strong turbulent mixing across the base of the layer on the right side of the track and upwelling of the thermocline due to net wind-driven current transport away from the storm center along the track (Price 1981; Gill 1982; Sanford et al. 1987; Shay et al. 1992). Shear-induced mixing effects deepen and cool the OML as colder thermocline water is entrained from below. This entrainment heat

Corresponding author address: Lynn K. Shay, Division of Meteorology and Physical Oceanography, Rosenstiel School of Marine and Atmospheric Science, 4600 Rickenbacker Causeway, Miami, FL 33149.
E-mail: nshay@rsmas.miami.edu

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flux subsequently causes the OML temperature (and by proxy the SST) to decrease, limiting air–sea heat and moisture fluxes that reduce intensity. This negative feedback mechanism is particularly effective when the OML depths are shallow or when storms become stationary or move slowly. By contrast, in regimes where the OML (and the depth of 26°C water) is deep, the ocean heat content (OHC) can be quite large (Leipper 1967; Leipper and Volgenau 1972; Shay et al. 2000). As this deeper OML has greater vertical extent, more turbulence is required to entrain colder water to cool these deeper layers. A well-studied example of this effect is the response of Hurricane Opal (1995) that intensified rapidly as it crossed a warm core ring in the Gulf of Mexico (GOM; Shay et al. 2000) under favorable atmospheric conditions (Bosart et al. 2000). When Opal encountered this deeper, warmer oceanic regime, the storm unexpectedly intensified from a category 1 to a category 4 hurricane status in 14 h as atmospheric conditions were favorable. Sensitivity studies with a coupled ocean–atmosphere model (Hong et al. 2000) showed that the central pressure of Opal was more than 10 hPa higher when the warm core ring (WCR) was removed in the prestorm state. Using a hurricane-season (June–November) climatology, Mainelli (2000) extended the Opal investigations across the Atlantic Ocean basin including the Caribbean Sea and the GOM and introduced a hurricane-season climatology instead of an annual climatology (Goni and Trinanes 2003).

More recent examples included both Hurricanes Katrina and Rita encountering the Loop Current (LC) and WCR in the central GOM in 2005 (Scharroo et al. 2005; Shay 2009; Jaimes and Shay 2009). For both hurricanes, Shay (2009) showed that the sea level pressure decreases were directly correlated to large values of the 26°C isotherm depth and OHC than just SSTs, which were essentially flat. Using the Statistical Hurricane Intensity Prediction Scheme (SHIPS; DeMaria et al. 2005), Mainelli et al. (2008) found that the OHC parameter contributed an average of 5%–6% to the reduction in intensity forecasting using the seasonal climatological approach. Lin et al. (2009) showed that the translation speeds of typhoons may impact its intensity because for fast movers (>6 m s⁻¹) only 60 kJ cm⁻² are necessary for intensification to severe status in the western Pacific Ocean basin compared to more than 100 kJ cm⁻² for slower-moving typhoons. These findings support the premise that oceanic regimes with high OHC contribute to hurricane intensification where SST cooling is reduced (e.g., less negative feedback) beneath the storm and maintaining or enhancing surface sensible and latent heat fluxes.

Consistent with this framework, recent improvements to SHIPS have shown that OHC parameter has reduced intensity forecast errors in the Atlantic Ocean basin (DeMaria et al. 2005; Mainelli et al. 2008). These studies have shown that a seasonal OHC climatology reduced forecast errors in intensity by an average of 2% when averaged over all storms between 1995 and 2003 compared to less than 1% from an annual analysis (Goni and Trinanes 2003). Notwithstanding, if only the western part of the Atlantic Ocean basin is used (60°–100°W), intensity errors are reduced from 4% to 6% when averaged over all storms. In some cases, the reduction in intensity errors is considerably more dramatic as observed during Hurricane Ivan in 2004. SHIPS with seasonal OHC showed as much as a 22% reduction in forecast intensity errors (Mainelli et al. 2008). This result underscores the importance of OHC variability in the warm pool of the Caribbean Sea and the Gulf of Mexico’s LC and WCR as well as other eddy-rich regimes such as the western Pacific and Indian Oceans (Lin et al. 2005; Ali et al. 2007).

The eastern Pacific Ocean basin (hereafter referred to as EPAC) is also a region of significant upper-oceanic variability given the warm pool centered at 10°N, 95°W and gradual shoaling of the oceanic thermocline from west to east, and the westward propagation of WCRs forced either by low-level jets through mountain gaps (Hurd 1929; Kessler 2002) or current instabilities (Hansen and Maul 1991; Fig. 1). This warm eddy feature moved southwestward at 13–15 cm s⁻¹ and dissipated within the Eastern Pacific Investigation of Climate (EPIC) domain in late October 2001 (Fig. 2). Over this oceanic regime, tropical cyclogenesis often begins at between 10°–15°N and 90°–100°W, which is characterized by SST gradients and OHC variations that impact hurricane intensity change (Raymond et al. 2004). Of particular importance to the maintenance of the SST and OML temperature structure are the sharp thermal gradients across the OML base starting about 30–35 m beneath the surface and the 20°C isotherm separates the upper from the lower layer in a two-layer model. This approach is applied to Hurricane Juliette in September 2001 (Shay and Jacob 2006), and its subsequent intensification to category 4 status, the SST cooling was less than 1°C in this warm regime (Raymond et al. 2004).

This investigation updates the original approach by Shay et al. (2000) and Mainelli (2000) in developing a seasonal hurricane climatology to monitor isotherm depths and OHC from multiple radar altimetry platforms. This approach differs from that of Goni and Trinanes (2003) in that they formed an annual climatology to represent a global tropical cyclone (TC) heat potential throughout the entire year. Since TCs only occur in specific seasons in both the northern (boreal summer) and Southern Hemisphere (boreal winter),
upper-ocean thermal energy is smeared over an entire year compared to a TC season that may lead to significant over- (under) estimations in the southern (northern) parts of the domains compared to thermal energy available to TCs in each basin.

The $20^\circ$C isotherm depth is used as the level for a two-layer model to estimate reduced gravities (i.e., density differences between the upper and lower ocean layers; O’Brien and Reid 1967; Kundu 1990; Goni et al. 1996). Radar altimetry data are merged and blended each day when new sea surface height anomaly (SSHA) data become available. Two or three sets of altimeter data are then objectively analyzed to the same grid as a hurricane-season climatology derived from U.S. Navy’s Generalized Digital Environmental Model (GDEM) for the two-layer model application (Teague et al. 1990). The resultant isotherm depths (particularly the $26^\circ$C isotherm) are used to estimate OHC when combined with SST and a climatological OML depth over the hurricane season. Isotherm depths and OHC estimates are carefully compared to temperature structure measurements from the Tropical Atmosphere Ocean (TAO) mooring data spanning the Pacific Ocean equatorial waveguide, the EPIC field program (Weller et al. 1999; Raymond et al. 2004), and volunteer observing ship expendable bathythermograph (XBT) transects to build an evaluated-hurricane-season climatology in the EPAC (May–November).

Finally, a stratification parameter is introduced to provide a measure of basin-to-basin OHC variability and the strength of the vertical density gradients to form an
equivalent OHC. Data resources are described in section 2 and the approach is given in section 3. Detailed comparisons and assessment of the altimeter-based isotherm depths and OHC estimates are discussed in section 4. We introduce the concept of comparing estimates from basin-to-basin for possible use in forecasting in section 5 with an application for Hurricane Juliette (September 2001) followed by a summary and concluding remarks in section 6 including prospects for future improvements.

2. Data resources

In this section, data resources are described for developing a hurricane-season climatology for the EPAC as in the Atlantic Ocean basin (Mainelli 2000; Mainelli et al. 2008). The TAO moorings (see Fig. 6) relative to the EPIC domain are superposed on a September 2001 SST image. These TAO moorings were originally deployed as part of the Tropical Ocean and Global Atmosphere program in the 1990s to monitor the equatorial waveguide and were enhanced at 95°W during EPIC (Cronin et al. 2002).

a. Radar altimetry

The Ocean Topography Experiment (TOPEX)/Poseidon (T/P) and Jason-1 radar altimeters measure sea level every 9.9 days along repeat ground track spaced 3° longitudinally at the equator. The European Remote Sensing Satellite-2 (ERS-2) mission and U.S. Navy Geosat Follow-On-Missions (GFO) have repeat tracks of 35 and 17 days, respectively. More recently, Envisat data has been added since the ERS-2 mission is no longer providing data. The availability of such altimetry measurements is shown in Fig. 3 for several satellites that carry altimetry sensors. For example, the National Aeronautics and Space Administration (NASA) TOPEX mission began in 1992 (Cheney et al. 1994; Ali et al. 1998) followed by Jason-1 in 2000 as the follow on to the TOPEX mission. As noted in Fig. 1, altimetry data are useful in tracking oceanic mesoscale features such as warm and cold features as observed during the EPIC field program from August to October 2001 using a 7-day Archiving, Validation, and Interpretation of Satellite Oceanographic data (AVISO) multimission product of altimeter data, developed by Collecte Localisation Satellites as part of the Developing Use of Altimetry for Climatic Studies. The ring pathway, tracked over a 3-month period based on successive SSHA images, suggests that the warm feature moved at speeds of 13–15 km day$^{-1}$ toward the west southwest as its diameter decreased during its spindown process in the EPIC domain. Tracking of oceanic features has been done in other regions such as the GOM to study WCR shedding processes from the LC (Leben 2005).
Although the 10- and 17-day repeat cycles utilized in the OHC estimates are usually long compared with the hurricane life cycles, they are reasonable compared to the time scales of upper-ocean variability (Shay et al. 2000), such as the ocean rings being analyzed. Moreover, this analysis system is designed to be updated daily to incorporate the latest altimetry measurements from all sensors including Envisat, which has further improved the OHC estimates, particularly in areas where the signal-to-noise (SNR) ratios are large. This is particularly important for hurricane intensity studies since track-dependent ocean features affect intensity changes (Shay et al. 2000; DeMaria et al. 2005; Mainelli et al. 2008). For example, during August and September 2005 (the Katrina and Rita time period), comparisons of the updated 1-day product from multiple platforms to the AVISO product revealed correlation coefficients of more than 0.92, regression slopes of O(1), and biases of 2–5 cm in the GOM basin (Jaimes and Shay 2009).

b. Objective analysis

In the forthcoming analysis, only periods when there are at least two sets of radar altimeter data available are used here. These SSHA data are objectively mapped to a 0.25° grid using the approach of Mariano and Brown (1992). This oceanic analysis decomposes a scalar observation into three components using parameters derived from the Hurricane Gilbert dataset (Shay et al. 1992). The first component of this scalar field is the large-scale or trend field. The second component in this analysis is represented by the synoptic time scale or the field variability on the oceanic mesoscale. The composite SSHA field from the 10 and 17 days of altimeter tracks from the various platforms is thus considered synoptic in time as each day this field is updated with the latest tracks of SSHA data. The third component in this objective mapping approach represents unresolved scales such as noise and mapping errors. Each day, final field estimates of the SSHA data are a sum of the trend field and the objectively mapped deviation field in space (Mainelli et al. 2001). In this procedure, the mapping noise is significantly reduced by including two or more sets of altimeter data. This analysis procedure allows the SSHA data to accurately depict mesoscale features as well as delineate areas of horizontal gradients each day when the latest data are ingested into the scheme.

c. EPIC data

To understand this EPAC upper-ocean variability, the EPIC field program, was conducted in and over the warm pool and along the 95°W transect to improve our understanding of these upper-ocean processes and determine their relationship to the atmospheric boundary layer (Weller et al. 1999; Raymond et al. 2004). During the field program, oceanic current, temperature, and salinity measurements from airborne expendable current profilers (AXCPs), airborne expendable conductivity-temperature and depth (AXCTD) profilers, and airborne expendable bathythermographs (AXBTs) were acquired from NOAA WP-3D and the National Center of Atmospheric Research WC-130 research aircraft (see Table 1). Profilers were deployed from 19 research flights encompassing the warm pool and the intertropical convergence zone (ICTZ) and along the 95°W equatorial transects in September and October 2001 (Raymond et al. 2004). Oceanic profilers were complemented with flight-level winds and atmospheric profiler data from GPS sondes. Flight tracks were located on adjacent sides of the R/V Ron Brown and R/V New Horizon centered on the 10°N TAO mooring in September 2001 (Wijesekera et al. 2005).

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Table 1. Summary of EPIC research flights from the NOAA WP-3D and NCAR WC-130 where Gr is grid flight, 95°W is transect flight, JI is a Juliette Invest flight, SST is a cold tongue flight, CP is a current profiler, CTD is a temperature and salinity profiler, and BT is the temperature profiler only. Failures are to the right of the slash under each category.

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Objective mapping the thermal structure from the grid measurements revealed the WCR structure (Fig. 4) is consistent with SSHA measurements shown in Fig. 1. The OHC values estimated from these snapshots range between 50 and 55 kJ cm\(^{-2}\) compared to more than 100 kJ cm\(^{-2}\) in the LC (Shay 2009). The 26°C isotherm depth exceeded 45 m in the WCR (consistent with Fig. 1 for the 20°C), and decreased to 20 m outside of it. As the WCR propagated southwestward in October, the isotherm depth decreased to about 15 m in the approximate region of the Costa Rica Dome along the eastern portion of the domain (Hofmann et al. 1981). The spatial variability of the WCR structure impacts the OHC distributions as observed in other regimes (Shay et al. 2000; Lin et al. 2005).

d. TAO mooring data

NOAA TAO moorings in the EPAC are part of the long-term monitoring efforts by the Pacific Marine Environmental Laboratory (PMEL). These moorings provide time series of thermal structure at various depths and during the EPIC program, these TAO moorings were enhanced to acquire temperature measurements at 1, 5, 10, 20, 40, 60, 80, 100, 120, 140, 180, and 500 m, and salinity at 1, 5, 10, 20, 40, 80, and 120 m at 8°, 10°, and 12°N, 95°W (Cronin et al. 2002). TAO temperature structure data from moorings at 95°, 110°, and 140°W will be used here to assess the OHC climatology based on satellite sensing.

As shown in Fig. 5, the thermocline is depressed (i.e., 20°C isotherm depth) during the passage of a WCR at the TAO mooring located at 10°N, 95°W as suggested by the higher SSHA values in Fig. 1a. That is, the ring has warmer water at depth, which causes a positive SSHA detected by radar altimeters. At this position, the OHC values exceeded 40 kJ cm\(^{-2}\) as the warm feature began to spin down and weaken, consistent with Figs. 1 and 4. Over several years of measurements, TAO time series from May to November are used to evaluate remotely sensed isotherm depths and OHC values.

e. VOS XBT transects

Ship-based expendable bathythermographs (XBTs) are acquired through the volunteer observing ship (VOS), Ship of Opportunity Program (SOOP), which includes NOAA, University-National Oceanographic Laboratory System (UNOLS), and Coast Guard vessels. These data are routinely provided to the Atlantic Oceanographic and Meteorological Laboratory (AOML) via a Windows based real-time ship and environmental data acquisition and transmission system. This software creates a series of reports, which describe ship movements that are transmitted in a real time via the Global Telecommunication System (GTS) to operational databases to be used by scientists. Specifically, XBT profiles from the ships involved in the SOOP are transmitted to receiving stations onshore via satellite. These data are placed in the GTS and are made publicly available. The global spatial data

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**Fig. 4.** Averaged OHC (kJ cm\(^{-2}\)) from in situ thermal profiles during EPIC in (left) September and (right) October based on AXCPs and AXCTDs temperature structure from the NOAA WP-3D research aircraft. Contour lines denote the 26°C isotherm depth.
are available since 1999. Here, data were accessed from the NOAA/AOML Web site and the data from the EPAC from 2000 to 2008 were reprocessed (see Fig. 6). Since minimal quality control is applied to these data, profiles were checked prior to making any calculations (more information on the method is available online at http://www.aoml.noaa.gov/phod/trinanes/SEAS/).

3. Approach

Based on recent analysis, the approach in the Atlantic Ocean basin by Shay et al. (2000) and Mainelli et al. (2008) has been revised for application in the EPAC as well as for the Atlantic Ocean basin. The rationale is that the stratification is much stronger in the EPAC and in particular the EPIC domain where the buoyancy frequency (i.e., the vertical derivative of the density structure) has a maximum ($N_{\text{max}}$) value of 24 cycles per hour (cph; Wijesekera et al. 2005) compared to values of 6–12 cph in the Atlantic Ocean basin (Fig. 7, right panel). Vertical salinity changes (not shown) at the base of the oceanic mixed layer ($h$) also contribute to these density changes. However, upper-ocean salinities in the EPAC tend to be less than in the GOM because of the ITCZ where excess rainfall reduces the mixed layer salinities compared to those in GOM (Gill 1982). The shoaling thermocline from west to east sets this large value of the buoyancy frequency, which is a close proxy to the 26°C isotherm depth needed for the OHC estimations (Palmen 1948).

a. Two-layer model

The two-layer model used here is based on reduced gravity where the upper and lower layers are separated by the depth of the 20°C isotherm:

$$g' = \frac{g (\rho_2 - \rho_1)}{\rho_2},$$

where $g$ is the acceleration of gravity (9.81 m s$^{-2}$), $\rho_2$ represents the density of the lower layer, and $\rho_1$ represents the density of the upper layer (O’Brien and Reid 1967; Kundu 1990; Goni et al. 1996). The total depth of the 20°C isotherm is given by

$$H_{20} = \frac{\bar{H}_{20}}{\bar{H}_{20}} \frac{h}{g'},$$

where $\bar{H}_{20}$ represents the average depth the 20°C isotherm from a hurricane-season climatology and $\eta'$ is the SSHA from the blended and objectively analyzed altimetry measurements from multiple platforms. Each day, the SSHA field is updated with new tracks of data in the domain from the various radar altimeters, which implies that the 20°C isotherm depth is also updated. The key issue is then the relationship of the 26°C isotherm depth (Palmen 1948). The updated total depth of the 26°C isotherm is determined from the relationship:

$$H_{26} = \frac{\bar{H}_{26}}{\bar{H}_{20}} H_{20},$$

where $\bar{H}_{26}$ is the average depth of the 26°C isotherm from hurricane-season climatology. Given this depth based on satellite altimetry and the SST from satellite measurements, the OHC relative to the 26°C isotherm is given by

$$Q = \rho_1 c_p \int_{H_{26}}^\eta [T(z) - 26°C] dz,$$

where $T(z)$ is the upper-ocean temperature structure that includes the SST; $c_p$ is the specific heat of seawater at

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**FIG. 5.** OHC (kJ cm$^{-2}$) and the 20°C isotherm depth (m) determined from the TAO mooring at 10°N, 95°W during September and October of 2001 during the EPIC field program.
constant pressure, which has a value of 1 cal gm\(^{-1}\) \(^{\circ}\)C\(^{-1}\); and \(p_1\) is 1.026 g cm\(^{-3}\) (Leipper and Volgenau 1972). Note that we use the 26\(^{\circ}\)C isotherm here since it is the temperature assumed for tropical cyclogenesis (Palmen 1948).

b. Empirical representation

To estimate the OHC from space-based measurements, we use the climatological OML \((h)\) and assume that the SST from Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI; Gentemann 2007) is a proxy for the OML temperature. From the surface to this depth (Fig. 7, gray shaded area), the OHC in the mixed layer is proportional to the product \([\text{SST} - 26^{\circ}\text{C}]/h\]. The second contribution to this estimate is underneath this layer from \(h\) to the depth of the 26\(^{\circ}\)C isotherm using the SST given by \(0.5(H_{26} - h)/(\text{SST} - 26^{\circ}\text{C})\). The total OHC is then the addition of the OML contribution plus the contribution from the layer depth to the 26\(^{\circ}\)C isotherm depth. There are few underlying assumptions involved with this approach:

1) Seasonal climatological OML has its own density that sits on top of a two-layer fluid; and,
2) The OML depth is time invariant since we do not have surface heat fluxes, wind stress, or current shear at the base of the OML to determine its evolution.

Variations in the SSHA field have maximum impact in the seasonal thermocline (i.e., 20\(^{\circ}\)C isotherm) depth and have a minimal impact on \(h\) similar to Cooper and Haines (1996). This approach has been tested in the Atlantic Ocean basin to assess whether there is an improvement in the OHC estimates. Comparisons have indicated good agreement between satellite-inferred and observed isotherm depths and OHC estimates in the GOM given its weaker thermal structure and vertical density changes compared to that observed in the EPAC and the Gulf Common Water (GCW) (Shay 2009).

c. U.S. Navy generalized digital environmental model

Central to this vertical structure issue is its representation in climatology and in situ data (Fig. 8). In comparing
the two climatologies (V2.1 and V3 GDEM) on monthly and seasonal time scales, the OML depth is included in the approach (Teague et al. 1990). For example, the ocean mixed layer \( (h) \) by definition is well mixed in properties such as temperature and salinity. Close inspection of the salinity profiles for September reveal differences between salinity in the upper 10–15 m of the water column (not shown). That is, the salinity structure in V2.1 reveals no constant salinity in this near-surface layer; by contrast, the salinity is relatively constant in the layer from 15 m to the surface. While these vertical salinity changes affect the density structure, the thermal structure dominates the density in the upper ocean, which seems to be more realistic in GDEM V2.1 than V3.0 (see Fig. 8). For example, the well-mixed OML (temperature only) depth decreases from a maximum in May of 35–40 m to a minimum of about 15 m in October. Such thermal structure behavior is more consistent with observed thermal structure variations during from the EPIC CTD profiles. That is, this layer shallowing from the EPIC data suggests that the OML is about 20–25 m deep in the warm pool or about 5–10 m deeper than the September climatology. Of more importance here is the vertical temperature structure beneath the OML in GDEM V2.1, which is more in line with the CTD profile from the R/V Ron Brown than GDEM V3.0. However, when averaged over the EPAC hurricane season (May–November) at 10°N, 95°W, climatologies suggest a general OML shallowing.

For these reasons, we chose to use the earlier version of GDEM climatology for the analysis contained herein. This is also the same version used for the Atlantic Ocean basin to ensure consistency between the two hurricane active basins. Climatologically, the SSTs exceed 28°C in the EPAC as shown in Fig. 9a. The spatial variations in the surface OML depths are shown in Fig. 9b based on an average from May through November. Notice how the OML shoals toward the east where mean values range between 15 and 20 m as compared to OML of more than 80 m west of 120°W. These spatial changes in the OML are now reflected in the approach. As shown in Figs. 9c,d, the seasonal mean depths of the 20°C and 26°C isotherms are based on an average over a hurricane season using GDEM profiles objectively analyzed to 0.25° resolution. Notice the general shoaling of the isotherm depths from west to east that forces tighter vertical
gradients in the warm pool’s upper-ocean thermal structure (e.g., shallower OML depth), which is markedly consistent with Fig. 8. Generally, the 20°C isotherm depths range from 30 to 50 m compared to more than 100 m west of 140°W. The corresponding 26°C mean isotherm depth ranges between 15 and 25 m in the warm pool (12°N, 95°W) and north of 20°N, the 26°C isotherm shoals to the surface. This surface shoaling, known as ventilating of the 26°C isotherm, implies that once a hurricane reaches that area, it will begin to lose their oceanic heat source and begin to weaken as the air–sea fluxes will diminish quickly (e.g., negative feedback).

The reduced gravity \( (g') \) distribution \([\text{from (1)}]\) and the ratio between the 26°C and 20°C isotherm depths are shown in Figs. 9e,f. East of 120°W, reduced gravities are about \( 5 \times 10^{-2} \text{ m s}^{-2} \), which is indicative of the strong stratification of the EPAC. West of this longitude, \( g' \) decreases to about \( 3.5 \times 10^{-2} \text{ m s}^{-2} \) whereas toward the northern part of the domain, reduced gravities decrease to about \( 2 \times 10^{-2} \text{ m s}^{-2} \). For example, in the area of the Hurricane Norbert experiment in 1984 (Sanford et al. 1987), the observed buoyancy frequency was \( \sim 12 \text{ cph} \) compared to more than 20 cph in the warm pool (Raymond et al. 2004). Such spatial variations in the stratification have a pronounced impact on OML cooling and the generation of a cold wake or trail left behind by a hurricane. In general, the strong stratification in the EPAC often precludes a cold SST wake at these low latitudes since wind-driven current shear instabilities are insufficient to lower the Richardson numbers to below criticality. An example of this was observed during Hurricane Juliette in 2001. The cold SST wake of more than 4°C only began toward the north and west of the warm pool (Shay and Jacob 2006).

d. Ocean heat content estimation

As shown in Fig. 10, the OHC is estimated using GDEM V2.1 climatology (Fig. 9) using the SSHA and TMI-derived SST fields for mid-September 2001 to coincide with the EPIC experiment (Raymond et al. 2004). The approach uses SSHA from TOPEX, GFO, and ERS-2 altimetry data (not the blended AVISO in Fig. 1) where repeat tracks are 9.9, 17, and 35 days, respectively. Jaimes and Shay (2009) showed virtually no difference between this product and the AVISO product during September 2005 in the GOM where correlation coefficients exceeded 0.92. These fields are blended and objectively analyzed to a 0.25° grid from the coast to 180° and from the equator to 30°N and are then combined to estimate isotherm depths and OHC. As shown in Fig. 10a, the warm SSTs exceeded 27.5°C north of the equatorial cold tongue and extended longitudinally from the coast to 180°. Cooler SSTs are observed north of 20°N, and decrease to below 26°C at about 24°N. The mean 20°C isotherm depths suggest a general shoaling of the thermocline from west to east to a relative minimum of about 40 m in the EPIC domain. Notice the general shape of this minimum that apparently was affected by a warmer feature between the two cold cells (Fig. 10b). This may be a manifestation of the Costa Rica Dome, which is a semipermanent feature of the EPAC forced by a cyclonic mean wind stress curl (Hofmann et al. 1981). The 26°C isotherm depths also show a similar
pattern except that the relative minimum is about 20–25 m. Finally, the resultant OHC distribution shows values of \(50 \text{ kJ cm}^{-2}\) at \(14.8^\circ N, 95^\circ W\).

\[e. \text{ Sea surface temperatures}\]

The surface boundary condition is central to these satellite retrievals as suggested in Fig. 7. As shown in Fig. 11, SSTs from Reynolds, TMI, and TAO mooring data are compared using regression techniques. For example, TAO mooring derived data suggest warmer SSTs than those derived from Reynolds analysis (slope of the least squares fit is 0.63) with RMS differences of about 0.6°C. The better comparison is between TAO and TMI data where the RMS difference is 0.51°C. Note that in this comparison, the slope of the regression curve is 0.93, which is indicative of a better fit with a bias of 1.67°C. Finally, the Reynolds analysis is compared to the TMI SSTs. RMS differences are 0.54°C and the slope of the regression curve is 0.72 with a bias of 8.19°C. In general, Reynolds-derived SSTs tend to be higher than the TMI SST since TMI data are corrected for diurnal cycling (Gentemann 2007). Thus, not all SST products are equal, and when it represents the surface boundary condition, careful analyses must be done to optimize the choice of the SST product in the data stream to estimate OHC for input into forecasting models (DeMaria et al. 2005).

\[4. \text{ Observational comparisons}\]

\[a. \text{ TAO moorings}\]

An important aspect of this approach is to understand differences between directly observed values and those inferred from satellite remote sensing techniques (Fig. 12). Observed SSTs, 26°C isotherm depth, and OHC values from the TAO mooring at 10°N, 95°W (as well as CTD profiles from R/V Ron Brown) are compared to satellite-derived fields that use the TMI and the GDEM V2.1-derived hurricane climatology. Isotherm depths were estimated by fitting the observed temperature profile to a cubic polynomial. Ocean mixed layer depths were then determined using a temperature threshold of 0.5°C from the surface, using the “well mixed” assumption to the cubic fit temperature profile.

SSTs from 1-m depth on the TAO mooring are in good agreement with those from the satellite-derived SSTs over the time series. Here, the approximate 0.5°C differences are attributed to the fact that the TMI instrument senses only skin temperature whereas the TAO data represent a bulk measurement in the upper part of the OML.

**Fig. 9.** TC seasonal climatology values for (a) SST (°C), (b) \(h\) (m), (c) \(H_{20}\) (m), (d) \(H_{26}\) (m), (e) \(g' \times 100\) (m s\(^{-2}\)), and (f) mean ratio of \(H_{26}/H_{20}\) from GDEM V2.1 for use with the empirical approach outlined in Fig. 7.
However, the depth of the 20°C isotherm suggests about an approximate 10-m difference between observed and satellite-inferred values. The TAO mooring data suggests a shallower 20°C isotherm depth compared to those from satellites. Since the TAO mooring data are acquired at discrete depths of 20, 40, and 60 m, the comparison to the CTD profiles from the R/V Ron Brown indicate that the derived depth from the TAO moorings is slightly deeper. As shown in Fig. 12b, however, the satellite-inferred 26°C isotherm depth shows a much higher correlation to this observed isotherm depth. Note that as the WCR passes the mooring in mid-September, both isotherm depths reflect a deeper, warmer layer. The corresponding OHC (Fig. 12c), derived from the satellite data, range in values of 20–30 kJ cm$^{-2}$ prior to ring passage. As the ring passes over the mooring, OHC increases to 38 kJ cm$^{-2}$ where the TAO mooring suggests a slightly larger value of 43 kJ cm$^{-2}$.

In addition to the 10°N, 95°W TAO moorings during EPIC, comparisons have been made at three other TAO sites 12°N, 95°W; 8°N, 110°W (within or nearby the EPIC domain); and 9°N, 140°W (northernmost mooring closest
FIG. 11. (left) Regression analysis and (right) histograms of differences in SST (°C) measurements between (top) Reynolds and the TAO 1-m mooring data, (middle) TMI and TAO, and (bottom) Reynolds and TMI, at 10°N, 95°W from May to October 2001. The regression line (including the bias and slope) and RMS differences are given for each analysis.
to the XBT line we studied) to assess spatial variations in the satellite algorithm to reflect the EPAC variability and not just the warm pool from 2000 to 2003 (Table 2, Fig. 13). However, it should be pointed out that the moorings are located along the equatorial waveguide where energetic Kelvin waves can distort the isothermal structure during El Niño years. Climatologies may contain this oceanographic phenomenon where temperatures may be smoothed out and caution needs to be applied to these comparisons along the periphery of the waveguide. In addition, the method to determine the isotherm depth is based on a linear interpolation in the vertical direction from $T(z)$ measurements which may induce small biases (e.g., 110°W). The slopes between TAO and satellite-inferred values lie between 0.4 and 0.6 for the 26°C isotherm depth compared to results at 10°N, 95°W. As shown in Fig. 13, the 4-yr comparisons for OHC are promising in that the RMS differences range from 8 to 10 kJ cm$^{-2}$ at all four locations. The dynamic range of the OHC ranges has maximum values as high as 80 kJ cm$^{-2}$. Thus, these values reflect about 10%–18% of the differences assuming no vertical displacements in the thermistor chain in the TAO moorings due to off-equatorial upwelling processes and equatorial Kelvin wave passage along the equator.

b. XBT transects

As shown in Fig. 6 (top panel), 6 yr of XBT data from this transect are used to estimate OHC and compare it to the 6-yr average from satellite-inferred values along the same transect as well as the closest moorings (Fig. 14). The dark transect line represents the depth of the 26°C isotherm in the bottom panel. Note the marked agreement between the XBT, mooring, and satellite-derived OHC values. In addition, there are no significant differences between the in situ and the remotely sensed values at 95% confidence along the repeat transect.

Over a broader spatial scale from 2000 to 2008 during the EPAC hurricane season, isotherm depth and OHC value statistics are listed in Table 3 from 6420 in situ data points. The depth biases range from 8 to 23 m for the 20°C isotherm depths and 8 to 18 m for the 26°C isotherm depths. The corresponding regression slopes range from 0.7 to 0.9 for these depths where RMS differences are typically 10%–15% of the total depth range depending on the year. Satellite-derived and in situ measurements suggest RMS OHC differences were 13–21 kJ cm$^{-2}$ or 12%–15% of the maximum values. Of particular relevance here is the fact that the regression slope for satellite-derived OHC values is 0.9 with RMS
difference of 17 kJ cm$^{-2}$ where the dynamic range lies between 113 and 190 kJ cm$^{-2}$. This latter value is a maximum rather than an average (see Fig. 15). Thus, RMS differences are within expected uncertainties and within 15% the maximum signals. The corresponding regression analysis and the histograms of the differences indicate consistency with these statistical results in Table 1. Of particular importance here is that the OHC regression line has a slope $\sim$0.9 and a bias of $-1$ kJ cm$^{-2}$. With 6420 data points from these transects, this value suggests quality satellite retrievals for OHC using this simple and crude modeling approach that uses the GDEM 2.1 climatology (Teague et al. 1990).

5. Basin-to-basin variability

An important aspect of this problem is the considerable variability in OHC estimates between basins due to the different temperature and salinity characteristics, and more importantly the strength of the thermocline and halocline. Temperatures and salinities vary in response to incoming radiation and precipitation (ITCZ) as well as the air–sea fluxes (Gill 1982). As suggested in Fig. 7, profiles from the LC subtropical water and the tropical EPAC illustrate marked differences in the temperatures and the resultant buoyancy frequency profile. In an OML, the buoyancy frequency ($N$) are nearly zero by virtue of the well-mixed assumption in temperature and salinity. By contrast, the maximum buoyancy frequency ($N_{\text{max}}$) of the LC subtropical water mass ranges from 4 to 6 cph and remains relatively constant. Below the 20°C isotherm depth ($\approx$250 m), $N$ decreases exponentially with depth. By contrast, the GCW buoyancy frequency is typically 12 cph across the OML base (Shay et al. 1992).

In the EPAC, however, $N_{\text{max}}$ is $\approx$20 cph due to the sharpness of the thermocline and halocline (pycnocline) located at the OML base (i.e., 25–30 m). Beneath this maximum, $N \geq 3$ cph are concentrated in the seasonal thermocline over an approximate thermocline scale (b) of 200 m and exponentially decay with depth approaching 0.1 cph. Such behavior has important implications for shear-instability and vertical mixing processes. In the EPAC, this implies that for large $N$, wind-forced shears have to be significantly larger for vertical mixing to occur compared to the LC or the GCW. Given a large $N_{\text{max}}$ and lower latitudes (12°N) where the inertial period is long in the EPAC warm pool, SST cooling and OML deepening will be much less than in the GOM as observed during Hurricane Juliette in September 2001 (Shay and Jacob 2006). This is precisely why few cold wakes (>2°C) are found in the EPAC warm pool regime. By contrast, significant SST cooling of more than 5°C occurred when Juliette moved northwest where $N_{\text{max}}$ decreased to about 14 cph at $\approx$20°N (e.g., shorter inertial period). While for the same hurricane in the GOM, similar levels of SST cooling would be observed in the GCW, but not in the LC water mass because the 26°C isotherm depth is 3–4 times deeper. These regional- to basin-scale variations in oceanic structure and the resultant stratification represent a paradox for hurricane forecasters, which is the rationale underlying the use of satellite radar altimetry in mapping isotherm depths and estimating OHC from SSHA fields and assimilating them into oceanic and coupled models.

a. Stratification parameter

To place these differing variations into context, a stratification parameter ($S$) is introduced that allows us to understand such differences. This parameter ($S$) is given by $\sqrt{N_{\text{max}}/N_o}$, where $N_{\text{max}}$ represents the maximum buoyancy frequency located across the OML base and $N_o$ is the reference buoyancy frequency for a given reference density (i.e., temperature and salinity). Here we use a reference buoyancy frequency of 3 cph, which is consistent with other basins, representing the value at the base of the seasonal thermocline. As shown in Fig. 16, $S$ has a maximum value in the EPAC warm pool where $N_{\text{max}}$ ranged between 20 and 24 cph observed during EPIC (Raymond et al. 2004). Thus, this ratio is approximately 2.8–3 in the warm pool compared to values of about 1.5–2 on the periphery of the warm pool and the northern tier of the hurricane-prone domain. Farther west, $S$ ranges from 2.6 to 2.8 between 8° and 12°N and between 130° and 145°W. However, the stratification parameter decreases to values less than 1.8 west of this patch of higher values. As shown in Fig. 7, $S$ is
FIG. 13. (left) Scatterplots and (right) histograms of differences between TAO (abscissa) and satellite-derived (ordinate) OHC (kJ cm$^{-2}$) for (a) 8°N, 140°W; (b) 8°N, 110°W; (c) 12°N, 95°W; and (d) 10°N, 95°W from 2000 to 2004. The black line in the scatterplots shows the perfect fit; the RMS differences in the top left of each panel are based 4 yr of data.
approximately unity in the LC and WCR regimes. In the western Pacific, the stratification parameter is 1.2–1.4.

b. Equivalent OHC

The stratification parameter is determined empirically from in situ measurements and climatology keeping in mind that the strength of the stratification at the OML base is an important parameter in vertical mixing processes through the Richardson number (Price 1981; Sanford et al. 1987; Shay et al. 1992). Here we then introduce equivalent OHC given by

\[
OHC_E = OHC \sqrt{N_{\text{max}}/N_0},
\]

where OHC is the vertically integrated thermal structure from the surface to the depth of the 26°C isotherm.

---

**Fig. 14.** Average OHC (kJ cm\(^{-2}\)) from repeat XBT transect (blue line), TAO moorings at 2°, 5°, and 8°N (140°W; colored boxes), and the corresponding satellite-derived values (black line) with \(\pm 2\sigma\) and the corresponding vertical temperature structure from the XBTs where the depth of the 20°C isotherm (white line) and the 26°C (black line). Data are averaged for the months of July from 2000 to 2005.

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**TABLE 3.** As in Table 2, but for the XBTs deployed from 2000 to 2008 where \(n\) = number of profiles compared to values derived from satellite altimetry. The depth and value represent an average over the profiles in a given year. The final column represents an average over all years. (The XBT profiles were from the web site: http://www.aoml.noaa.gov/phod/trinanes/SEAS/.)

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FIG. 15. (left) Scatterplots and (right) histograms of differences between XBT (abscissa) and satellite-derived (ordinate) for (a) $H_{20}$ (m), (b) $H_{26}$, and (c) OHC (kJ cm$^{-2}$) for over 6400 data points spanning the time scale from 2000 to 2008. The black line in the scatterplots shows the perfect fit.

bias = 12.2
slope = 0.8
RMS = 22.3

bias = 11.0
slope = 0.8
RMS = 17.8

bias = -1.0
slope = 0.9
RMS = 16.8
as above. This expression allows us to compare OHC values in differing basins or regions. For example, OHC in September 2001 in the warm pool ranged between 38 and 43 kJ cm$^{-2}$ as noted above. However given the strength of the stratification ($S \sim 3$), OHC$_E$ ranges from 114 to 129 kJ cm$^{-2}$, respectively (Fig. 17), which means the highly stratified water will act as a barrier to strong shear-induced mixing until such time that vertical shears develop to lower the Richardson number to below critical values. At these low latitudes from 10$^\circ$–14$^\circ$N, more OHC is available during hurricane passage through the air–sea fluxes as mixing will be suppressed for a long period of time. In general, entrainment mixing is what forces 65%–80% of the cooling and layer deepening during hurricane passage (Price 1981; Jacob et al. 2000).

During Hurricane Juliette in September 2001 (Shay and Jacob 2006), and the subsequent intensification to category 4 status, the SST cooling was less than 1$^\circ$C in the regime with strong vertical gradients (approximately 20–24 cph; Wijesekera et al. 2005). Wind-driven ocean current shear was insufficient to significantly cool the upper ocean through shear instability until Juliette moved into an area with much weaker stratification where SST cooling was 4$^\circ$–5$^\circ$C. Entrainment mixing across the OML base due to the ocean current shear did not lower the bulk Richardson number to below criticality. Hence, a larger fraction of OHC was available for Juliette through air–sea fluxes during the rapid intensification phase over less than 24 h (Raymond et al. 2004).

Second, these levels of OHC$_E$ are nearly equivalent to those observed in the western Atlantic basin. That is, OHC values in the northwest Caribbean Sea have values of 120–150 kJ cm$^{-2}$ with a corresponding equivalent OHC value because $S \sim 1$ in that regime. Similar values of OHC$_E$ are found in the subtropical water (e.g., LC). In the western Pacific, OHC values are about 100 kJ cm$^{-2}$, and when multiplied by 1.5, the corresponding OHC$_E$ are quite similar to those in the EPAC and LC–WCR complex. Thus, the relative import of this simple empirical relationship is that it allows forecasters to understand not only spatial variability in OHC levels, but provides a means of assessing these processes in differing basins based on stratification, which is in the numerator of the Richardson number (Price 1981; Shay et al. 1992).

6. Summary and conclusions

Sea surface height anomaly fields and an oceanic climatology were used to estimate isotherm depths and ocean heat content fields in the eastern Pacific Ocean.
basin. An objective analysis method was used to combine SSHA fields from several satellites. A realistic climatology for OHC relative to 26°C isotherm depth was developed for the eastern Pacific Ocean using GDEM V2.1 (Teague et al. 1990). This approach now includes a climatological mixed-layer depth coupled with the features developed by Mainelli (2000) for the Atlantic Ocean basin product. This approach differs from Goni and Trinanes (2003) in that they use an annual mean climatology to describe tropical cyclone heat potential, whereas we use a hurricane season only (May–November) for OHC relative to the 26°C isotherm depth. Including the entire year, the thermal structure climatology is oversmoothed and not necessarily representative of the hurricane season as OHC will be under- and overestimated in the northern and southern parts of the basins, respectively. Mainelli (2000) showed that the mean isotherm depths differed considerably between an annual and a hurricane-season average. This hurricane season approach was shown to be effective in SHIPS in the Atlantic Ocean basin for severe hurricanes (DeMaria et al. 2005; Mainelli et al. 2008) and the rapid intensification index (Kaplan et al. 2010). While Pun et al. (2007) used climatological values to evaluate an altimetry derived product in the western Pacific Ocean, efforts are under way to build and test an improved climatology that will provide daily climatological values for use with altimeter-derived fields.

We have shown detailed evaluations of satellite-inferred values to observed profiles including estimates of the isotherm depths and OHC fields from the EPIC experiment, time series from the TAO moorings, and several years of XBT transects from volunteer observing platforms. The comparisons show very high correlations with mooring and XBT data as well as shipboard measurements during the EPIC field experiment (Raymond et al. 2004). The statistics from 8 yr of these in situ measurements suggest that RMS differences are about 10–15 kJ cm\(^{-2}\) with regression slopes of \(O(0.8–1.0)\). As in the Atlantic Ocean basin, the values of OHC typically are 0.9–1.1 kJ cm\(^{-2}\) m\(^{-1}\) relative to the depth of the 26°C isotherm (Leipper and Volgenau 1972). Following the approach above, we will continue the detailed comparisons.

![Figure 17](image_url)

**Fig. 17.** Averaged (a) OHC and (b) OHC\(E\) (kJ cm\(^{-2}\)) during September 2001 during the EPIC field program [black box in (a)] and the track and intensity of Juliette (23 September–3 October). (c) The buoyancy frequency (cph) from 19 days of CTD measurements from the NOAA R/V Ron Brown at 95°N, 10°W where the maximum buoyancy frequency was about 22 cph consistent with AXCTDs deployed from the NOAA aircraft. Notice the OHC maps clearly delineate the Costa Rican Dome just east of the EPIC domain and the warm pool.
to ensure that the satellite-derived values are grounded in actual oceanic measurements from TAO mooring data, XBT transects, and ARGO profiling floats.

To place the OHC into context with other basins, we have developed an empirical parameter based on the maximum buoyancy frequency and a reference buoyancy frequency. This stratification parameter \( \sqrt{N_{\text{max}}/N_0} \) provides a normalization based on the strength of the stratification observed at the OML base where shear-induced mixing occurs. These mixing events are often driven by vigorous near-inertial motions forced by the wind stress and its curl (Shay et al. 1989). Thus, this empirical approach then allows us to compare values and assess the threshold values currently used in SHIPS of 60 kJ cm\(^{-2}\) in the Atlantic Ocean basin (DeMaria et al. 2005; Mainelli et al. 2008). There is observational evidence that this threshold is fairly large compared to the coupled measurements acquired during Isidore and Lili (Shay and Uhlhorn 2008) where surface heat losses were more on the order of 10 kJ cm\(^{-2}\) for surface air–sea fluxes of 1.4–1.7 kW m\(^{-2}\). This is an important issue that needs to be resolved in coupled models aimed at forecasting intensity change.

Finally, the approach is based on a reduced gravity modeling approach (O’Brien and Reid 1967; Goni et al. 1996) with broad assumptions. Notwithstanding, the comparisons to in situ data suggest that this relatively crude approach works reasonably well in the absence of global ocean models that assimilate data from various sources such as floats, moorings, SOOP XBTs, and satellite altimetry (Marks et al. 1998). In the Atlantic Ocean basin, we have found marked consistency with the Hybrid Coordinate Ocean Model (HYCOM) that assimilated data from Global Ocean Assimilation Experiment product (Halliwell et al. 2008, 2009, manuscript submitted to Mon. Wea. Rev.) and the Navy Coupled Ocean Data Assimilation (Cummings 2005) products for pre-Isidore and Lili in 2002 and pre-Ivan in 2004. The added benefit of the primitive equation model is that after the hurricanes pass by, one can carefully assess the relative importance of advection and mixing on the three-dimensional response including the heat losses. This would provide considerable insights for ocean models that will be coupled to atmospheric models aimed at forecasting hurricane intensity in research and operational models.

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