Limitation of One-Dimensional Ocean Models for Coupled Hurricane–Ocean Model Forecasts

RICHARD M. YABLONSKY AND ISAAC GINIS
Graduate School of Oceanography, University of Rhode Island, Narragansett, Rhode Island

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ABSTRACT

Wind stress imposed on the upper ocean by a hurricane can limit the hurricane’s intensity primarily through shear-induced mixing of the upper ocean and subsequent cooling of the sea surface. Since shear-induced mixing is a one-dimensional process, some recent studies suggest that coupling a one-dimensional ocean model to a hurricane model may be sufficient for capturing the storm-induced sea surface temperature cooling in the region providing heat energy to the hurricane. Using both a one-dimensional and a three-dimensional version of the same ocean model, it is shown here that the neglect of upwelling, which can only be captured by a three-dimensional ocean model, underestimates the storm-core sea surface cooling for hurricanes translating at \(<5\) m s\(^{-1}\). For hurricanes translating at \(<2\) m s\(^{-1}\), more than half of the storm-core sea surface cooling is neglected by the one-dimensional ocean model. Since the majority of hurricanes in the western tropical North Atlantic Ocean translate at \(<5\) m s\(^{-1}\), the idealized experiments presented here suggest that one-dimensional ocean models may be inadequate for coupled hurricane–ocean model forecasting.

1. Introduction

It has become well established that hurricanes require heat energy from the ocean surface to develop and intensify, but the wind stress imposed on the upper ocean by the storm can limit the intensification process primarily via shear-induced mixing of the upper ocean and entrainment of cooler water into the upper oceanic mixed layer (OML) from below (e.g., Ginis 1995, 2002, and references therein). Since this shear-induced mixing is a one-dimensional (1D) process, some recent studies suggest that coupling a 1D ocean model to a hurricane model may be sufficient for capturing the storm-induced sea surface temperature (SST) cooling in the region providing heat energy to the hurricane (Emanuel et al. 2004; Lin et al. 2005, 2008; Bender et al. 2007; Davis et al. 2008). If in fact a 1D model is sufficient, valuable computational resources can be saved as compared to coupled models that employ a fully three-dimensional (3D) ocean component.

The purpose of this study is to evaluate the difference in the SST response to hurricane wind forcing underneath the storm core using both a 1D and a 3D version of the same ocean model. While the ocean’s 3D baroclinic response to a moving hurricane has been extensively studied, with summaries provided by Anthes (1982), Gill (1982), and Ginis (1995, 2002), the focus here is on the potential impact of the first upwelled region behind the storm center. If the storm is translating slowly, this upwelled region, which can only be captured with a 3D ocean model, may occur under the storm core, thereby allowing for enhanced shear-induced mixing and subsequent SST cooling (Price 1981). Advection may also contribute to the ocean response in the 3D model, especially when background currents associated with ocean fronts or eddies are present (e.g., Oey et al. 2007), but here, background currents are avoided by prescribing the initial density field to be horizontally homogeneous (see section 2). Finally, evaporative heat flux to the atmosphere, while vital to the hurricane, contributes far less than mixing/entrainment to storm-core SST cooling in the deep ocean (Price 1981; Shen and Ginis 2003; D’Asaro et al. 2007), so for this study, heat flux between the ocean and atmosphere is neglected.

2. Experimental design

The 3D experiments in this study are performed using a version of the Princeton Ocean Model (POM; Blumberg...
and Mellor 1987; Mellor 2004) that is similar to the version used in the operational Geophysical Fluid Dynamics Laboratory–University of Rhode Island (GFDL–URI) coupled hurricane–ocean prediction system (hereinafter the GFDL model) for the Atlantic Ocean basin (Bender et al. 2007). The 1D experiments use this same version of POM, except the advection and pressure gradient terms are removed so that at each grid point, there is no interaction among surrounding grid points in the horizontal.1 This 1D simplification is fully consistent with the 1D version currently used in the GFDL model for the eastern Pacific Ocean basin (Bender et al. 2007); it is also generally consistent with the 1D layer models used in Emanuel et al. (2004), Lin et al. (2005, 2008), and Davis et al. (2008), except for differences in the way vertical mixing is parameterized and the fact that POM is a level model instead of a layer model.2 While these differences can impact the mixing magnitude and hence the magnitude of the storm-core sea surface temperature cooling, all of these 1D models (including the one used herein) are limited by their inability to account for up-welling.

For both the 1D and 3D experiments, the ocean grid spans from 108.5° to 60°W longitude and from 10° to 47.5°N latitude. Unlike the operational GFDL model, the ocean grid is set on an f plane, where the earth’s rotation rate and the longitudinal grid spacing assume constant latitude of 22.4°.3 There are 508 (449) grid points in the x (y) direction, yielding a horizontal grid spacing of 9.8 (9.3) km in the x (y) direction. The entire domain is assumed to be a 2500-m-deep ocean (no land or bathymetry), and the 23 half-sigma levels are placed at the following depths: 2.5, 7.5, 12.5, 17.5, 22.5, 27.5, 32.5, 40, 47.5, 55, 67.5, 87.5, 125, 187.5, 275, 387.5, 550, 775, 1050, 1400, 1800, 2250, and 2500 m.

In all experiments, the ocean is initialized with a horizontally homogeneous temperature (T) and salinity (S) profile and no background currents. Two T profiles are tested: in the “GCW” experiments, the T profile is based on the 0–2500-m portion of the Generalized Digital Environmental Model (GDEM) climatological profile in the Gulf of Mexico Common Water during the month of September (Teague et al. 1990), while in the “CRB” experiments, the T profile is based on the 0–2500-m portion of the September GDEM climatological profile in the Caribbean Sea, except for the OML temperature, and hence sea surface temperature, is adjusted slightly to match the OML temperature in the GCW experiments (Fig. 1). The main difference between the two temperature profiles in the upper ocean is the increased OML depth in the CRB profile relative to the GCW profile.

Once the T and S are defined everywhere on the POM grid, wind stress is applied (Fig. 2a), with the storm center initially located at (22.4°N, 71.7°W). The wind stress distribution is based on the wind field derived from an analytic model of the wind and pressure profiles in hurricanes (Holland 1980), where in this case central pressure (p_c) = 950 hPa, environmental pressure (p_e) = 1013 hPa, maximum wind speed (V_m) = 50 m s⁻¹, radius of maximum winds (RMW) = 55 km, air density (ρ_a) = 1.28 kg m⁻³, and exponential decay parameter (B) = V_m^2 p_a/(p_e – p_a), where e = 2.718 28. Asymmetry is included in this otherwise axisymmetric wind field by adding half of the storm translation speed, as in Price (1981). Once the wind field is calculated, the wind stress magnitude is calculated using the bulk formula, in which the drag coefficient (C_D) is calculated as an empirical function of the 10-m wind speed, similar to Moon et al. (2007) but modified to decrease C_D at high wind speeds to be more consistent with observations, as suggested by Tung (2008). In all experiments, this wind stress field translates from right to left (i.e., “westward”) with a prescribed speed, and model integration continues until the average SST cooling under the storm core reaches the quasi–steady state (≈5 days). Translation speeds from 0.2° to 2° longitude (6 h⁻¹) (i.e., ~1–10 m s⁻¹) [with an increment of 0.1° longitude (6 h⁻¹) (i.e., ~0.5 m s⁻¹)]
are tested so that the impact of translation speed on the difference in storm-induced SST cooling between the 1D and 3D versions of the model can be determined.

Since the goal of this study is to quantify the magnitude of SST cooling only within the region providing most of the heat energy to the storm, the average SST cooling is calculated within a 60- and a 200-km radius around the storm center for each experiment once the quasi–steady state is reached. The exact radius over which ocean heat flux significantly impacts storm intensity is not well known and likely varies depending on storm size, but 60 km includes what may be considered the storm “inner core” because it is 5 km larger than the radius of maximum winds and is consistent with the inner-core definition of Cione and Uhlhorn (2003). Note that the axisymmetric component of the wind speed at a 60-km radius is \( \approx 50 \, \text{m s}^{-1} \) in the experiments presented here (Fig. 2b). Using a 200-km radius captures what Cione and Uhlhorn (2003) define as the “inner-core wake,” and idealized model sensitivity experiments performed by Shen et al. (2002) suggest that ocean heat flux may be important for storm intensity even at this large radius from the storm center. Note that the axisymmetric component of the wind speed at a 200-km radius is \( \approx 25 \, \text{m s}^{-1} \) in the experiments presented here (Fig. 2b).

3. Results and discussion

The average SST cooling within a 60- and 200-km radius of the storm center for all experiments after a quasi–steady state is reached is shown in Fig. 3. The most striking aspect of this result is the large difference in SST cooling between the 3D and 1D experiments with translation speeds \( \approx \, 5 \, \text{m s}^{-1} \) (\( \approx \, 3.5 \, \text{m s}^{-1} \)) in the GCW (CRB) experiments. For the slowest translation speed tested, \( \approx 1 \, \text{m s}^{-1} \), the 3D simulation for the GCW (CRB) experiment yields an average SST cooling of \( \approx 10 \, ^\circ\text{C} \) \( \approx 4.5 \, ^\circ\text{C} \) within the 60-km radius, while the 1D simulation for the GCW (CRB) experiment yields an SST cooling of only \( \approx 2.5 \, ^\circ\text{C} \) \( \approx 0.5 \, ^\circ\text{C} \) within the 60-km radius (Fig. 3a). Even for storms translating at \( \approx 2 \, \text{m s}^{-1} \), the
3D simulation produces double the cooling of the 1D simulation. Similar results are found for the average cooling within the 200-km radius (Fig. 3b), although the magnitude of cooling is \( \sim \frac{1}{3} \) less than the 60-km radius cooling for the corresponding experiment.

Unfortunately, to the authors’ knowledge, there have been few if any oceanic measurements made within the core of a nearly stationary major hurricane, so there is currently insufficient observational evidence to support or refute SST cooling of \( \sim 5^\circ-10^\circ+8^\circ \)C within the core of such a storm. However, this magnitude of cooling is not beyond the realm of possibility because observations within the cold wakes of slow-moving Hurricane Hilda (1964) and Typhoon Virginia (1979) indicated cooling of \( \sim 6^\circ \)C (Leipper 1967; Pudov 1980), and observations within the cold wakes of Typhoons T8914, T8915, and Kai-Tak (2000) indicated cooling of up to \( 9^\circ \)C (Sakaïda et al. 1998; Lin et al. 2003). Also, note that it may be exceedingly difficult for a major hurricane to maintain its strength for enough time for such observations to be made in the face of such significant cooling, which of course is precisely why it is necessary for a forecast model to capture this cooling if and when it exists. Regardless, it appears likely that the \( \sim 5^\circ-10^\circ+8^\circ \)C of inner-core SST cooling indicated by the 1–2 m s\(^{-1}\) GCW-3D experiments is more plausible than the \( \sim 2^\circ-2.5^\circ \)C of inner-core SST cooling indicated by the equivalent GCW-1D experiments.

While significant differences in average SST cooling within the storm core between the 1D and 3D simulations for slower-moving storms imply the increased importance of upwelling relative to vertical mixing alone, it is helpful to examine temperature cross sections of the upper ocean along the storm track in these simulations to provide further evidence for this claim. First, to illustrate the concept, a time series is presented for stationary 1D and 3D experiments with the GCW initial profile (Fig. 4). Contours of turbulent kinetic energy (TKE) are overlaid on the temperature cross sections to indicate the regions of strongest mixing.\(^4\) In the 1D experiment (Figs. 4a,c,e,g), the OML deepens and cools over time in and around the RMW where strong mixing is occurring, but otherwise, the temperature field remains unchanged. In the 3D experiment (Figs. 4b,d,f,h), however, upwelling in and around the storm center dramatically changes the temperature field and contributes to continuous, enhanced SST cooling within the storm core.

Next, we focus on the 1.0, 2.4, 4.8, and 7.1 m s\(^{-1}\) translation speed experiments with the GCW initial profile once a quasi–steady state is reached (Fig. 5), as in Fig. 3. For the 1.0 m s\(^{-1}\) experiments (Figs. 5a,b), the difference between 1D and 3D is striking. In the 1D experiment (Fig. 5a), the OML depth steadily increases from an initial depth of \( \sim 35 \) m at \( >200 \) km ahead of the storm center to \( \sim 70 \) m at \( >80 \) km behind the storm center. As the OML depth increases, the OML (and hence sea surface) temperature steadily decreases by \( \sim 3^\circ \)C. In the 3D experiment (Fig. 5b), it is apparent that upwelling begins to impact the OML temperature even \( >200 \) km ahead of the storm center. However, the greatest impact begins between 110 and 40 km ahead of the storm, where the OML depth increases from \( \sim 60 \) to \( \sim 120 \) m and the OML temperature cools by \( \sim 6^\circ \)C. Significant cooling continues from front to back through the storm core, and the first upwelling maximum occurs \( \sim 40–110 \) km behind the storm, which when combined with continued mixing causes the OML temperature to be \( 12^\circ–15^\circ \)C cooler than the prestorm value in this region. Further behind the storm, the alternating pattern of upwelling and downwelling associated with internal inertia–gravity waves is evident, but this pattern occurs below and behind the region of strong vertical mixing, yielding little additional cooling in the OML.

For the 2.4 m s\(^{-1}\) experiments (Figs. 5c,d), the difference between 1D and 3D is still significant, but it is less dramatic than the difference for the 1.0 m s\(^{-1}\) experiments. This result can be attributed to the fact that for the 3D 2.4 m s\(^{-1}\) experiment, the first upwelling maximum is of lesser magnitude than in the 3D 1.0 m s\(^{-1}\) experiment, and this maximum is elongated toward the rear of the storm (i.e., farther from the region of maximum wind forcing), causing the OML cooling to be of lesser magnitude and to be less concentrated within the storm core. This trend continues for the 4.8 and 7.1 m s\(^{-1}\) experiments (Figs. 5e–h), such that for the 7.1 m s\(^{-1}\) translation speed (Figs. 5g,h), the difference between 1D and 3D is \( <1^\circ \)C until \( >200 \) km behind the storm center. Although not surprising, it is also interesting to note that the presence of upwelling behind the storm center in the 2.4 and 4.8 m s\(^{-1}\) experiments causes the mixing maximum to be closer to the surface in the 3D experiments relative to the 1D experiments; this upward shift may allow the cooling to be realized at the surface faster in the 3D experiments than in the 1D experiments.

Since all temperature cross sections presented thus far have been for the GCW experiments (Figs. 4 and 5), a brief analysis of the temperature cross sections for the CRB experiments may prove useful for understanding the impact of the initial temperature profile on the upper-ocean response near the storm core (Fig. 6). For all four translation speeds, the pattern of mixed layer deepening in the 1D CRB experiments (Figs. 6a,c,e,g) and upwelling

\(^4\) TKE in POM is calculated using the Mellor–Yamada level 2.5 turbulence closure scheme (Mellor and Yamada 1982; Mellor 2004).
in the 3D CRB experiments (Figs. 6b,d,f,h) is quite similar to the corresponding GCW experiments (Fig. 5). Even the distance from the storm center to the first upwelling maximum behind the storm is similar in the 3D CRB and 3D GCW experiments, but this distance is slightly less in the former than in the latter, most likely because the fastest baroclinic mode wave speed ($c$) is faster in the 3D CRB experiments (2.87 m s$^{-1}$) than in the 3D GCW experiments (2.01 m s$^{-1}$). According to linear wave theory, the longest waves that are steady relative to a storm

5 MATLAB programs sw_bfrq.m and dynmodes.m are used to calculate $c$ (Morgan and Pender 2009; Klinck 2009).
translating at speed $U$ are excited in the along-track direction with wavelength $L = (U^2 - c^2)^{0.52} \pi f$, assuming $U > c$ (Geisler 1970; Ginis 2002). Since the distance to the first upwelling maximum behind the storm center is proportional to $L$, then for a given $U$, $L$ should be smaller in the 3D CRB experiment than in the 3D GCW experiment. Regardless, the main reason storm-core SST cooling appears to be restricted in the CRB experiments (Fig. 6) relative to the GCW experiments (Fig. 5) is the deeper initial OML depth in the former relative to the latter.

Finally, it is instructive to revisit the 1.0, 2.4, 4.8, and 7.1 m s$^{-1}$ translation speed experiments with the GCW initial profile once a quasi–steady state is reached, but this time, to examine spatial plots of the sea surface temperature and current velocity vectors (Fig. 7) to supplement the aforementioned temperature cross sections.

**FIG. 5.** Along-track, quasi–steady state ocean temperature cross sections for the (left) GCW-1D and (right) GCW-3D experiments with translation speeds of (a),(b) 1.0; (c),(d) 2.4; (e),(f) 4.8; and (g),(h) 7.1 m s$^{-1}$. Contours and symbols are the same as in Fig. 4.
Generally, the SST cooling pattern immediately left and right of the storm track is similar to the SST cooling pattern along the storm track, thereby justifying the use of the along-track temperature cross sections to infer the dominant mechanisms of upper-ocean cooling in the storm core. However, it is important to note that in both the 1D and 3D experiments, the maximum surface current velocity, and hence the region of maximum SST cooling, are shifted to the right of the storm track. The rightward bias in the surface current field is a characteristic feature of the ocean response to a moving storm and is well known from previous observational and numerical studies (e.g., Price 1981); it can be explained by the superposition of the wind stress vector’s rotation due to the storm’s forward motion, inertial rotation due to the
Coriolis force, and to a lesser extent, the asymmetry in the wind stress magnitude. The rightward bias in the SST cooling is a result of the rightward bias in the shear-induced turbulent mixing in the water column. Note that the magnitude of this bias does not appear to be affected dramatically by the inclusion (or deletion) of upwelling.

4. Summary and conclusions

By neglecting upwelling, 1D ocean models miss an important mechanism for hurricane-induced SST cooling, and for storms translating at $\sim 3.5$–$5$ m s$^{-1}$ or less (depending primarily on the OML depth in the initial upper-ocean temperature profile), this upwelling-enhanced SST

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**Fig. 7.** Quasi–steady state SST and current vectors for the (left) GCW-1D and (right) GCW-3D experiments with translation speeds of (a),(b) 1.0; (c),(d) 2.4; (e),(f) 4.8; and (g),(h) 7.1 m s$^{-1}$. Temperature contours and symbols are the same as in Fig. 4; dashed rings indicate 60- and 200-km radii; the 1 m s$^{-1}$ vector scale is shown in the lower left.
cooling can occur within the storm core where heat flux from the ocean to the atmosphere can impact hurricane intensity. For storms translating at <2 m s\(^{-1}\), the idealized experiments presented here suggest that upwelling more than doubles the magnitude of storm-core SST cooling relative to vertical mixing alone. According to the idealized experiments presented here, upwelling more than doubles the magnitude of storm-core SST cooling relative to vertical mixing alone. Therefore, 1D ocean models may be inadequate for coupled hurricane–ocean model forecasting.

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