Impacts of Ocean–Atmosphere Coupling on Tropical Cyclone Intensity Change and Ocean Prediction in the Australian Region

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ABSTRACT
This study investigates the impact of atmosphere–ocean coupling on predicted tropical cyclone (TC) intensity change and the ocean response in the Australian region. The coupled model comprises the Australian Bureau of Meteorology’s Tropical Cyclone Limited-Area Prediction System (TC-LAPS) and a regional version of the BLUElink ocean forecasting system. A series of case study forecasts are presented and the differences between coupled and uncoupled forecasts, operational forecasts, and posterior objective analyses are compared. A coupled model ensemble is also developed that uses different first-order approximations of the effects of surface waves on surface stress in an inertial coupling method. In each of the cases, the use of reanalyzed sea surface temperatures significantly improves the prediction of TC intensity change in the intensification phase. The results show that dynamic air–sea coupling has a modest impact on intensity in cases where SST cooling is significant and is likely to be important for predicting the rate of TC intensification, peak intensity, and deintensification. Results also show that there is a definite coupled signal and suggest inherent biases in the atmospheric model that could potentially be removed. With different parameterizations of surface wave effects, results show modest sensitivity in TC intensity of up to 10 hPa in minimum surface pressure; however, in some cases there was significant sensitivity in the predicted ocean response. The results also highlight the relative increased complexity of tropical cyclone prediction in the Australian region compared to other regions. In cases where the forecast TC track was reasonably skillful, there were improvements in the predicted ocean response with respect to observations compared to an ocean reanalysis.

1. Introduction
An important source of energy for tropical cyclones (TCs) is heat and moisture from the upper ocean. Large-scale environmental and climatological factors that contribute to the formation and support of tropical cyclones include warm sea surface temperatures (SSTs), usually above 26°C, a relatively deep oceanic mixed layer, large low-level cyclonic absolute vorticity, low vertical wind shear over the disturbance, mean upward motion, and high midlevel humidities (Gray 1968; DeMaria and Kaplan 1994; Dare and Davidson 2004). The TC speed and intensity are important factors governing air–sea interaction. In particular, an intense slow moving tropical cyclone is likely to be strongly coupled to upper-ocean heat content (Leipper and Volgenau 1972). The passing of the cyclone induces ocean currents close to the inertial frequency and strong vertical upward mixing that entrains cooler water from the base of the mixed layer (Chang 1985). With increasing winds, oceanic vertical mixing reduces SST causing a reduction of surface heat fluxes and a negative feedback on TC intensity (Schade and Emanuel 1999).

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The ocean response to tropical cyclones is characterized by the formation of persistent cool wakes that are associated with sea level anomaly (SLA) depressions of around 20–30 cm (Shay et al. 1990). The wakes are mainly caused by upward vertical mixing of cooler subsurface water and upwelling from surface mass divergence induced by the wind field. The presence of the cooler water lowers the upper-ocean heat content and gives rise to steric effects and the resulting baroclinic pressure gradients lead to geostrophic currents (Ginis and Sutyrin 1995; Shay and Chang 1997). The role of the ocean in TC intensification has been investigated in various case studies in the Gulf of Mexico (Hong et al. 2000; Jacob and Shay 2003) and theoretical studies such as Ginis et al. (2004). These studies suggest warm core mesoscale eddies may act as heat reservoirs for cyclones and be related to intensification in some cases. In situ observations, from several ship-based studies, of surface cooling induced by TCs, once they have passed, suggest a response in the range 1°–6°C (Price 1981). Surface cooling may also be important for TC intensity change in slow-moving TCs. Emanuel (1999) suggests that TC intensity change depends mainly on initial intensity, the atmospheric thermodynamic environment, and heat exchange under the core with the upper ocean. In terms of the ocean’s role in TC intensity change, SST cooling under the TC core is likely to be more important than the cooling in the wake. The cooling may not only be induced by surface heat fluxes and vertical mixing but also by natural transition of the TC to cooler oceanic environments such as that which may occur when moving to higher latitudes. Regardless, an SST cooling under the core of 2.5°C may be sufficient to completely shut down energy production (Emanuel 1999).

In this paper, the impact of ocean–atmosphere coupling on TC and ocean prediction in the Australian region is investigated with a series of case studies. Tropical cyclones impact across a broad region of northern Australia where large areas of both shelf sea and deep ocean exist. In the shallow regions there may not be a source of deep cooler water to entrain to the surface. Other processes such as advection, internal waves, and the presence of density fronts may influence the ocean response at the shelf break. Furthermore, climatological sea surface temperatures are approximately 2°C warmer and salinities are several psu lower in the equatorial western Pacific Ocean than in the equatorial Atlantic Ocean. This suggests that the work required to vertically mix cooler waters to the surface, which depends on the potential energy of the stratification and the density structure, may be greater in the Australian region. These factors may contribute to important regional differences in TC and ocean interaction.

The next section presents details of the coupled model, a brief outline of the methods, and a description of the case studies. Following this, the results that illustrate the role of SSTs and air–sea coupling on TC intensity prediction are presented. Following this, an assessment of the ocean response for each of the case studies in terms of both modeling and observations is examined. Finally, the conclusions are given.

2. Model description

The model used is the coupled limited-area model (CLAM), which consists of a regional version of the global BLUElink Ocean Forecasting Australia Model (OFAM; Brassington et al. 2007; Oke et al. 2008) coupled to the Bureau of Meteorology’s Tropical Cyclone Limited-Area Prediction System (TC-LAPSv4; Davidson and Weber 2000) using the Ocean Atmosphere Sea Ice Soil (OASIS) coupler version 3.1 (Valcke et al. 2003). The ocean model is based on version p1 of the Modular Ocean Model (MOM4p1; Griffies 2009) with 1/6° horizontal resolution and 47 levels, the upper 20 levels having 10-m resolution. For the cases studied here, version 2.1 of the BLUElink Reanalysis (BRAN2.1; Oke et al. 2005) is used for initial and boundary conditions. The atmospheric model is configured at 0.15° horizontal resolution and has 29 vertical σ levels. The domain of both models is initialized with the vortex in the center of a 30° × 30° region. The coupling frequency adopted is 2 h and time-averaged fields are passed between the models. TC-LAPS surface heat fluxes drive ocean model heating–cooling. Ocean model SST is a surface boundary condition in TC-LAPS influencing heat fluxes, atmospheric pressure gradients, and winds. An inertial coupling method for surface stress (Bry and Wolff 1999) is used to determine a shared surface stress in both models. This algorithm incorporates relative differences in winds, currents, and fluid surface densities and accounts for surface wave field Stokes drift velocity. The surface wave field Stokes drift velocity is estimated from the 10-m winds. Surface stress drives currents and ocean vertical mixing and influences marine atmospheric boundary layer shear. Vertical mixing induces entrainment of cooler subsurface water, influencing upper-ocean heat content (OHC) and SST. OHC influences the expansion or contraction of the water column producing the SLA and currents.

3. Initialization

The procedure for a coupled hindcast is shown in Fig. 1. Ocean model initial conditions are acquired from 48 h behind the forecast base time. The first 24 h involves a dynamical spinup of the ocean model. The ocean model is initialized with BRAN2.1 temperature and salinity, but
with zero sea level and currents. Incremental analysis updating to BRAN2.1 analyzed sea level and currents is carried out to recover the analyzed dynamical ocean state. The 24-h period is longer than the inertial period and sufficient for geostrophic adjustment to take place. Temperature and salinity are not constrained in order to allow initial adjustment of SST and heat content to tropical cyclone conditions. Also in the initialization phase, ocean model fluxes are linearly adjusted from the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis 6-hourly surface fluxes to TC-LAPS surface fluxes at the analysis base time. The spinup of the surface fluxes and ocean state is an attempt to dynamically adjust the ocean model to the observed TC fluxes while also producing an initial ocean response to the TC. After the ocean spinup, a coupled run is initialized. This is referred to as the coupled analysis period in which the atmospheric model is nudged to a synthetic vortex through air temperature, surface pressure, mixing ratio, and wind vorticity. A technique for nudging upper-atmospheric thermal properties based on GMS cloud imagery is also used (Davidson and Weber 2000). In the atmospheric model nudging period, the system is coupled in order to allow some adjustment of the ocean to the vortex. From the forecast base time the coupled model is integrated forward for 72 h without constraints.

4. Coupled surface fluxes

The atmospheric and ocean models are coupled through surface momentum, heat, and freshwater fluxes. The surface heat fluxes calculated in the atmospheric model induce modest changes in SST; however, surface stress imparts momentum flux into the ocean and drives horizontal and vertical mixing, accounting for most of the induced surface cooling. The semi-implicit quadratic law for wind stress \( \tau \) at the ocean surface typically used to force ocean models is

\[
\tau = C_D \rho_A \mathbf{u}_A \sqrt{u_A^2 + v_A^2},
\]

where \( \rho_A \) is air density, \( C_D \) is the 10-m drag coefficient, and \( \mathbf{u}_A \) represents either \( u_A \) or \( v_A \), that is, the components of the 10-m wind velocity vector. There are various empirical formulations for \( C_D \) based on observations. It can be derived by measuring boundary layer wind or current profiles in different sea states. A correct choice of \( C_D \) could account for all the physical processes governing surface stress. This, however, becomes an area of uncertainty in high wind (\( >40 \text{ m s}^{-1} \)) conditions where few observations exist to determine the relationship between wind speed, sea state, and \( C_D \). The Powell et al. (2003) dataset suggests a reduction of \( C_D \) at wind speeds above \( \sim 40 \text{ m s}^{-1} \). Bye and Jenkins (2006) developed a unified boundary layer model, based on an inertially coupled wind–wave–current system that also predicted reduction of \( C_D \) at high wind speeds. Moon et al. (2004) compare various parameterizations of \( C_D \) at high wind speeds, including those from Powell et al. (2003), and suggest that \( C_D \) at extreme wind speeds under a TC has strong spatial dependence on translation speed and the wave and wind fields.

An alternative way to estimate surface wind stress is to account for more of the known physical processes. The inertial coupling method of Bye and Wolff (1999) can be used to determine an inertially coupled surface stress \( \tau_S \) that is dependent on the wave field and the relative free-stream velocities in the respective fluids:

\[
\tau_S = \frac{\rho_A K}{(1 + \epsilon)^2} \left[ \mathbf{u}_A - (1 - \epsilon) \mathbf{u}_L \right] \left[ \mathbf{u}_A - (1 - \epsilon) \mathbf{u}_L \right],
\]

where

\[
\epsilon = \sqrt{\frac{\rho_A}{\rho_S}},
\]

and \( \rho_A \) is air density, \( \rho_s \) is seawater density, \( u_A \) is wind speed, \( u_s \) is current speed, and \( u_L \) is the spectrally integrated surface wave phase speed. Here \( \epsilon u_L \) is the spectrally integrated Stokes velocity and \( K \) is a constant drag coefficient evaluated at the edge of the wave boundary layer independent of wind speed. The inertial coupling method gives lower stress than the quadratic method for winds in the same direction as currents. Later in this article we present results from the coupled model regarding the relationship between wind speed and surface stress.

FIG. 1. Illustration of the initialization and coupled forecast procedure.
that infer an implicit drag coefficient. This relationship is compared to drag coefficients obtained from Bender et al. (2007) and Bye and Jenkins (2006).

In the coupled model, $\tau_s$ is used as the surface stress that governs the shear in the marine atmospheric boundary layer and drives ocean currents and mixing. $\tau_s$ is evaluated using 10-m winds and $\rho_A$ from TC-LAPS and surface currents and $\rho_s$ from the ocean model. Based on the recommendations of Bye and Wolff (1999), $K$ is chosen to be 0.0012 and the assumption that $u_L$ is proportional to the 10-m winds is also made.

5. Case studies and sensitivity experiments

The experiments involve 5 tropical cyclone case studies that intensified to category 4 or 5 on the Saffir–Simpson scale (Table 1). The tracks for these cases are shown in Fig. 2 and are based on best-track data from the Joint Typhoon Warning Center. Tracks are colored in terms of normalized intensity and a circle and date, corresponding to 0000 UTC, is used to signify the initial position. A cross is marked on each track every 24 h. Experiments Zoe and Inigo use a 19-$\sigma$-level tropical cyclone model configuration based on an earlier version of the operational system. The remainder of cases use a 29-$\sigma$-level configuration, which is the current operational version. The cases occurred within the BRAN2.1 reanalysis period (1992–2006) where sufficient information is available for estimating the prestorm ocean state. In each case, a forecast is run from a base time that is approximately prior to intensification.

<table>
<thead>
<tr>
<th>Name</th>
<th>Period</th>
<th>96-h forecast database</th>
<th>Region</th>
<th>Estimated min surface pressure (hPa)</th>
<th>Highest category</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inigo</td>
<td>1–8 Apr 2003</td>
<td>1 Apr 2003</td>
<td>Southeast Indian Ocean</td>
<td>900</td>
<td>5</td>
<td>Most intense TC in Australian region</td>
</tr>
<tr>
<td>Fay</td>
<td>16–28 Mar 2004</td>
<td>16 Mar 2004</td>
<td>Southeast Indian Ocean</td>
<td>910</td>
<td>5</td>
<td>Strong SST cooling and SLA depression</td>
</tr>
<tr>
<td>Ingrid</td>
<td>4–17 Mar 2005</td>
<td>7 Mar 2005</td>
<td>Coral Sea</td>
<td>930</td>
<td>5</td>
<td>Made landfall 3 times with long track and lifetime</td>
</tr>
<tr>
<td>Floyd</td>
<td>19–27 Mar 2006</td>
<td>22 Mar 2006</td>
<td>Southeast Indian Ocean</td>
<td>927</td>
<td>4</td>
<td>Intensified and decayed over ocean</td>
</tr>
</tbody>
</table>

Fig. 2. Tracks for the TC case studies based on best-track data from the Joint Typhoon Warning Center. Tracks are colored in terms of normalized intensity for each TC. Circles and dates, corresponding to 0000 UTC, denote initial positions. Crosses marked on each track are 24 h apart.
Coupled model sensitivity experiments were carried out according to Table 2 using representations of the spectrally integrated peak phase speed at different percentages of wind speed. Also shown are the corresponding spectrally integrated Stokes velocities. Modifying the Stokes velocity term in the inertial coupling equation is akin to altering the drag coefficient. Experiments without inertial coupling and with zero wave effects in the inertial coupling method are also carried out.

6. Results

CLAM forecasts are first compared to operational TC-LAPS forecasts and analyses. The CLAM forecasts compose the ensemble listed in Table 2 and an uncoupled simulation using fixed BRAN2.1 reanalysis SSTs. Both the operational and CLAM forecasts use the same atmospheric model physical parameters, initial, and boundary conditions. It should be noted that the operational TC-LAPS analysis and forecast system uses spatially varying SSTs from a global 1° × 1° horizontal resolution optimal interpolation analysis (Smith 1995). The analysis uses in situ data from ships, buoys, and expendable bathythermographs (XBTs) and measurements from the Advanced Very High Resolution Radiometer (AVHRR) carried out on the National Oceanographic and Atmospheric Administration (NOAA) polar-orbiting satellites. The analysis–forecast cycle is 12-hourly in which each new analysis is updated with new SST data. In this system, SST is fixed in time for the period of the forecast. The modeled TCs in each analysis–forecast cycle experience different SSTs. Figures 3–7 show, for each case, minimum surface pressure, maximum wind speed, and mean SST following the storm for a 72-h period from the analysis base time. Data for mean SST following the storm frame of reference is acquired from within a circular area where the perimeter is taken to be twice the radius of maximum winds from the center of the storm. Showing mean SST underneath the storm helps characterize the SSTs that the storm is in contact with, assuming that SSTs over the larger-scale environment have less significant impacts. This also helps identify the cases that produce an overestimate of ocean surface cooling.

The patterns of changing minimum surface pressure, maximum wind speeds, and corresponding mean SST under the storm for the TC Inigo case are shown in Fig. 3. The ensemble of coupled model runs is denoted by the gray area and shows the spread due to the different parameterizations given in Table 2. The solid line shows the operational forecast and the heavy solid line is the operational analysis. There are significant improvements in the CLAM forecasts of intensity change compared to the operational model. The overall difference between the CLAM simulations and the operational model prediction can be attributed to warmer mean SSTs from the BRAN2.1 reanalysis. The ensemble spread in CLAM is relatively low and the uncoupled simulation with fixed BRAN2.1 reanalysis SST,
which remains a fraction warmer, produces a modestly more intense storm and is in slightly better agreement to observations. This appears to be related to the weak ocean surface cooling response in this case. The differences between coupled and uncoupled runs suggest that air–sea coupling creates important differences even with small changes in SST. The minimum threshold SST required for intensification in the model in this case appears to be \( \sim 29.5^\circ C \), since intensification did not develop in the operational forecast TC at SSTs <29.5°C. Throughout the intensification phase of the storm, fixed SSTs used in the operational forecast were biased cooler than TLCAPS analysis SSTs and the CLAM model. The simulations in CLAM produce TCs that intensify at a rate resembling the observed TC until a departure occurs at \( \sim 24 \) h. Since there was no significant SST change in the CLAM model, or a significant increase in forecast track error (not shown), the departure is likely to have resulted from atmospheric factors in the TC model.

The CLAM model simulations produce a significantly improved prediction of intensity change compared to the operational model in the TC Floyd case (Fig. 4). This again appears to be mainly due to the use of BRAN2.1 reanalysis SST, but is also related to a reliable track forecast. The operational model TC was exposed to SSTs that were biased cool compared to observations. The uncoupled CLAM simulation with fixed BRAN2.1 reanalysis SST intensifies earlier than suggested by observations, yet achieves the best forecast estimate of minimum surface pressure. The differences between coupled and uncoupled simulations in minimum surface pressure show that air–sea coupling has an important effect on the modeled TC. All simulations do not achieve the maximum winds suggested by observations and may indicate suboptimal physical parameterizations in the atmospheric model. Inertial coupling provides a slightly higher possible maximum wind speed for an equivalent intensity in terms of lowest central pressure compared to simulations that are either uncoupled or not inertially coupled. It appears that the SST cooling due to the ocean response is minimal and most of the cooling is due to the transition of the storm to higher latitudes. However, the spread in the coupled ensemble in both mean SST and minimum surface pressure indicates that air–sea coupling has an influence on both intensity and surface cooling. The case with no inertial coupling produces the most SST cooling in the ensemble (lowest values in the gray area represent this case). This case is the ensemble member that imparts the largest amount of wind-driven momentum flux into the ocean. The ensemble member with inertial coupling without surface wave effects produces the least intense storm compared to the other ensemble members. This suggests that first-order parameterized wave effects may be important for determining surface stress and TC intensity.

Figure 5 presents the results for TC Ingrid from the analysis base date of 7 March 2005. This time corresponds to 2 days after genesis when the storm was a category 3. In this particular case, at the time of initialization, the initial vortex structure and large-scale environmental conditions were not adequately described by the observations that were assimilated into the operational analysis. This resulted in the storm being of incorrect intensity at the forecast base time. During the 24-h period prior to forecast base time, nudging ensures that little difference is seen in the results between the operational and CLAM simulations. Once this constraint is removed, however, the representation of the tropical cyclone in the CLAM simulations intensifies to values closer to observations in the first 24 h of the forecast. The uncoupled CLAM simulation shows a marginal increase in intensity (decrease in lowest central pressure) compared to the coupled
runs where evolution of SSTs are permitted. The rapid rise in minimum surface pressure after 48 h is associated with Ingrid’s landfall over the Cape York Peninsula. The decay in intensity is more rapid and significant in the CLAM simulations compared to observations. The simulation of the uncoupled TC, however, deintensifies at a later time compared to the ensemble of coupled simulations. This is most likely because the storm developed a slower translation speed. The coupled model ensemble is spread by about 10 hPa at the time of maximum intensity. The ensemble member that produced the least intense storm is the case with no inertial coupling. This member created an unrealistic ocean response with too much SST cooling, at forecast hour 24, compared to observations (the lowest of the mean SSTs in the spread are from this case). The other cases produce a reasonable ocean response in terms of mean SST cooling following the storm compared to observations (shown later). This suggests that in the case with no inertial coupling, either the representation of surface stress was too high and/or the underlying initial ocean heat content was inaccurate. The operational model experienced warmer mean SSTs than the CLAM model from 18 h after forecast base time, yet did not continue to intensify. Since SSTs in CLAM were significantly warmer during the initialization and the first part of the forecast, this suggests that the energy supplied from the ocean during this relatively short period lead to intensification of the storm. During intensification, maximum winds in the coupled model were stronger in all ensemble members compared to the operational model. The ensemble member with the largest degree of wave effects predicted the strongest winds due to the fact that this parameterization produced the lowest surface stresses in the atmospheric model, compared to the other parameterizations. The low spread in the ensemble for maximum wind, however, suggests that, in this case, the atmospheric model prediction of maximum wind is relatively insensitive to coupling.

The TC Fay case is shown in Fig. 6. Here, coupled model minimum surface pressure is close to observations

![Fig. 5](image1.png)  ![Fig. 6](image2.png)

**Fig. 5.** As in Fig. 3, but for TC Ingrid. Analysis base time is 7 Mar 2005.

**Fig. 6.** As in Fig. 3, but for TC Fay. Analysis base time is 20 Mar 2004.
for the first 24 h of the forecast; however, after this the TC does not decay in intensity. In all Fay experiments, the CLAM simulations failed to capture the deintensification because the simulated TC moved southwest rather than west to an area of deeper ocean away from the shelf (shown later in Fig. 10). Large errors began to appear in the forecast track around 24 h from the forecast base date. The rapid growth in track error from this time shows that none of the models were capable of resolving the environmental control of the storm. The meander of the real track of TC Fay at the time of recurvature suggests significant environmental control was taking place. Most cases where TC tracks are not well forecasted result from uncertainty and model inadequacies in resolving the large-scale environmental atmospheric flow (N. Davidson 2009, personal communication). Minimum surface pressure and maximum wind speed do not show a large ensemble spread in this case. This may be due to either the small storm-induced SST cooling in the coupled model or that SST was above the threshold for sustaining the required surface fluxes for intensification and support of this storm. The results show that the TC in the uncoupled model experiences warmer SSTs, during the forecast, creating a slightly more intense storm than any of the coupled model simulations. Despite the fact that the track error lead to a significant divergence between the predicted and analyzed storms, the results show that air–sea coupling has an important effect on the modeled TC and the ocean response and is likely to be able to remove some of the inherent biases in the atmospheric model.

The results for TC Zoe (Fig. 7) are, in some ways, similar to TC Ingrid because the initial representation of the storm prior to the forecast was inadequately described because of a lack of observations. Because of this, neither the uncoupled, coupled or operational simulations were able to reproduce the correct intensity of the TC in the first 24 h prior to forecast base time. However, as with the Ingrid case, the TC in the CLAM simulations underwent rapid intensification from the forecast base time, whereas the operational model TC did not intensify to the same degree. Figure 7c shows that the most significant difference was the mean SST initial conditions under the storm that were on average $1.8^\circ C$ warmer in the BRAN2.1 reanalysis SST. As expected, the uncoupled simulation produced a slightly more intense TC because no feedback of cooler SSTs was permitted. The bias between initial TC-LAPS analysis SST and BRAN2.1 reanalysis SST highlights the difference between these products.

### 7. Assessment of the ocean response

It is typically difficult to obtain observations of the ocean state in tropical cyclone conditions for many reasons. These include the fact that TCs are localized and transient and have erratic tracks that rarely coincide with instruments such as moorings and Argo floats. Also, the present altimetry system does not resolve the spatial and temporal scales required and SST infrared sensors and microwave sensors cannot penetrate clouds or areas of precipitation, respectively. Most of the available ocean observations in the Australian region capture either prestorm conditions or cool wakes once the TC has passed. The correct combination of SLA and SST may indicate that the initial ocean heat content conditions, surface forcing, and vertical mixing are reasonable. This does not guarantee, however, that one or more of these are not incorrect and other terms are compensating. If the model reproduces a consistently reasonable ocean response for all cases and is in good agreement with the observations then an adequate balance is likely to be being achieved.

The BLUElink Ocean Data Assimilation System (BODAS; Oke et al. 2008) and BRAN2.1 are used
together with satellite superobservations (upscaled observations) to examine the ocean response and the sensitivity of inertial surface stress parameterizations in the coupled model. BRAN2.1 uses the BODAS ensemble optimal interpolation method to ingest the satellite altimetric sea level anomaly from a combination of altimeters such as the Environmental Satellite (Envisat), Jason-1, and the Ocean Topography Experiment [(TOPEX)/Poseidon]. It also assimilates microwave SST from the Advanced Microwave Scanning Radiometer for Earth Observing System (AMRS-E) and in situ temperature and salinity observations from Argo and ship-based XBTs. The coupled model ocean response to each of the cases is assessed by examining the degree of SST cooling in the TC wakes and the associated sea level height anomalies. It should be noted that the results shown for BRAN2.1 and BODAS are 24-h means whereas the results from the coupled model are 2-h means. BRAN2.1 is also forced with ECMWF reanalysis fluxes that typically underestimate TC intensity; however, they have relatively low error in track.

The ocean response for the Ingrid case is pronounced in both SLA and SST and was captured in the superobservations (Fig. 8). SLA in the BODAS analysis shows a deeper and differently positioned depression than in BRAN2.1. There are only several superobservations of SLA showing the depression to be < 25 cm. The coupled model experiments appear to reach this level. However, the track error caused the shape of the wake to be different. Currents and sea level patterns in Fig. 8 indicate there is approximate geostrophic balance around the wake. SST superobservations suggest the cool wake encompasses a larger area than suggested by SLA superobservations. The coupled model experiment 2 (see Table 2), with inertial coupling and wave effects creates the most realistic surface temperatures in the wake, whereas experiment 5 (no waves) and experiment 6 (no inertial coupling) appear to overestimate the cooling.

Tropical Cyclone Fay left a pronounced ocean response, which is not surprising as it was slow moving at maximum intensity for around 24 h near 14°S, 122°E. At this location it created a large area of anomalously cool SSTs and low sea level. This is captured in the superobservations in both SST and SLA (Fig. 9). After this, Fay underwent recurvature, deintensified, tracked south, and made landfall at around 20°S, 120°E. BRAN2.1 and BODAS are in reasonable agreement with the superobservations. The CLAM experiments, however, highlight a deficiency in the coupled model with its reliability on a well-forecasted track. Figure 10 illustrates the difference between the analyzed track from the Joint Typhoon Warning Center and the average forecast track from all the model runs. Track error in CLAM and the TC-LAPS operational model increased during the forecast period due to errors in prediction of the larger-scale atmospheric flow. The path of the TC in CLAM was over the shelf break prior to maximum intensity, meaning a source of cooler deeper water was not available to entrain to the surface as the storm approached maximum intensity. The intensity of the forecasted TC persisted longer as it tracked west, where it then created a cool SST wake and associated low SLA to the left of the track.

In the ocean response for TC Inigo (Fig. 11), BRAN2.1 underestimates the SLA depression compared to superobservations. CLAM places the depression in a different location because of error in the forecast track; however, it makes an improved estimate of the magnitude of the depression. A cool SST patch is observed around 15°S, 113.5°E. Both CLAM and BRAN2.1 seem to underestimate the degree of cooling, however, CLAM has a small area of SSTs close to observed values. CLAM experiment 5 has approximately 1°C warmer SSTs outside the storm environment due to weakened outgoing heat fluxes.

In the TC Floyd case (Fig. 12), the ocean response in the coupled model shows a SLA depression and region of maximum cooling centered near 17°S, 106°E. This is the location and time of maximum intensity and where, soon after, Floyd underwent recurvature. Prior to this, Floyd moved in from the northeast without leaving a cool wake in SST. Currents and SLA are mainly in geostrophic balance indicating a strong baroclinic response. The CLAM results in Fig. 12 show a nongeostrophic wavelike structure in currents and sea level to the southwest of the position of TC Floyd (19°S, 106°E) in the simulation. Tropical Cyclone Floyd then continued to track in a south–southeast direction and weakened as it passed into areas of significantly lower SSTs. The superobservations hint that there is a cool wake, which is more pronounced in SLA than SST. BRAN2.1 and BODAS seem to underestimate the size of the depression compared to the observations, whereas CLAM potentially overestimates this.

Figure 13 illustrates the ocean response in the Zoe case. There is a large sea level depression associated with a low heat content anomaly in the area where Zoe reached maximum intensity at 12°S, 169°E. The sea level depression of < 25 cm is in good agreement with SLA superobservations. BRAN2.1 seems to underestimate the depression and the depth-averaged currents in the upper 200 m. The three CLAM experiments show that, as surface stress is lowered through inertial coupling, the extent and magnitude of the depression decreases. SST cooling is not quite as large as that given by the superobservations indicating that prestorm ocean conditions from BRAN2.1 may not have had the correct upper-ocean fluid temperature.
Fig. 8. Ocean response to TC Ingrid on 9 Mar 2005. (a) BODAS-analyzed SST. (b) SST super-observations from the AMSR-E microwave sensor. (c) BLUElink ReANalysis (BRAN2.1) SST. Resulting SST from coupled limited-area model experiments (d) 2, (e) 5, and (f) 6. (h)–(m) As in (a)–(f), but illustrating corresponding sea level anomalies and depth-averaged currents up to maximum of 200-m depth.
FIG. 9. As in Fig. 8, but for TC Fay on 29 Mar 2004.
thermal structure. The true intensity of TC Zoe was far greater than was predicted by the model, so it is likely that the ocean response in the model is underestimated because the surface fluxes are underestimated. This probably explains why in this case, the CLAM experiment with no inertial coupling appears to be closest to observed cooling. SST in CLAM experiment 2 is remarkably close to BRAN2.1 without assimilation of observations. It should also be noted that any barotropic ocean response to balance TC winds propagates rapidly once the winds have passed. This means that the SLA superobservations may show areas of disagreement to the model due to phase errors in the barotropic waves.

The relationship between wind speed and wind stress obtained using the inertial coupling method and how this compares when using parameterizations of $C_D$ as constant $= 0.0013$ and as a function of wind speed from Bender et al. (2007) and Bye and Jenkins (2006) is shown in Fig. 14. The quadratic relationship given in Eq. (1) is used for the solid curves. The spread obtained in the ensemble experiments is derived from all of the case studies. It is clear that inertial coupling dynamically modifies surface stress through wave effects. The first-order effect is that through increasing wave effects, surface stress is reduced. It also appears that experiment 1 has unrealistically large wave effects that result in an untenable relationship between wind speed and wind stress. Inertial coupling in experiments 3, 4, and 5, however, place the relationship in between the bounds of the relationship obtained using a constant $C_D = 0.0013$ and those obtained using the Bender et al. (2007) and Bye and Jenkins (2006) formulations. Further justification for the use of inertial coupling can be found by assessing the affects of the ensemble members on ocean surface currents. We found unrealistically large surface ocean currents under high wind conditions when we did not use inertial coupling. An example is provided in Fig. 15 that shows the wind field for TC Floyd at a forecast time of 48 h and surface currents in experiment 3 (inertial coupling) and experiment 6 (no inertial coupling). Here, experiment 6 has unrealistically large surface currents, the maximum being $\sim 5.3 \, \text{m s}^{-1}$, whereas in experiment 3 the maximum current is significantly lower at $\sim 3.1 \, \text{m s}^{-1}$. Also, the area where relatively strong surface currents exist in experiment 3 is less than half that in experiment 6. This is a general issue for all the case studies.

8. Conclusions

The impact of coupling on tropical cyclone intensity forecasts and ocean prediction has been investigated in the Australian region using a coupled limited-area model (CLAM) that is based on a regional version of the BLUElink ocean forecasting system and the Bureau of Meteorology’s Tropical Cyclone Limited-Area Prediction System (TC-LAPS). In each of the case studies, the CLAM model improves the prediction of TC intensity change compared to the operational model. In all cases, the improvement in resolving intensification is attributable to the use of warmer BRAN2.1 reanalysis SST rather than due to the evolution of SST in the coupled model. The effect of air–sea coupling in the CLAM model consistently and modestly decreased the intensity of the predicted TCs compared to the uncoupled runs as has been found elsewhere (Bender and Ginis 2000). Cases where the rate of change of SST is enough to influence TC intensity show that air–sea coupling has an important effect on the rate of intensification, peak intensity, and deintensification of the modeled TC. As the coupling was found to have no negative impacts in cases where the rate of change of SST was insufficient to change the TC intensity, we recommend coupling be used in general. The coupling is also likely to be able to remove some of the inherent biases that appear to be in the atmospheric model. The results also show that inertial coupling, in general, provides a higher possible maximum wind speed for an equivalent intensity in terms of minimum surface pressure in the model TC. Although TC intensity appears to be relatively insensitive to inertial coupling, including this can provide a more physically realistic representation of the ocean response. The introduction of shock in the atmospheric model through the use of different SST initial conditions is unlikely to have had a significant effect on the forecasts since the 24-h initial nudging to the analyzed vortex will have assisted in minimizing this.
FIG. 11. As in Fig. 8, but for TC Inigo on 9 Apr 2003.
FIG. 12. As in Fig. 8, but for TC Floyd on 29 Mar 2006.
FIG. 13. As in Fig. 8, but for TC Zoe on 1 Jan 2003.
The coupling did not show significant impacts on the forecast TC track compared to the operational model in the cases studied here. Cases where initial atmospheric storm environmental conditions prior to the forecast were inadequately described because of a lack of observations showed that the CLAM simulations were able to rapidly intensify to the analyzed storm in the first 24 h because of the use of more accurate SSTs from BRAN2.1.

FIG. 14. Maximum wind stress as function of maximum wind speed obtained from the ensemble experiments of the case studies compared to what would be obtained using the parameterizations of the drag coefficient $C_D$ as constant $= 0.0013$ and as function of wind speed from Bender et al. (2007) and Bye and Jenkins (2006).

FIG. 15. (a) Wind and surface pressure field shown overlying surface skin temperature for TC Floyd at forecast hour 48 from base date at 0000 UTC 24 Mar 2006. Corresponding 2-h time-averaged surface currents in (b) experiment 3 (inertial coupling) and (c) experiment 6 (no inertial coupling).
Tropical Cyclone Fay is an important case that highlights the complexity of TC and ocean prediction in the Australian region. The significance of accurate track prediction for the ocean response is paramount when the TC is likely to pass over significantly different oceanic areas. In this case, the predicted track was oriented with the shelf break and the degree of cooling on the left side of the track was not realized in the simulation because there was not an available source of cooler water on the shelf. The actual track was farther offshore in the deeper ocean, where at maximum intensity, the real TC induced cooling was highly significant, as seen in the observations, and caused deintensification and recurvature.

In the cases where TC track error was not a significant factor, the simulated ocean response appeared to improve compared to the ocean reanalysis product. Remotely sensed observations of the surface cooling response in these cases were within the expected range of 1°C–6°C found by Price (1981). An ensemble of coupled model runs using the inertial coupling method, with different parameterizations of the effect of surface waves, showed TC intensity sensitivity of up to 10 hPa in minimum surface pressure. In general, waves modestly altered the representation of TC intensity in the atmospheric model, but improved the ocean model response. Uncertainty in the parameters used in the inertial coupling method could be resolved by maintaining an ensemble as performed here.

This study illustrates that simulated TC intensity in the present system possesses a large degree of sensitivity to SSTs. Therefore, forecasts using a coupled model will be sensitive to prestorm upper-ocean heat content provided by the operational ocean analysis and forecasting system. There are many interactive factors that determine the ocean surface cooling such as potential energy of the stratification, mixed-layer depths, levels of induced vertical mixing, water depth, and TC translation speed. The potential advantage of a coupled model for TC prediction is that if initial conditions are properly constrained to observations and there is skill in the forecast component models, then the coupled model has the potential to account for the many factors that govern the ocean-related aspects of TC intensity change. Air–sea coupling is likely to improve future prediction efforts in terms of the ocean response, since a more accurate representation of the TC, compared to representations from coarser-resolution global models, can be used to drive the ocean response. Furthermore, a coupled model offers closure of the surface flux budget and a better representation of the true physical system that in nature is coupled.

Overall, the relative improvements show that the CLAM system is likely to add value to tropical cyclone and ocean prediction in Australia. Further work could be carried out investigating the impacts of different data assimilation and initialization strategies, parameterizations of ocean vertical mixing, and coupling methodologies. The recommendation from these findings is that the operational system is likely to benefit in terms of TC intensity prediction, from the use of real-time and forecast SSTs derived from the present BLUElink ocean forecasting system. The results suggest that coupling may play an important role in predicting TC deintensification over the ocean in the Australian region. This would possibly provide improved guidance for issuing severe weather warnings and be of significant benefit to forecasters. Another aspect that this study highlights is the notion that different threshold SSTs probably exist for every storm. Storm threshold SSTs may also be time dependent during the life of the storm. This is an area open to further research.

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