Atmosphere–Ocean Coupled Dynamics of Cyclones in the Midlatitudes

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(Manuscript received 14 July 2003, in final form 8 April 2004)

ABSTRACT

It is well known that hurricane intensity is influenced by factors such as the storm’s initial intensity, the spatial extent of the storm, the thermodynamic state of the atmosphere through which it moves, the storm propagation speed, and sea surface fluxes along the storm track. Although several of these factors are also known to modulate the strength of midlatitude cyclone systems, little is known about the impact of air–sea interactions on storms outside the Tropics. To investigate the atmosphere–ocean dynamics of midlatitude North Atlantic storms, the Canadian Mesoscale Compressible Community (MC2) atmospheric model is coupled to the Princeton Ocean Model. Case studies include midlatitude extratropical storm Earl (1998) and an intense winter storm from January 2000, hereafter denoted Superbomb. On one hand, late-summer storms such as Earl encounter a thin mixed layer and produce a cold wake by inducing strong currents. Sea surface temperatures (SSTs) can be depressed as much as 5°C or more. On the other hand, winter storms such as Superbomb occur when the mixed layer is quite deep. Although impacts on SSTs and the upper-ocean temperature profile tend to be weak, about 1°C or so, storm-induced ocean currents can be large. In the specific cases of Earl and Superbomb, little impact on cyclone strength was detected.

1. Introduction

Emanuel (1999) illustrated that tropical cyclone intensification tends to be controlled by internal variability and dynamics, in relation to interactions with the ambient environment through which they move. Using a simple coupled ocean–atmosphere model, he demonstrated that hurricane intensity depends mainly on the storm’s initial intensity, the atmospheric thermodynamic state through which it moves, and the upper-ocean properties along the storm track. Related studies by Schade and Emanuel (1999), Bender and Ginis (2000), and Bao et al. (2000) have also shown that ocean interactions with tropical systems can strongly influence storm intensity.

When a cyclone passes over the ocean, the surface roughness extracts momentum from the cyclone, and in turn, the cyclone obtains heat and moisture through sensible and latent heat fluxes from the ocean surface. Thus, two interaction processes are present. Positive feedback tends to dominate the early development stages, when enhanced fluxes from the ocean coincide with cyclone intensification. Negative feedback occurs with the generation of storm-induced surface currents, which pass momentum to the upper ocean, enhance vertical exchanges within the upper ocean, and result in SST cooling. Thus, sea surface fluxes are reduced and cyclone intensity is diminished (Chang and Anthes 1979; Sutyrin and Khain 1984; Bender et al. 1993). SST cooling, also denoted the “cold wake effect,” can be as much as −6°C and depends on the storm’s intensity, spatial extent, and propagation speed, and the oceanic mixed layer thickness (Leipper 1967; Wright 1969; Price 1981; Stramma et al. 1986; Nelson 1998; Wentz et al. 2000; Lin et al. 2003). Bender et al. (1993) found an average cooling, for slow-, medium-, and fast-moving storms of 5.3°C, 3.5°C, and 1.8°C, respectively, in an analysis of 16 tropical cyclones.

Sutyrin and Khain (1984) used an axisymmetric hur-
hurricane model coupled with a 3D ocean model to numerically simulate the effect of storm movement on storm intensity. They showed that slowly moving storms and shallow oceanic mixed layer depths lead to a strong negative feedback of the ocean on the storm (see also Schade and Emanuel 1999; Mao et al. 2000; Chan et al. 2001). Simulation of real storms, using composite atmosphere–ocean model systems, confirmed these idealized studies; the cold wake induced by tropical cyclones can significantly influence storm intensity, particularly for slow-moving storms (Bender and Ginis 2000; Bao et al. 2000). Conversely, the SST response is negligible for quickly propagating tropical cyclones over deep oceanic mixed layers, and the resulting oceanic impact on storm intensity is small (Schade and Emanuel 1999). Cyclonically rotating storm winds cause the oceanic mixed layer currents to diverge from the storm center. The high velocity current shear produces mixing of the cooler thermoline water into the mixed layer, increasing the mixed layer depth and cooling the sea surface. Thus, the storm-related physical processes that control SST are radiation and air–sea heat exchange, as well as the dynamic processes of the upper ocean; the heat and mass budgets in the mixed layer as determined by entrainment mixing (Price 1981; Jacob et al. 2000; D’Asaro 2003).

Although tropical cyclones tend to weaken because of increased wind shear and lower SST as they move to higher latitudes, some are able to reintensify, becoming very deep midlatitude storms. These are extratropical transitions (ET) of tropical cyclones, and they can be stronger than the hurricanes that spawn them (Thorncroft et al. 1993; Thorncroft and Jones 2000; Hart and Evans 2001). When a cyclone is generated over warm tropical SSTs, large sensible and latent heat fluxes at the sea surface can have important impacts on its development and intensification. Indeed surface heat fluxes during a cyclone’s very early development stages have a much greater influence on its ultimate intensity, compared to their influence when they occur during a later period (Kuo et al. 1991; Sanders 1986, 1987; Reed and Simpsons 1991). During later stages of the cyclone’s life, it may cross the Gulf Stream, move over colder waters, and undergo rapid baroclinically driven ET. Thus, the surface heat flux in the vicinity of the storm center becomes weaker and may reverse, making the ocean a heat sink, which often leads to a spindown of the storm’s circulation.

Most studies of coupled atmosphere–ocean systems concentrate on tropical cyclones and their ocean interactions in the low latitudes. The purpose of this study is to investigate the role of air–sea fluxes on extratropical cyclones and the upper ocean in the midlatitudes. We consider two cyclone case studies over the North Atlantic using an atmosphere–ocean model. These cyclones represent two quite different genres of storms that occur in this region: a hurricane undergoing ET, and an explosively developing winter storm. A brief description of the atmosphere–ocean model is presented in section 2. Storm cases are described in section 3. The experimental design of our simulations is presented in section 4. Numerical simulation results are discussed in sections 5 and 6, and conclusions are given in section 7.

2. Model description and experimental design

All simulations are performed using the Canadian Mesoscale Compressible Community (MC2) atmospheric model coupled to the Princeton Ocean Model (POM), passing SST from POM to MC2 and momentum, heat, and moisture fluxes from MC2 to POM.

a. Atmospheric model

The MC2 model is a state-of-the-art fully elastic non-hydrostatic model solving the full Euler equations on a limited-area Cartesian domain with time-dependent nesting of lateral boundary conditions. The latter are provided by a larger-scale model, MC2 uses semi-Lagrangian advection and a semi-implicit time-differencing dynamic scheme. As a modeling tool, MC2 is quite versatile, achieving successful simulations of extratropical cyclones (Benoit et al. 1997; McTaggart-Cowan et al. 2001, 2003).

The model domain is 24.25° to 56.25°N and 79.5° to 40°W, with a horizontal resolution of 0.25° and 30 vertical layers. We use latitude–longitude projection, and the integration time step is 600 s. Initial conditions and boundary conditions are determined from the Canadian Meteorological Centre (CMC) analysis fields as described by Chouinard et al. (1994). A force–restore scheme, described by Benoit et al. (1997), is used to calculate surface heat and moisture fluxes over land. The surface fluxes above the sea are calculated using Monin–Obukhov similarity theory. Deep cumulus convection is parameterized following Kain and Fritsch (1990, 1993).

b. Ocean model

POM (Blumberg and Mellor 1987) is used to simulate the oceanic component of our coupled model system. It is a three-dimensional, primitive equation model with complete thermohaline dynamics, using a sigma vertical coordinate and a free surface. POM is capable of simulating the coastal areas adjacent to the deep ocean, including the continental shelf and slope. A second-order turbulence closure scheme (Mellor and Yamada 1982) is used to represent the mixed layer dynamics. The model is implemented on the domain from 20° to 57.5°N and from 82° to 40°W with a horizontal resolution of 0.16°, using latitude–longitude projection. In order to accurately represent the cyclone-related mixed layer dynamics, 23 vertical layers are used, with higher resolution in the upper-ocean mixed layer (eight levels within the upper 80 m). Ocean topography is determined.
from the Earth Topography and Ocean Bathymetry Database (ETOPO5) at 5-min resolution, interpolated to POM’s model grid (U.S. National Geophysical Data Center, http://www.ngdc.noaa.gov/). The model’s minimum depth at the coastal boundary is 10 m. Radiative boundary conditions are prescribed for momentum and thermal variables at the lateral open ocean boundaries, as well as barotropic transports to correctly position the Gulf Stream (Mellor 1998).

The initial conditions and inflow open boundary conditions for ocean temperature and salinity are determined by the monthly averaged profiles from the Naval Oceanographic Office Generalized Digital Environmental Model (GDEM) following Bender and Ginis (2000). The spinup process includes two steps, starting with a null initial velocity. First the ocean model is integrated for 1 yr using the climatic monthly mean wind stress, heat flux, and freshwater flux from the analysis data produced by the National Centers for Environmental Prediction (NCEP). Second, the ocean model is integrated for a second year, using NCEP data for a given storm case, in order to get a more realistic prestorm ocean representation.

c. Coupling technique

The model system exchanges information between the atmosphere and ocean at the air–sea interface at every coupling time step. Each storm simulation begins with the integration of MC2 for 50 min with fixed SST from CMC. Wind stress, sensible and latent heat fluxes, radiative flux, and freshwater flux from MC2, are passed to POM. POM is then integrated for 50 min, which represents two time steps for its baroclinic mode. A new SST field, as produced by POM, is then passed to MC2 and it is then integrated forward for 50 min. Because of the different horizontal resolutions of the two models, data exchange between MC2 and POM is accomplished through interpolation. NCEP data are used in regions where the domains for MC2 and POM do not coincide.

3. Storm cases

We consider two midlatitude cyclones representative of the storms that occur in the Northwest Atlantic: the ET phase of ex-Hurricane Earl (September 1998) and an intense winter storm, Superbomb (January 2000).


Tropical Storm Earl formed 930 km south-southwest of New Orleans at 1800 UTC on 31 August, subsequently reaching hurricane status 230 km south-southwest of New Orleans at 1200 UTC on 2 September. The storm thereafter rapidly intensified to category 2 (winds in excess of 42 m s$^{-1}$) and made landfall near Panama city, Florida, as a category 1 hurricane at 0600 UTC on 3 September. The system continued its northeastward track across the Carolinas and continued to fill. Exiting the continental United States north of Cape Hatteras, North Carolina, just after 1200 UTC on 4 September, the storm lay 350 km off the New Jersey coastline by 0000 UTC on 5 September with a central sea level pressure (SLP) of 1000 mb as analyzed by the National Hurricane Center (NHC).

The subsequent 36 h saw ex-Hurricane Earl track northeastward and reintensify to 964 mb in the waters 100 km northeast from Newfoundland, Canada’s Avalon Peninsula. McTaggart-Cowan et al. (2001) suggest that a strong midlevel trough interacted with the system over this period, contributing to a rapid spinup of the lower-level vortex. A cyclonic vortex rollup was observed during reintensification, suggesting that baroclinic processes were prevalent. At 1200 UTC on 6 September, Earl made a sharp anticyclonic track change, thus completing recurvature, and accelerated to the eastern North Atlantic, where it merged with remnants of Danielle. The central SLP, 1000–500-hPa thickness, and 10-m winds ($U_{10}$) for Earl’s extratropical phase are shown in Figs. 1 and 2, from the coupled MC2–POM simulation.

b. Superbomb (2000): Rapid oceanic cyclogenesis

Superbomb’s genesis can be traced to a CMC analysis that placed a surface center in southern Illinois on 20 January, with a reduced SLP of 1006 mb. Over the following 12 h, westerly flow to the south of the circulation center began to feel the warm (20°–24°C) coastal waters south of Cape Hatteras. Exiting the coast, the system’s circulation quickly extended southeastward as warmer marine boundary layer air wrapped rapidly around it. A secondary center superposed on a warm (>24°C) Gulf Stream eddy amid warm localized mixed layer temperatures. Associated upper-level forcings including stratospheric vorticity advection and baroclinity are additional factors contributing to the cyclogenesis that developed and the absorption of the original circulation center by the incipient marine system, subsequently leading to explosive cyclone generation.

The newly formed center began to track rapidly northeastward parallel to the coast from 1200 UTC on 20 January to 1200 UTC on 21 January, deepening from 997 to 955 mb. At midlevels, the CMC analysis suggests that a short wave traveled quickly around the base of a deepening larger-scale trough centered over Hudson Bay. The system was still under the influence of a jet to its south and west, thus benefiting from the divergent upper-level forcing associated with the left-exit region. Phase locking occurred shortly after this time, and the vertically stacked system began to curl northward under the influence of the flow at midlevels. The rate of intensification leveled off, and central SLP remained at 956 mb until 0000 UTC on 22 January. The system made landfall in Cape Breton, Nova Scotia, Canada, shortly before 0000 UTC on 22 January and tracked
northwestward to the Gulf of St. Lawrence, as the central SLP weakened to 972 mb. Filling continued as it moved toward Baffin Bay, and ultimate dissipation. Figures 1 and 2 show Superbomb’s central SLP, 1000–500-hPa thickness, and $U_{10}$ fields in comparison with Earl.

4. Experimental design

Numerical experiments are described in Table 1. Simulations denoted uncoupled MC2 are control runs, carried out with MC2, using CMC analysis data to specify fixed SSTs during the integration period. Coupled model simulations use the coupled MC2–POM model system. Simulations denoted coupled model with no wind stress, are performed using the MC2–POM coupled model system with no wind stress to drive POM. Simulations denoted, coupled model with no heat flux, use the MC2–POM coupled system with no heat fluxes to drive POM.

Differences between coupled and uncoupled simulations are obtained by comparing MC2 control and coupled MC2–POM simulations. The ocean response to wind stress is obtained by comparing MC2–POM coupled simulations, with and without wind stress. The ocean response to heat flux is obtained by comparing
Table 1. List of numerical experiments. Ex-Hurricane Earl was simulated from 0000 UTC 5 Sep to 1200 UTC 7 Sep 1998, Superbomb, from 0600 UTC 20 Jan to 1200 UTC 22 Jan 2000.

<table>
<thead>
<tr>
<th>Experiment name</th>
<th>Description</th>
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<tbody>
<tr>
<td>Ex-Hurricane Earl</td>
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<tr>
<td>EARL-un</td>
<td>Uncoupled MC2</td>
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<tr>
<td>EARL-coup</td>
<td>Coupled MC2-POM model</td>
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<tr>
<td>EARL-no-heat</td>
<td>Coupled MC2-POM model; does not use heat flux to drive POM</td>
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<tr>
<td>EARL-no-wind</td>
<td>Coupled MC2-POM model; does not use wind stress to drive POM</td>
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<tr>
<td>Superbomb</td>
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<td>BOMB-un</td>
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Fig. 2. As in Fig. 1, comparing $U_{10}$ winds (m s$^{-1}$), for (left) Earl and (right) Superbomb.
MC2–POM coupled simulations, with and without heat flux. Comparisons of Earl and Superbomb give differences in the storms, oceanic responses, mixed layers, and ocean stratifications.

5. Experimental results: Ex-Hurricane Earl

The setup and results of the control simulation (EARL-un in Table 1) is very similar to that of McTaggart-Cowan et al. (2001), and most of the small differences are attributable to recent improvements in MC2. Comparisons between EARL-un and the NHC storm track are shown in Fig. 3. Slight initial differences in the central SLP on 0000 UTC 5 September result from analysis discrepancies between the NHC and CMC analysis low positions. Thereafter, storm propagation speed in the MC2 control is very similar to that of NHC analysis, as Earl progresses along a preexisting baroclinic zone along the eastern seaboard. Crossing eastern Newfoundland (0600 UTC 6 September), the model simulation compares well with an analysis by McTaggart-Cowan et al. (2001).

a. The oceanic response

Earl is a typical ET, beginning over warm subtropical waters and passing to colder waters north of the Gulf Stream, with shallow mixed layers. We discuss SSTs, the temperature profile, and currents in the upper ocean.

1) SSTs and storm track

The SST cold wake is weak during our coupled (MC2–POM) simulation of Earl’s early midlatitude ET development stage, because Earl propagates rapidly (>15 m s⁻¹) during the first 24 h of the simulation (Fig. 4a). During the second day of the simulation (to 48 h), Earl’s propagation slows considerably (~5 m s⁻¹), and the cold wake becomes widely distributed, occupying most of the 40°–55°N region (Fig. 4b). The maximum SST cooling is almost 5°C and occurs to the right of the storm track, over the Grand Banks and related waters. This cooling trend can be verified from satellite SST data from The Johns Hopkins University (http://fermi.jhuapl.edu/avhrr). Associated oceanic mixed layer currents, produced by Earl’s cyclonically rotating, asymmetric winds (Fig. 2), are also biased to the right side of the cyclone track (Figs. 4a,b).
2) Upper-ocean Responses

The storm-induced SST cold wake is not simply an ocean surface feature. Figure 5a shows the ocean temperature structure along 43°N, at 0000 UTC on 5 September, when Earl is south of the Gulf Stream. This is a typical autumn temperature profile for this latitude, with a shallow mixed layer, colder water underneath, and a sharp temperature gradient in the upper thermocline. Figure 5b shows the associated impacts on the upper-ocean temperature from the coupled simulation, EARL-coup. At this time, the storm is at (49.75°N, 49.75°W), and moving eastward. A cooling zone extends ~500 km to the right of the storm track (shown in Fig. 4b), to 10-m depth, with warming in deeper (15 to 50 m) waters, by as much as 1.5°C (shown in Fig. 5b). The latter is consistent with Price’s (1981) results that entrainment causes cooling in the mixed layer and warming at depths below the initial mixed layer.

Storm-induced currents are a dominant mechanism in forming the storm’s SST depression. Because of the asymmetry of Earl’s wind fields (Fig. 2), the strongest currents are produced to the right of the storm center. This is shown in the coupled simulation (EARL-coup) in Fig. 4, with speeds up to 1.8 m s⁻¹, which is similar to currents generated by tropical storms (Bender and Ginis 2000). The horizontal component of these currents is shown in Fig. 5c, as a function of depth along 43°N, at 24 h. Earl’s impact on current shear extends to ~500 km in the cross-track direction (shown in Fig. 4b) with the strongest storm-induced currents in the upper 15 m (Fig. 5c).

Time series of screen-height air temperature, T_a, U_10, SST, and SLP, are shown in Fig. 6 for 45.25°N, 48.5°W, where the SST depression due to Earl is greatest (see Fig. 4b). Passage of the storm corresponds to minimum
SLP and maximum $U_{10}$, followed by SST and $T_a$ depressions occurring about an inertial period later ($\sim 12$ h). The associated time series for ocean temperature and horizontal and vertical currents is given in Figs. 7a–c. The role of the storm-induced vertical currents to depress SST and modify the ocean temperature profile is notable. Although the horizontal and vertical currents are weak prior to Earl’s passage, Earl causes strong upwelling under the mixed layer and vertical current shear within the upper 50 m. Thus, entrainment is enhanced, colder water beneath the mixed layer is mixed with warmer water in the mixed layer, and SSTs are cooled. This is consistent with Jacob et al. (2000) and De Young and Tang (1990).

3) THERMAL AND DYNAMIC PROCESSES

Coupled model simulations allow consideration of the competitive roles of thermal and dynamic processes. This is achieved by comparing the EARL-no-heat and EARL-no-wind simulations. The former does not use MC2 heat fluxes to drive POM; the latter does not use MC2 wind stress to drive POM. Results are shown in Figs. 8a and 8b. When the surface heat fluxes are disconnected, the resultant SST depression and surface currents are almost as strong as those of the fully coupled simulation (EARL-coup) after 48 h, as shown in Fig. 4b. Moreover, surface current patterns are quite similar in the simulations of Figs. 4b and 8a, showing the dominance of Earl’s cyclonically rotating, asymmetric winds. On the other hand, when the wind stress is disconnected, no significant cold wake or surface currents develop. Therefore, the dynamic processes represented by wind stress, dominate over thermal processes represented by heat fluxes.

b. Effects on atmospheric development

The processes described in the previous section, such as the time-varying SST depression along the storm track; represent negative feedback in the ocean–cyclone system. This can be investigated using the coupled MC2–POM simulations. Kuo et al. (1991) suggested that ocean surface fluxes could potentially have an impact on storm intensity, depending on the phase of storm life cycle in which they are felt. In turn, impacts on
storm intensity can potentially modify the wind stress and thermal fluxes that determine the upper-ocean response. Bender and Ginis (2000) and Bao et al. (2000) showed that tropical-mode hurricanes with slow propagation speeds and large SST cooling can experience large ocean impacts.

1) IMPACTS ON STORM INTENSITY

In Earl’s case, we have shown (Fig. 3) that ocean coupling (MC2–POM) does not appreciably shift the storm’s track, for example, in response to increased lower-level stability over cooled SSTs. Moreover, in terms of storm intensity, the maximum impact of ocean coupling on Earl’s central SLP and $U_{10}$ is small. It occurs at about the maximum storm intensity, about 36–48 h into the simulation (Fig. 9). The reduced intensity (~3 hPa or 3 m s$^{-1}$) reflects SST cooling and is a negative feedback during Earl’s intensification. This is determined by factors such as Earl’s rapid propagation over warm SSTs deceleration and slower propagation over relatively colder SSTs.

2) RESPONSE OF THE LOWER ATMOSPHERE

In order to estimate the impact of the ocean on the lower atmosphere, it is important to consider the relationship between the SST depression, lower-level atmospheric winds, and related variables. SST depression cools the atmospheric surface layer, resulting in a spatial $T_a$ distribution that is quite similar to the SST results shown in Fig. 8a (and also Figs. 4a,b). This figure was obtained by comparing the fully coupled simulation, EARL–coup, to the partially coupled simulation, EARL–no-wind, which passes heat fluxes from MC2 to POM, but no wind stress. Corresponding to the SST effects, ocean impacts on specific humidity are given in Figs. 10a and 10b, and on winds $U_{10}$ in Figs. 11a and 11b.

Decreases in the (2 m) specific humidity imply that the atmosphere takes up less moisture and the latent
FIG. 10. The difference in specific humidity $\Delta q$ at 2 m ($\text{g kg}^{-1}$) between the EARL-coup and EARL-no-wind simulations, at (a) 24 and (b) 48 h. The black cross is the location of the cyclone center, from EARL-coup. Storm track from EARL-coup simulation indicated by the black line.

FIG. 11. As in Fig. 10, for differences in $U_{10}$ winds (m s$^{-1}$). Cross sections A–B and C–D are used in Fig. 13.
heat flux from the ocean must also decrease. The corresponding $U_{10}$ fields show that the impact of ocean coupling on $U_{10}$ is small during the first day, in accord with the small storm-induced SST cooling. After 2 days, the negative impact on winds increases slightly ($\sim 3 \text{ m s}^{-1}$) and shows a degree of coincidence with the SST depression regions (see Figs. 4b and 8a). The tendency is to decrease storm intensity, expressed in terms of $U_{10}$ fields, to the right of the storm track. These figures show that cooler SSTs and lower specific humidity correspond to a weaker cyclone, expressed in terms of $U_{10}$. Cooler SSTs and less latent heat imply a lower-level spindown resulting from increased stability and reduced coupling with the low-level jet.

Comparisons of the associated latent and sensible heat fluxes from MC2 control and coupled MC2-POM simulations (EARL-un and EARL-coup) are shown in Figs. 12a–d. As expected, latent heat flux constitutes a dominant factor in the coupling of atmosphere and ocean. The impact of ocean coupling is to reduce the latent heat fluxes in the EARL-coup simulation compared to those of EARL-un. Moreover, SSTs and Earl’s propagation speed are important factors affecting the ocean’s impact on latent heat fluxes. While sensible and latent heat flux have similar distributions, Earl propagates so fast that after 24 h it has moved over cold water and both latent and sensible heat fluxes are negative in the storm-center neighborhood. Although extensive positive sensible heat fluxes remain, they are confined to the rear of the storm and do not strongly affect further storm development.

The reaction of the lower atmosphere to SST cooling is strong. Smaller sensible and latent heat fluxes from the ocean surface tend to cool and dry the atmospheric boundary layer in the coupled simulation EARL-coup, compared to the control EARL-un. Concomitantly, the $U_{10}$ winds decrease and the cyclone tends to weaken in the coupled simulation. However, because most of the significant interactive processes occur during the peak intensification period, after the initial 24 h in our simulation of Earl, the ocean impact on intensity is not large.

Fig. 12. Uncoupled simulation (EARL-un) at 24 h of (a) latent heat flux and (b) sensible heat flux. Coupled simulation (EARL-coup) at 24 h of (c) latent heat flux and (d) sensible heat flux. Units: W m$^{-2}$.
3) THE MIDLATITUDES

Wallace et al. (1989) suggested that as an air mass moves over cold water, the atmospheric boundary layer stabilizes because of negative sensible surface heat fluxes. Vertical mixing is suppressed, the vertical wind shear increases, and as surface drag effects decrease, wind speeds aloft increase. Our results are consistent with these studies.

Figure 13 shows cross sections of differences of temperature $\Delta T$, geopotential height $\Delta G$, and horizontal wind speed $\Delta U$, between the simulations EARL-coup and EARL-no-wind. These cross sections are taken through the storm center (49.75°N), at 48 h when SST cooling has developed. The impact of MC2–POM coupling on the vertical atmospheric structure is visible in the middle of the troposphere, although at reduced strength compared to its impact at lower levels, and disappears in the upper troposphere. Within the boundary layer (approximated >800 hPa), the cyclone is cooler and winds weaken in the EARL-coup simulation, compared to EARL-no-wind. In the midtroposphere, the cyclone becomes warmer by about 0.5°C and more intense, with slightly lower geopotential height and stronger winds surrounding the storm center. Thus, ocean impacts in the midtroposphere are opposite to their atmospheric boundary layer effects.
6. The 2000 Superbomb: Rapid winter cyclogenesis

Superbomb is typical of a series of winter storms that tracked along the North American east coast. It is notable because it developed explosively, following a storm track that was similar to that of a previous storm (hereafter denoted “previous storm”). At the time of our Superbomb simulation initialization, the previous storm was located at 41.25°N, 56.5°W, as indicated in Figs. 1 and 2, with a central SLP of 971 hPa. Thereafter, it propagated northeastward, reaching a central SLP of 966 hPa at 42.25°N, 54.5°W after 6 h. Continuing along this storm track it slowly filled and moved out of our model domain.

a. The oceanic response

The setup and control simulation of Superbomb is identical to that of Li et al. (2003), extending for 54 h, from 0600 UTC 20 January until 1200 UTC 22 January. We consider oceanic responses in conjunction with associated atmospheric impacts.

1) SSTs and storm track

Comparisons between the control simulation (BOMB-un), the coupled simulation (BOMB-coup), and the CMC analyzed Superbomb track are shown in Fig. 14. In the simulations, the storm progresses parallel to the eastern seaboard at a rate similar to the analyzed system except for a minor perturbation at 0600 UTC 21 January. From simulations with and without ocean coupling, estimated tracks are close to the CMC analysis and converge near the end of the cyclone life cycle.

Corresponding SST impacts and surface currents are given in Figs. 15a and 15b, including the location of the previous storm that occurred a few days before Superbomb and developed along a similar storm track. The SST cooling along Superbomb’s track does not exceed 1°C because of the deep oceanic mixed layer. Additional modest SST cooling occurs to the southeast of Newfoundland, because of the previous storm. Although Superbomb’s SST cooling is modest, significant surface currents are generated along its storm track, with some bias to the right side of the track because of its strong, asymmetric winds (Fig. 2).

2) Upper-ocean responses

During the autumn, the North Atlantic oceanic mixed layer is generally shallow, whereas in the winter, it is much deeper. In the former case, a shallow mixed layer allows the upper ocean to be easily stirred with the...
Fig. 16. Longitude–depth sections from the BOMB-coup simulation along 43°N (line A–B in Fig. 15a) for (a) initial ocean temperature $T_0$ (°C), (b) temperature difference at 36 h minus initial state $\Delta T_0$ (1°C contours), (c) horizontal current $U$ at 36 h (0.15 m s$^{-1}$ contours), and (d) vertical current $w$ at 36 h ($1 \times 10^{-4}$ m s$^{-1}$). Black dot indicates that at 36 h Superbomb is at 43.0°N, 62.5°W. White colour in (a) and black in (b)–(d) are the model ocean bottoms.

passage of an intense storm. Thus Earl can generate a much stronger cold wake than that produced by Superbomb, although Superbomb’s storm-induced surface currents are similar in magnitude to those of Earl.

The mixed layer depth during Superbomb’s development is in excess of ~500 m, as shown in the depth–longitude section at 43°N in Fig. 16a. The vertical and horizontal oceanic temperature gradients are small in the vicinity of the cyclone, and indeed over most of the ocean domain, accounting for the lack of any strong SST cooling. Although the largest SST cooling is about 1°C, a significant area of warming occurs in the region 40° and 45°N, with peak values of nearly 3°C. The SST response to Superbomb results from a combination of factors. First, there is a strong horizontal ocean temperature gradient defining the Gulf Stream’s northern edge. Second, there is a warm water anomaly (Fig. 16a) dominating the upper ocean and centered at about 100-m depth, resulting in the SST warming region within about 300 km to the east of the storm track (Figs. 15b and 16b). The associated horizontal and vertical current speeds are given in Figs. 16c and 16d, with values as high as 1 and $4.5 \times 10^{-4}$ m s$^{-1}$, respectively. They suggest that the local SST warming is due to vertical mixing.

Time–depth sections of ocean temperature and horizontal and vertical current speeds are shown in Fig. 17, at the location with the largest SST warming (43°N, 57°W). This location has an effective shallow mixed layer with strong vertical temperature gradients, because of the warm water anomaly at 100 m (Fig. 16a). The two black dots indicate the previous storm and Superbomb. These two storms (Fig. 15a) generate strong inertial oscillations, upwelling, and entrainment, which result in the upward transport of warmer water from the warm anomaly beneath the surface, and warmer SST. This process differs from that occurring in much of the rest of the ocean, where a deep mixed layer and small vertical temperature gradients allow only small SST changes. Associated time series for $T_0$, $U_10$, and SLP are given in Fig. 17d, for location 43°N, 57°W. Superbomb’s pas-
sage occurs at a locally minimum SLP, maximum $U_{10}$, with a SST warming a few hours later.

3) Sea Surface Fluxes

Latent and sensible heat fluxes from the uncoupled simulation (BOMB-un) are compared to those of the coupled simulation (BOMB-coup) in Fig. 18. This figure shows that latent heat fluxes are larger than sensible heat fluxes, and the impact of ocean coupling is to make the BOMB-coup heat fluxes less powerful than those of the BOMB-un simulation, as in the simulations of Earl (Fig. 12). It is notable that while both fluxes have similar spatial distributions, and Superbomb’s simulated initial propagation speed is less than Earl’s, Superbomb still propagates rapidly enough that at 24 and 48 h it has moved over cold water, and both latent and sensible heat fluxes are negative in the neighborhood of the storm center. Extensive positive heat fluxes remain, but as they are confined to the rear of the storm, they no longer strongly influence storm development.

Latent heat fluxes for Superbomb are much larger than Earl’s. Moreover, sensible heat fluxes are also much larger than those of Earl (Fig. 12). This follows from the fact that Superbomb involves a much cooler, drier air mass with higher wind speeds than Earl. The MC2 interfacial fluxes at the sea surface are computed using Monin–Obukhov similarity theory, which leads to a bulk turbulent flux formulation for the interfacial momentum and heat fluxes. Key factors in the sensible heat flux are the air–sea temperature difference $\Delta T$ and $U_{10}$. For the latent heat flux, key factors are the specific humidity difference $\Delta q$ and $U_{10}$. When heat fluxes are maximal in Superbomb, $\Delta T$ is about 15°C, $U_{10}$ is about 25 m s$^{-1}$, and $\Delta q$ is slightly greater than 10 g kg$^{-1}$. By comparison, maximal heat fluxes for Earl occur when $\Delta T$ is less than 5°C, $U_{10}$ is about 12 m s$^{-1}$, and $\Delta q$ is slightly less than 10 g kg$^{-1}$. Thus, heat fluxes are greater for Superbomb.

To consider the competition between thermal processes and dynamic processes, we compare the partially coupled simulations Bomb-no-heat (does not pass MC2 heat fluxes to POM) and Bomb-no-wind (does not pass MC2 wind stress to POM). In the former case (Fig. 19a), the SST and surface current fields are very similar, in strength and in current patterns, to those of the fully coupled model (see Fig. 15b), particularly between 40° and 45°N, showing the dominance of Superbomb’s asymmetric winds. This is analogous to results obtained
from the Earl simulations (Fig. 8a), although weaker. In the latter case (Fig. 19b), essentially no SST cooling or surface currents develop. Therefore, wind-stress-related dynamic processes dominate over heat flux processes in the upper-ocean, as in Earl. It is notable that a slight SST warming occurs to the northeast of Newfoundland (Fig. 19a) in the BOMB-no-heat simulation, and a much more extensive warming (Fig. 19b) occurs in the BOMB-no-wind simulation. This suggests that heat flux coupling is the key to this warming effect and reflects an effect from the storm previous to Superbomb. Analogous results are shown for Earl (Fig. 8b).

b. Effects on atmospheric development

Superbomb originates and develops entirely within the midlatitudes. It is more intense than Earl in terms of central SLP (Fig. 20a), with maximum $U_{10}$ in excess of 40 m s$^{-1}$ (Fig. 20b). Although it was first evident over the warm waters east of Cape Hatteras where SSTs...
Fig. 19. SST differences between the 48-h simulation minus initial state, using partially coupled models: (a) BOMB-no-heat and (b) BOMB-no-wind. Arrows are surface currents in m s$^{-1}$. Superbomb location is indicated by $\mathbf{H}$.

Fig. 20. Comparison of Superbomb (a) center SLP (hPa) and (b) maximum $U_{10}$ wind (m s$^{-1}$), from simulations with ocean coupling (BOMB-coup) and without ocean coupling (BOMB-un), in comparison to (a) CMC SLP and (b) NCEP/Quikscatterometer blended winds.

exceed 20°C, its trajectory takes it over cold northern waters with SSTs as low as 0°C. Figures 20a and 20b show that the maximum ocean coupling impact on $U_{10}$ occurs about 12 h prior to Superbomb’s central SLP, well within the rapid intensification phase of the storm. In magnitude, the atmosphere–ocean coupling effect on winds is about 4 m s$^{-1}$; on central SLP, it is about 4 hPa. However, the ocean coupling impact on Superbomb’s SSTs is negligible, compared to Earl, because of the deep oceanic mixed layer. Thus, the SST feedback to the atmosphere is small, despite Superbomb’s slow propagation speed and higher wind speeds, compared to those of Earl.

To consider the relationship between SST change and lower-level atmospheric winds, we compare the coupled BOMB-coup simulation to the partially coupled simulation, BOMB-no-wind, in Fig. 21. This is for longitude 43°N passing through the maximum SST warming region (Figs. 16a–d) after 48-h simulation. Figure 21 shows that $T_a$ correlates strongly with SST, for both the

Fig. 21. Differences between BOMB-coup and BOMB-no-wind after 48-h simulations, for SST (°C), $T_a$ (°C), $U_{10}$ wind (m s$^{-1}$), and specific humidity, $Q_a$ (g kg$^{-1}$) on 43°N (line A–B in Fig. 15a).
Fig. 22. Vertical profiles of differences in (a) temperature $\Delta T$ (°C), (b) specific humidity $\Delta Q$ (g kg$^{-1}$), and (c) horizontal wind $\Delta U$ (m s$^{-1}$), for BOMB-coup minus BOMB-no-wind at 43°N, 50°W after 48-h simulation. This location experiences the greatest SST cooling in Fig. 21. Profiles are averaged for 200 km $\times$ 200 km areas.

A warming region between 53° and 57°W and the cooling region, 47°–53°W, from the storm that preceded Superbomb. The SST and $T_a$ warming and enhanced $U_{10}$ (~1 m s$^{-1}$) suggest locally enhanced latent heat fluxes from the ocean to the atmosphere. However, this region is too small and too remote from the storm track to influence the storm’s development and intensity.

The adjacent cooling region (47°–53°W) experiences modest SST and $T_a$ depression, decreased $U_{10}$, and reduced specific humidity, suggesting a reduction in the local ocean region’s ability to flux moisture into the atmosphere and local weakening in storm intensity. These temperature changes result in hydrostatic height changes and associated wind adjustments that are largely restricted to the lower atmosphere and boundary layer variables. Figures 22a–c show vertical profiles for differences in temperature $\Delta T$, specific humidity $\Delta q$ and horizontal wind $\Delta U$ between BOMB-coup minus BOMB-no-wind runs. This implies that the ocean impacts on Superbomb’s atmospheric temperature, specific humidity, and winds are notable at the surface but become progressively weaker with increasing height. For all three variables, ocean effects are negligible above 700 mb.

7. Conclusions

This study is concerned with the implications of using a coupled atmosphere–ocean model to simulate intense northwest Atlantic storms and upper-ocean responses. Our model system consists of the MC2 atmospheric model coupled to the POM by air–sea fluxes of heat, momentum, and mass. To illustrate the impacts of ocean surface processes, we consider two North Atlantic storms. Hurricane Earl (1998) represents an autumn ET, with a rapid propagation speed and associated with a warm, shallow oceanic mixed layer. Superbomb (2000) was an explosively developing slow-moving North Atlantic winter cyclone with a cold deep mixed layer. In both Earl and Superbomb, temperatures range from the comparatively warm Gulf Stream waters to comparatively cold northern waters; latent and sensible heat fluxes are enhanced over Gulf Stream waters, while waters north of the Gulf Stream are a heat sink.

Our results suggest that feedbacks in the coupled midlatitude cyclone–ocean system are enhanced by a thin oceanic mixed layer, slow storm propagation speed, strong thermal stratification below the oceanic mixed layer, and low relative humidity in the atmospheric boundary layer. Impacts of coupling include SST cooling and surface heat fluxes, resulting in modification of the cyclone’s intensity, in accord with related studies by Kuo et al. (1991), Schade and Emanuel (1999), and Chan et al. (2001).

In the coupled simulations of Earl, an SST cold wake develops with a maximum cooling of 5°C in the right-rear quadrant of the storm, consistent with satellite SST observations. Associated cyclonically rotating winds produce oceanic mixed layer currents that are biased to the right side of the cyclone track because of Earl’s asymmetric wind structure. SST cooling is mainly due to entrainment of cooler water from beneath the mixed layer, and mixing associated with the strong vertical shear. Stronger oceanic responses occur in Earl during the second and third integration days, because the propagation speed decelerates at that time and because Earl’s near-surface winds have increased dramatically. Oceanic impacts on the atmosphere also become more important during the second and third integration days, weakening Earl’s central SLP by about ~3 hPa, and $U_{10}$ by about ~3 m s$^{-1}$, compared to the uncoupled simulation. Coupling Earl with the ocean results in a cooler,
dryer more stable atmospheric boundary layer, and thus a less intense cyclone. The impact is also felt in the mid troposphere, although at reduced magnitude compared to boundary layer effects. The mid troposphere becomes slightly warmer, with slightly lower geopotential height and higher winds.

While impacts of ocean coupling on Superbomb’s storm track are minor, impacts on central SLP (≈4 hPa) and maximum $U_{10}$ (≈4 m s$^{-1}$) are notable. Wind-induced surface currents are as significant as those of Earl, because Superbomb’s winds are hurricane strength (≥40 m s$^{-1}$). However, the deep mixed layer allows for the development of only very mild SST cooling compared to Earl.

The impact of ocean coupling is to reduce sensible and latent heat fluxes, in both Earl and Superbomb. However, localized warming does occur in Superbomb as a result of the presence of a local warm water mass centered at 100-m depth between the Scotian Shelf and the Grand Banks. This feature is typical for this region of the northwest Atlantic over the winter months, and it creates an apparent local shallow mixed layer with strong vertical temperature gradients. Resultant localized SST warming is accompanied by localized near-surface air temperature warming, modest enhancement to near-surface winds, and enhanced air–sea latent heat fluxes, during Superbomb. However, since these ocean impacts are largely relegated to a localized region of the atmospheric boundary layer (approximated as >800 hPa), and occur far from the storm track, they do not substantially impact Superbomb’s development and intensity.

The coupling of the ocean–atmosphere system results in important structural modifications for both Earl and Superbomb. For Earl, these changes occur in boundary layer, SST, and the upper ocean, in response to a shallow mixed layer and high-wind forcing of the intensifying storm. Since synoptic-scale processes provide much of the impetus for the system’s baroclinic-mode redevelopment, the cooled ocean surface plays only a secondary role in determining atmospheric structures. Conversely, the differences between the coupled and uncoupled simulations of Superbomb occur primarily in the lower atmosphere as rapid boundary layer destabilization processes dominate. The deep, stable winter oceanic mixed layer in the coupled simulation prevents significant atmospheric feedbacks, particularly north of the Gulf Stream.

Although these storm cases show differences in both the atmospheric and oceanic responses to the coupled modeling system, the relevance of coupling concepts to mesoscale modeling and severe storms diagnosis has been shown. In the case of either Earl or Superbomb, model comparisons show that the thermodynamic structure of the upper ocean beneath the cyclone is affected more by the wind than by the sensible and latent heat fluxes at the ocean surface.

Acknowledgments. This study was funded by the Panel on Energy Research and Development (PERD) of Canada Projects on Severe Storms and Waves. A Canada Foundation for Climate and Atmospheric Studies (CFCAS) grant to John Gyakum is gratefully acknowledged. We also want to especially thank Prof. Issac Ginis for providing the University of Rhode Island version of POM, which formed the basis for the ocean component of this investigation.

REFERENCES


Reed, R. J., and A. J. Simmons, 1991: Numerical simulations of an explosively deepening cyclone over the North Atlantic that was unaffected by concurrent surface energy fluxes. Wea. Forecasting, 6, 117–122.


Wright, R., 1969: Temperature structure across the Kuroshio before and after typhoon Shirley. Tellus, 21, 409–413.
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