# Global Ocean Surface Wave Simulation Using a Coupled Atmosphere–Wave Model

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#### ABSTRACT

This study describes a 29-yr (1981–2009) global ocean surface gravity wave simulation generated by a coupled atmosphere–wave model using NOAA/GFDL's High-Resolution Atmosphere Model (HiRAM) and the WAVEWATCH III surface wave model developed and used operationally at NOAA/NCEP. Extensive evaluation of monthly mean significant wave height (SWH) against in situ buoys, satellite altimeter measurements, and the 40-yr ECMWF Re-Analysis (ERA-40) show very good agreements in terms of magnitude, spatial distribution, and scatter. The comparisons with satellite altimeter measurements indicate that the SWH low bias in ERA-40 reanalysis has been improved in these model simulations. The model fields show a strong response to the North Atlantic Oscillation (NAO) in the North Atlantic and the Southern Oscillation index (SOI) in the Pacific Ocean that are well connected with the atmospheric responses. For the NAO in winter, the strongest subpolar wave responses are found near the northern Europe coast and the coast of Labrador rather than in the central-northern Atlantic where the wind response is strongest. Similarly, for the SOI in the Pacific Ocean, the wave responses are strongest in the northern Bering Sea and the Antarctic coast.

#### 1. Introduction

Ocean surface gravity waves (OSGW) play a significant role in many physical processes at the air–sea interface. The momentum, heat, and moisture fluxes through the air–sea interface not only depend on the atmospheric state (i.e., the surface wind speed) but also on variations of the OSGW field (Fan et al. 2009a; Li and Garrett 1997; Grachev and Fairall 2001; Hanley and Belcher 2008).

Information on the variability in the wave climate also has a wide application to human activities in coastal

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areas. The Intergovernmental Panel on Climate Change (IPCC) Working Group (WG) II has recognized (Nicholls et al. 2007) that estimates of changes in the global wind– wave climate are one of the main requirements for the assessment of the effects of climate change on coastal erosion, risks to coastal population, and ecosystems. The IPCC WG I affirms (Christensen et al. 2007) that more information on projected wave conditions are required to enable assessments of the effects of climate change on coastal erosion. Additionally, changes to wave climate will have implications for a range of offshore operations like the oil industry and ship route planning.

Dynamical projections of wave climate are usually carried out through regional wave modeling studies, where downscaled global climate model (GCM) projections are used to force regional wave models. Forcing conditions are typically obtained from selected emission

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scenarios. After comparing several wind–wave reanalysis datasets against both buoy and altimeter measurements, Caires et al. (2004) point out that the differences between the wave datasets are larger than those between the wind speed datasets, and differences in wave datasets produced by different wave models exist even when the forcing winds used to produce the different wave reanalyses are the same. Thus, it is important to use the same model to generate both global wave climatology for the past and wave climate projection for the future. To our knowledge, no high-resolution global wave climatology and future wave climate projections have been produced by a single dynamical model, not to mention a coupled atmosphere-wave model.

Using an atmosphere–wave coupled model has additional advantages. 1) In the coupled model, the wind is passed to the wave model at every time step. When using a stand-alone wave model, saved winds, usually at a 3– 6-h interval, are used to force the wave model, and thus the wind field needs to be interpolated onto finer time intervals and thus be smoothed in time and space. 2) The wave model can feed back roughness to the atmospheric model and thus improve the boundary layer flux parameterization.

We have developed a high-resolution global simulation system by coupling the operational wave model, WAVEWATCH III, developed at the National Centers for Environmental Prediction/Environmental Modeling Center (NCEP/EMC; Tolman 1998), to the Geophysical Fluid Dynamics Laboratory (GFDL) High-Resolution Atmosphere Model (HiRAM) model (Zhao et al. 2009; Chen and Lin 2011). The aim of this paper is to evaluate the performance of this newly developed coupled model on wave climatology simulations. In this study, a 29-yr (1981-2009) Atmospheric Model Intercomparison Project (AMIP)-type simulation of wave climatology [including significant wave height (SWH), mean length, mean wave direction, period, etc.] is produced using this coupled system. This dataset is evaluated using available National Data Buoy Center (NDBC) buoy data, Ocean Topography Experiment (TOPEX)/Poseidon satellite measurements, and the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and is used to study the wave climate variability during the past 29 years. In a forthcoming paper, the same model is used to project the wave climate change in response to the SST/sea ice anomalies in the late twenty-first century generated by coupled models in the World Climate Research Program Coupled Model Intercomparison Project 3 (CMIP3) archive for IPCC Fourth Assessment Report (AR4).

The paper is presented in five sections. The coupled atmosphere–wave model is described in section 2.



FIG. 1. A schematic diagram of the atmosphere–wave coupled model. The arrows indicate the prognostic variables that are passed between the model components. In the diagram,  $z_0$ ,  $z_q$ , and  $z_h$ , are momentum, latent heat, and sensible heat roughness lengths, respectively. Also,  $T_{\text{air}}$ ,  $T_{\text{ice}}$ , and  $T_{\text{land}}$  are air, ice, and land temperatures at the surface, respectively.

Model evaluations against available observations are presented in section 3. Section 4 discusses the interannual wave climate variability. Finally, a summary is given in section 5.

#### 2. Methodology

A schematic diagram of the coupled atmospherewave model developed in this study is shown in Fig. 1. This coupled model makes use of the Flexible Modeling System (FMS; http://www.gfdl.noaa.gov/fms/) coupler for calculating and passing fluxes between its atmosphere, land, sea ice, and wave components. Both the atmospheric model and wave model have a time step of 20 min. At every time step, the atmosphere model exchanges fluxes with the land, ice, and wave model, and the ice model passes ice coverage to the wave model. The coupler computes and passes fluxes between the component models and does all the necessary regridding so that each component receives inputs and supplies outputs on its own grid. All fluxes are conserved to within machine precision.

## a. Atmospheric model

The atmospheric model used in this study is GFDL's HiRAM (e.g., Zhao et al. 2009; Chen and Lin 2011) designed for resolutions from 1 to 100 km, it features a finite-volume dynamical core (Lin 2004) formulated on a cubed-sphere grid (Putman and Lin 2007). The cubed-sphere grid is a projection of a cube onto the surface of a sphere represented as six adjoining equal-sized grid

faces. For this study, each face has 32 vertical levels and there are  $180 \times 180$  grid points on each level, corresponding to a horizontal resolution of ~50 km. As described in Zhao et al. (2009), compared to the GFDL Atmospheric Model, version 2.1 (AM2.1), this 50-km resolution version of HiRAM uses a simpler diagnostic cloud scheme and its convective parameterization is based on a closure for shallow convection by Bretherton et al. (2004), with much of the deep convection allowed to occur on resolved scales. The details of this shallow convection scheme are given in Zhao et al. (2009).

## b. Land and ice models

The GFDL Land Model, version 3 (LM3; P. C. D. Milly et al. 2012, unpublished manuscript) the land model coupled to HiRAM, is a new model for land water, energy, and carbon balance. In comparison to its predecessor, Milly and Shmakin (2002), LM3 includes more comprehensive models of snowpack, soil water, frozen soil–water, groundwater discharge to streams, and finite-velocity horizontal transport of runoff via rivers to the ocean. LM3 uses the same grid configuration as the atmospheric model.

When run in AMIP mode, HiRAM uses a simple sea ice model with implicit heat diffusion through a single uniform thickness ice layer with specified ice/ice-free coverage in a grid box (GFDL Global Atmospheric Model Development Team 2004).

#### c. Wave model

The coupled system utilized the surface gravity wave model, WAVEWATCH III (WWIII), developed and used operationally at the National Oceanic and Atmospheric Administration (NOAA)/NCEP (Tolman et al. 2002). It explicitly accounts for wind input, wave-wave interaction, dissipation due to white-capping and wavebottom interaction (negligible for deep-water waves, which is our focus of this study), and solves the spectral action density balance equation for directional wavenumber spectra. The wave spectrum of the model is discretized using 24 directions and 40 intrinsic (relative) frequencies extending from 0.0285 to 1.1726 Hz, with a logarithmic increment of f(n + 1) = 1.1f(n), where f(n) is the *n*th frequency. The intrinsic frequency is related to the wavenumber (magnitude) k through the dispersion relation. The wave model is built on latitude-longitude grid with a horizontal resolution of 1/2°. Because of the nature of this grid, unless an extremely small time step is used the model will generate instability at high latitude, thus the global simulation domain is cut off at 72°N at the northern boundary with prescribed artificial ice coverage north of it. Even though WWIII has been validated against observations over both global and regional-scale



FIG. 2. Scatterplot of momentum flux roughness  $z_0$  vs 10-m wind speed. The blue circles represent original WWIII  $z_0$  parameterization using Tolman and Chalikov (1996), the black circles represent  $z_0$  parameterized using Beljaars (1995), and the red circles represent  $z_0$  calculated by the wave model.

wave forecasts (Tolman 1998; Tolman et al. 2002), it was shown to overestimate wave heights under very high wind conditions in extreme tropical cyclones (Chao et al. 2005; Tolman and Alves 2005; Tolman et al. 2005). Thus, in the coupled system, the momentum roughness parameterization in WWIII is modified based on field observations as discussed in section 2d below and the appendix. The modified momentum roughness is also fed back to the atmospheric model as lower boundary conditions.

## d. Momentum roughness parameterization

The boundary conditions for the atmospheric model are the momentum, latent heat, and sensible heat fluxes at the surface. These fluxes are determined by Monin–Obukhov similarity theory and are in part determined by surface roughness lengths, which can be converted to drag coefficients, if desired, through the similarity theory. In this version of the wave-coupled model, the atmosphere sees only the momentum roughness produced by the wave model. In reality, wave breaking will lead to bubble generation and injects spray into the air, which affects latent heat flux and albedos. But for simplicity in this study, we do not use information from the wave model to modify the evaporation and sensible fluxes felt by the atmosphere.

The Beljaars (1995) formula has been used to parameterize the momentum, heat, and moisture roughness lengths in HiRAM over the ocean, as in AM2.1. The black circles in Fig. 2 show the scatterplot of momentum roughness length  $z_0$  over the ocean versus 10-m wind speed. We can see that  $z_0$  increases rapidly with wind speed at high winds. The blue circles give the original WWIII  $z_0$  parameterization using Tolman and Chalikov (1996), which is even higher than Beljaars (1995) and increase faster with wind speed. Over the past 20 years or so, field measurements, laboratory experiments, and modeling work have been conducted to study the behavior of momentum drag coefficient  $C_d$  over the ocean under different wind speeds. Using data from GPS dropsondes deployed in 15 hurricanes, Powell et al. (2003) estimated the drag coefficient at different wind speeds and reported that it would level off for wind speeds exceeding 30 m s<sup>-1</sup>. A similar relationship was established in laboratory experiments by Donelan et al. (2004), in which the drag coefficient levels off around  $33 \text{ m s}^{-1}$ , and remained approximately constant as the wind was increased up to  $50 \text{ m s}^{-1}$ . Measurements from the Coupled Boundary Layers Air-Sea Transfer (CBLAST) experiment using an instrumented aircraft in rain-free regions of two hurricanes also support the earlier studies that the drag coefficient does not continue to increase with wind speed, but rather levels off or even decreases (French et al. 2007).

Inspired by the observation findings, we have developed a new formula for  $z_0$  over the ocean as a function of both wind speed and nondimensional wave age (defined as  $c_p/u_*$ , where  $u_*$  is the friction velocity and  $c_p$ is the phase speed at the spectral peak frequency). The red circles in Fig. 2 shows the new  $z_0$  calculated by the wave model, which is similar to Beljaars (1995) at low winds but levels off at high winds. By utilizing this new roughness parameterization, we are able to simulate strong hurricanes at comparable intensities with observations, while the original HiRAM [which use the Beljaars (1995)  $z_0$  parameterization] was shown to be inadequate on simulating the intensity for strong hurricanes (Zhao et al. 2009). But, the difference in the monthly mean winds are not significant. More details about this new formulation are given in the appendix.

# e. Initialization and forcing

The free-running 29-yr (1981–2009) simulation in this study follows the typical setup for AMIP runs. In addition to prescribed SSTs from the Hadley Centre Sea Ice and SST version 1.1 dataset (HadISST 1.1; Rayner et al. 2003) as the lower boundary condition for the atmospheric model, the well-mixed greenhouse gases, volcanic aerosols, and both tropospheric and stratospheric ozone vary from year to year, following the procedure used in the GFDL Coupled Model, version 2.1 (CM2.1) historical simulations in the CMIP3 database (Delworth et al. 2006).

To generate the initial condition for the wave-coupled system, a 1-yr integration of the coupled system is conducted starting from 1 January 1980. For this 1-yr run, the wave model starts from calm sea; the atmospheric, and land initial conditions at 1 January 1980 are taken from the end of a 10-yr run of the HiRAM model that uses climatological SSTs. Note that the spinup time needed for the wave model by itself is less than a month, and is dominated by transit times of swell through the Pacific Ocean.

#### 3. Model evaluation

ECMWF has produced a 45-yr global reanalysis dataset from 1957 to 2002 (ERA-40). This dataset contains both atmospheric and OSGW information. The OSGW fields are calculated using the Wave Model (WAM) developed by the Wave Modeling Group established in 1984 on the initiative of K. Hasselmann (WAMDI Group 1988; Komen et al. 1994). The wave spectrum of WAM is discretized using 12 directions and 25 intrinsic (relative) frequencies. In addition to the several sets of wind observations assimilated into ERA-40, the European Remote Sensing Satellite-1 and -2 (ERS-1 and ERS-2) SWH altimeter measurements were also assimilated into the model for the period in which they are available (1992–2001). Hanley et al. (2010) judge the ERA-40 wave product as the best data currently available for global climatology of ocean wave processes and wind-wave interaction. The atmospheric/wave model used at ECMWF to produce ERA-40 is at roughly 1.125°/1.5° resolution, but the data freely available for both wind and wave is at 2.5° resolution, which is what we use for our model evaluation. Area-weighted averaging is used to average our model results at 0.5° resolution onto the ERA-40 grid for the comparison. We evaluate both the wind and wave field during a 21-yr period from 1981-2001 when our model results and ERA-40 jointly exist.

However, Caires et al. (2004) find that the ERA-40 data tends to underestimate SWH at wind speeds above 14 m s<sup>-1</sup>. Hemer et al. (2010a) also pointed out that as a consequence of the relatively coarse resolution of the atmospheric model and its limited ability to resolve storm systems, extreme wave heights are severely underestimated in ERA-40. Thus, we also compare with the corrected ERA-40 (C-ERA-40) SWH produced by Caires and Sterl (2005) for a better understanding of the model bias.

#### a. Wind speed

In evaluating the general climate simulated by this coupled system, it is particularly important to examine the 10-m wind speed because it is used to drive the wave model. The comparison of annual mean 10-m wind speed between our model and ERA-40, and the differences between the two are given in Fig. 3. From Fig. 3c, we can



FIG. 3. (a) Model and (b) ERA-40 annual mean 10-m wind speed for the period of 1981–2001, (c) their difference, and (d) model minus ERA-Interim annual mean 10-m wind speed.

see that the differences over the ocean are mainly found in four regions, where our model winds are higher than ERA-40. To understand the differences, we also calculated the 99th percentile wind speed from the 6-hourly wind data during the 21-yr period for both ERA-40 (Fig. 4a) and our model (Fig. 4c). These two wind fields have very similar global structure and the correlation between them is 0.87. However, our model 99th percentile winds are higher than ERA-40 everywhere over the global ocean, the higher the winds, the larger the differences are. This is consistent with the findings by Caires and Sterl (2003) that the ERA-40 has underestimations for high wind speeds. Notice that we can clearly see the presence of strong tropical storms in the west Pacific (between 0°-30°N in the North Pacific and between 0°-30°S to the east of Australia) from our model results. As described by Zhao et al. (2009), HiRAM produces a very realistic climatology of tropical cyclones, although it generates few storms stronger than category 2. On the other hand, ERA-40 hardly shows any signature of tropical storms at all. Apparently, this large discrepancy has resulted in the mean wind field differences in the north and southwestern Pacific (Fig. 3c).

The third difference area is found to be in the Gulf of Tehuantepec located in the east Pacific Ocean off the Mexico coast, where the strong gap winds are better resolved by our high-resolution model. Our model mean winds also appear to be larger in the Southern Ocean and at a higher latitude in the Northern Hemisphere. This difference is present throughout the year but is most pronounced during the winter [July-September (JAS) for the Southern Ocean and January-March (JFM) for the Northern Hemisphere], when the winter storms are most intense there (the maximum wind speed is around  $35 \text{ m s}^{-1}$  in our model). Since the ERA-40 wind speed bias are different for different latitude regions and are particularly strong for high latitude in both hemispheres (Caires and Sterl 2003), we also calculated the difference between our model winds and the ERA-Interim reanalysis (Dee et al. 2011) in Fig. 3d. Notice that our model winds compare very well with ERA-Interim at high latitude for both hemispheres, but large differences still exist



FIG. 4. ERA-40 99th percentile (a) wind speed, (b) significant wave height, and model 99th percentile (c) wind speed and (d) significant wave height during the 21-yr period (1981–2001).

in the Gulf of Tehuantepec and the tropical cyclone active regions in the Pacific Ocean.

# b. Significant wave height

#### 1) IN SITU DATA

SWH measurements at 25 buoy stations (Fig. 5) are used for model evaluation. The buoy data are obtained from the NOAA National Data Center (http://www.ndbc. noaa.gov/). The choice of locations for comparison is restricted to deep-water sites and to the stations having the longest data available (i.e., 42002, 46035) or providing larger area coverage even if the record is short (i.e., 32302, 52200, and 41041). Since the results presented in this study are from a free model run, no data assimilation techniques are used in either the atmospheric or the wave model simulations, the short-term weather events produced by this model can be very different from observations. Even though investigation of the stochastic properties of short-term weather events (i.e., 99th percentile SWH) is possible, point-to-point comparisons between our model results and observations are not possible on these short weather time scales. To eliminate the short-term weather events for the comparisons, we calculated the monthly mean SWH of both our model results and the buoy measurements, and then interpolated the model monthly mean SWH onto the NDBC buoy locations for comparison. The variability in monthly means is a mixture of forced responses to SST anomalies and interval variabilities. We have separated these buoy stations into regional groups and recorded the statistical comparison between model and NDBC data in Table 1. In general, the model mean SWHs are close to buoy observations with a bias less than  $\pm 10\%$ , except for some of the Bering Sea stations where the model significantly overestimate the SWH. This is because the Bering Sea is bounded by a chain of small islands, which separates the Bering Sea from the North Pacific Ocean. Several of these islands are not included in our simulation due to the relatively coarse model resolution. The omission of



FIG. 5. Map of NDBC locations used for model evaluation.

these islands reduced the wave sheltering effect in this region and thus resulted in higher waves.

1%–99% quantile-quantile (Q-Q) plot of combined NDBC buoy measurements by region with corresponding model data are presented in Fig. 6, which again demonstrates good agreement between model and buoy measurements except in the Bering Sea region. The northwest and southeast Pacific and northwest tropical Atlantic regions are excluded since the total number of buoy data is too small for Q-Q plots.

Table 1 also indicates high correlation between model SWH and buoy measurements. These correlations are mainly coming from the strong seasonal cycles in both model and observations; the correlations for deseasoned time series are much lower.

#### 2) SATELLITE MEASUREMENTS

The TOPEX/Poseidon along-track quality-checked deep-water altimeter measurements of SWH for the years 1998–2000 were obtained from the Jet Propulsion Laboratory of National Aeronautics and Space Administration (ftp://podaac.jpl.nasa.gov/allData/topex/L2/ tp\_ssha). The satellite measurements are performed about every second with a spacing of about 5.8 km. We form altimeter observations by grouping together the observations crossing a  $2^{\circ} \times 2^{\circ}$  region each month. The altimeter observations are taken as the mean of these grouped data points. To compare our model results with these constructed altimeter observations, we averaged our model monthly mean SWH onto corresponding  $2^{\circ} \times 2^{\circ}$ grids and generated 1%-99% Q-Q plots for these two datasets. These Q-Q plots show very good agreements between our model SWH and altimeter measurements for all three years (Figs. 7b,c,d).

Hemer et al. (2010b) performed a detailed analysis of altimeter-derived SWH in the Southern Hemisphere through a combination of several satellite measurements [Geosat, ERS-1, ERS-2, TOPEX/Poseidon, Jason-1, Envisat, and the Geosat Follow On-1 (GFO-1)] and compared them with ERA-40 SWH data. To evaluate the SWH trends in the ERA-40 dataset, they calculated the mean SWH difference of two 4-yr periods between 1993-96 and 1998-2001 for the altimeter-measured SWH (Fig. 8a), the ERA-40 SWH (Fig. 8c), and the corrected ERA-40 SWH (Fig. 8d). To evaluate our model SWH, we also calculated the SWH difference between the same two 4-yr periods (Fig. 8b). We found our model differences look more like the corrected ERA-40 and compare better with the altimeter-measured SWH than ERA-40. Our model was able to capture most of the positive and negative difference areas identified by both the observations and the corrected ERA-40 with similar magnitude. To find out the reason for the different behavior between our model and ERA-40, we calculated the 10-m wind speeds for the same period. The wind speed difference turns out to be very similar between ERA-40 (Fig. 8f) and our model (Fig. 8e). The major difference between them is that our model shows much larger increase of wind speed to the northeast of Australia, but this is unlikely to be connected to the bulk of the SWH differences. We suspect that it is the higher resolution of our wave model (both spatial and spectra) that results in this improvement and that this increased resolution is useful even in mean wave climatology studies. Tolman (1992) has pointed out that the influence of numerical errors on wave growth rates, response to turning winds, propagation of swells, and dynamical interaction between swells and wind sea is closely related to both the spatial and spectrum resolution

TABLE 1. Statistical comparisons of the model significant wave height time series vs NDBC buoy measurements. Here, RMSE is the root-mean-square error calculated as RMSE =  $\sqrt{\sum (Mod_{i,j} - Obs_{i,j})^2/mn}$ , where i = 1, n and n is the total number of measurements at the buoy stations.

NDBC	DBC Correlation Obs Mod		Model	RMSE	Month of obs data							
station	coef	coef mean (m) mean (n		(m)								
North Atlantic												
44011	0.72	2.05	2.05 2.01 0.44		144							
44004	0.62	2.00	1.99 0.48		156							
41001	0.60	1.98	1.91 0.45		92							
41002	0.66	1.87	1.78	107								
41048	0.75	1.93	1.84	0.34	24							
Gulf of Mexico and Caribbean												
42001	0.68	1.09	1.266 0.34		245							
42002	0.66	1.25	1.294	0.29	251							
42003	0.64	1.07	1.14	0.30	234							
42056	0.78	1.26	1.43	0.28	24							
Northwest tropical Atlantic												
41043	0.69	1.88	1.91	0.29	24							
41041	0.71	2.08	2.13	0.31	36							
Northeast Pacific												
46005	0.82	2.82	3.05	0.60	228							
46006	0.80	2.85	2.97	0.64	192							
46059	0.73	2.73	2.81	0.55	144							
46003	0.76	3.04	3.38	0.72	141							
46001	0.78	2.74	2.89	0.56	274							
46085	0.89	2.66	2.97 0.50		24							
Bering Sea												
46072	0.78	3.00	3.35	0.69	36							
46073	0.80	2.31	3.20 1.03		36							
46035	0.75	2.56	2.95 0.72		215							
46070	0.88	2.65	2.96	0.56	12							
Hawaii												
51001	0.79	2.43	2.52	0.36	167							
51003	0.74	2.23	2.41 0.34		144							
Northwest Pacific												
52200	0.59	1.57	2.29	0.85	36							
Southeast Pacific												
32302	0.67	2.18	2.45	0.38	72							

of the wave model. A significant part of the numerical errors can be eliminated by using a finer spatial resolution and extended frequency range. Furthermore, we suspect the coarse wave direction resolution in ERA-40 (twice as large as our model) is responsible for predicting the large decrease of SWH in the eastern Pacific. The Pacific Ocean is dominated by swells from the Southern Ocean. Accurately predicting the swell-propagating direction from the Southern Ocean is essential for correctly modeling the wave heights in this region. Also notice that both our model and ERA-40 overestimated the positive difference in the Southern Ocean (south of approximately 45°S). Ardhuin et al. (2011) have found that the distribution and variation of small icebergs appear to be very strongly associated with the anomalies in the SWH fields. Their preliminary parameterization of wave blocking by icebergs have shown significant reduction in wave model SWH errors in the region south of 45°S.

Our model also significantly overestimates the SWH differences at the northeast coast of Australia. This indicates that the tropical cyclones in this area may be too active in our model. Also, because of the resolution of our model, we missed the entire Santa Cruz Islands chain that located to the northeast of Australia. This will seriously reduce the wave-sheltering effect in this region and end up with much larger waves approaching the northeast coast of Australia than observed.

## 3) ERA-40

An example of comparison of the annual mean SWH for the period of 1981–2001 is shown in Fig. 9 and the comparison statistics of both annual and seasonal mean of SWH, mean wave direction, and mean wave period are recorded in Table 2. The ERA-40 SWH from January 1992 until May 1993 is corrupted because erroneous fast delivery product (FDP) *ERS-1* SWH measurements were assimilated into ERA-40 during this period (Caires et al. 2004); thus, it is not used in our model evaluation. Sea ice varies in time in both our coupled system and ERA-40. To avoid the complication in averaging and comparison due to the variation of ice, the area with ice variation (i.e., the Hudson Bay and the areas surrounding Antarctic) will not be compared.

We have also included model annual mean wavelength L and direction as arrow length and direction in Figs. 9a,b to illustrate the global wave propagation pattern. ERA-40 does not include information of mean wavelength, so we computed an approximate value by using mean wave period T in the dataset and applying the linear wave dispersion relation ( $f^2 = gk$ , where k = $2\pi/L$  and  $f = 2\pi/T$ ). We can see that our model annual mean SWH and wave propagation direction closely resembles that of the ERA-40. The high waves are found at the high latitudes with the highest waves in the Southern Ocean. This is because first, the surface westerlies in the Northern Hemisphere are much weaker compared to the Southern Hemisphere (Fig. 3); and second, the waves at the high latitude of the North Pacific and North Atlantic Ocean mainly propagate to the northeast and thus have relatively smaller fetch, while the wave propagation direction in the Southern Ocean are more aligned to the east and thus have an extended



FIG. 6. Q-Q plots (from 1%-99%) of model vs buoy significant wave height by region.



FIG. 7. Q-Q plots (from 1%–99%) of (a) ERA-40 (black dots) and corrected ERA-40 (gray circles) vs model SWH from 1981 to 2001; and model vs TOPEX/Poseidon SWH from January to December for (b) 1998, (c) 1999, and (d) 2000.

fetch around the globe (Figs. 9a,b). The combination of higher wind and extended fetch results in larger waves in the Southern Ocean.

The arrows in Fig. 9a indicate that the waves generated in the Southern Ocean can propagate northeast into the Pacific, Indian, and Atlantic Oceans as large swells. The directions of these swells gradually turn toward the north and propagate into the Northern Hemisphere. The waves generated in the high latitude of the Northern Hemisphere propagate southeast into the Pacific and Atlantic Oceans as large swells as well. Thus, even though the SWH decrease gradually from high latitudes toward the equator, large mean wavelength is found throughout the eastern part of Pacific and Indian Oceans (Fig. 9d). This indicates that although the SWH field closely follows the variation in the local wind fields, the mean wavelength field and thus the phase speed of the wave packets are not correlated with the local winds because of swell propagation. This effect has been nicely demonstrated by the inverse wave age plot generated by Hanley et al. (2010) using ERA-40.

As a result of the sheltering effect created by the southwestern coast of South America and the western coast of Africa, fewer swells propagate into the Atlantic Ocean. Thus, the swells in the Atlantic Ocean are much smaller and mostly confined in the South Atlantic. It is interesting to notice the well-confined tongue-shaped large mean wavelength region to the west of South America in the equatorial region, which coincide with the Niño-1+2 and Niño-3 region (black boxes in Fig. 9d) and where the east Pacific cold tongue exists. The interaction between the local winds and these long waves



FIG. 8. Mean SWH difference (m) between two 4-yr periods (1998–2001 and 1993–96) calculated using (a) TOPEX/Poseidon altimeter measurements, (b) GFDL coupled model, (c) ERA-40 data, and (d) C-ERA-40 data. Mean 10-m wind speed difference for the same period calculated using the (e) GFDL coupled model and (f) ERA-40 data. Both (a) and (b) are obtained from Fig. 8 in Hemer et al. (2010b).

could have an important effect on SST and the mixed layer depth in these regions.

Notice that our model SWH has much higher magnitude than ERA-40 especially in the Southern Ocean and high latitudes in the Northern Hemisphere. Also, there is a big difference in SWH in the western Pacific Ocean (Fig. 9c), where the tropical cyclones are most active. To illustrate this point, we also calculated the 99th percentile SWH from the 6-hourly data during the 21-yr period for both ERA-40 (Fig. 4b) and our model (Fig. 4d) results. These two SWH fields have very similar global structure and the correlation between them is 0.97. However, we can clearly see the strong tropical storm generated waves in the western Pacific with SWH above 7-8 m from our model results. This is consistent with the 10-m wind speed differences we discussed in the previous section. From Table 2, we can see that, for both the annual and seasonal means, our model results are highly correlated with ERA-40 data for all three variables. The SWH in our model is higher in magnitude than ERA-40 for both annual and seasonal means. Same as the 10-m wind speed, both the model and ERA-40 SWH have the highest spatial variability during the winter season in the Southern Hemisphere (JAS) when the westerlies reach maximum strength over the Southern Ocean. But, unlike the 10-m winds, the SWH field appears to have the lowest spatially variability during the months of JFM, the winter season in the Northern Hemisphere, which indicates that the SWH fields are nonlocal (not entirely dominated by the local wind field).

To evaluate the SWH differences between our model and ERA-40, we generated a Q-Q plot (1%–99%) of ERA-40/C-ERA-40 monthly mean SWH versus our model results at every ERA-40/C-ERA-40 grid point for all 12 months from January to December (Fig. 7a). The slope with ERA-40 is very similar to the Q-Q plot



FIG. 9. (a) Model and (b) ERA-40 annual mean significant wave height (color, m), mean wavelength (arrow length), and mean wave direction (arrow direction); (c) significant wave height difference (color, m); and (d) model annual mean wavelength (color, m) for the period 1981–2001.

(1%–99%) of ERA-40 SWH versus TOPEX/Poseidon data given by Caires et al. (2004). We can see that our model SWH show fairly good agreements with the corrected ERA-40 SWH below 4.5 m, and overestimate the mean SWH above it. One of the reason for this overestimation could be that our model has a high bias for large waves. The other reason could also be that since tropical cyclones are not resolved by the ERA-40 system because of its low resolution, the corrected ERA-40 data for the regions of the tropical storms may also have a low bias (Caires and Sterl 2005).

Table 2 also indicates that both the model mean wave period and mean wave direction are highly correlated with ERA-40 for both annual and seasonal means. It is interesting to notice that even though our model mean wave period has more spatial contrast than ERA-40, the average mean wave period is smaller in our model compared with ERA-40. This may be because both the spatial and spectrum resolution of our model (0.5° spatial resolution and 40 frequencies in spectrum resolution) is higher than ERA-40 (1.5° spatial resolution and 25 frequencies in wave spectrum resolution). The higher resolutions in our model may shift the average period to shorter waves. Even though the mean wave direction variability are similar between our model and ERA-40 for both annual and seasonal comparisons, the mean direction in our model is typically rotated slightly counterclockwise compared to ERA-40. This is more likely because our model solves the wave spectrum in 24 directions, while ERA-40 solves it in 12 directions.

## 4. Wave climatology variability

To evaluate the wind and wave field produced by our model on long time scales, we will calculate the regression of sea level pressure (SLP), 10-m wind, SWH, mean wave direction, and mean wave period against the North Atlantic Oscillation index in the boreal winter season (JFM) and against the Southern Oscillation index in both the boreal winter (JFM) and summer (JAS) season. Although the ERA-40 has low bias for high winds and waves, studies (e.g., Marshall 2003; Hemer et al. 2010b) have demonstrated its suitability for climate variability researches. Thus, we will evaluate our

TABLE 2. Statistical comparisons between the model results and ERA-40. Here WSP stands for 10-m wind speed, SWH stands for significant wave height, Dir stands for mean wave direction, *T* stands for mean wave period, and RMSE is the root-mean-square error calculated as  $\text{RMSE} = \sqrt{\sum (\text{Mod}_{i,j} - \text{Obs}_{i,j})^2 / mn}$ , where, i = 1, n, j = 1, m. The correlation coefficients given in this table are for spatial pattern correlations.

		Model		ERA-40			
		Mean	Std dev	Mean	Std dev	Correlation coeff	RMSE
Annual	WSP (m $s^{-1}$ )	6.00	2.62	6.01	2.29	0.90	1.14
	SWH (m)	2.47	1.24	2.18	0.96	0.96	0.51
	Dir (°)	170.4	69.5	180.5	69.7	0.92	29.5
	$T(\mathbf{s})$	7.96	2.55	8.49	2.13	0.87	1.39
JFM	WSP (m $s^{-1}$ )	5.93	2.64	5.94	2.35	0.92	1.02
	SWH (m)	2.62	1.07	2.38	0.85	0.94	0.47
	Dir (°)	177.8	73.2	196.8	71.4	0.90	38.2
	$T(\mathbf{s})$	8.17	2.15	8.94	1.71	0.81	1.48
АМЈ	WSP (m $s^{-1}$ )	5.97	2.68	6.00	2.40	0.90	1.22
	SWH (m)	2.56	1.31	2.25	1.01	0.96	0.54
	Dir (°)	178.0	69.4	187.8	68.2	0.93	28.0
	$T(\mathbf{s})$	8.34	2.35	8.95	1.89	0.84	1.42
JAS	WSP (m $s^{-1}$ )	6.15	2.86	6.12	2.52	0.89	1.33
	SWH (m)	2.63	1.43	2.21	1.06	0.96	0.65
	Dir (°)	177.4	64.8	183.2	64.1	0.90	29.6
	T(s)	8.16	2.50	8.49	2.09	0.88	1.24
OND	WSP (m $s^{-1}$ )	5.94	2.67	5.98	2.31	0.91	1.10
	SWH (m)	2.46	1.20	2.17	0.94	0.96	0.50
	Dir (°)	175.0	69.8	182.2	70.3	0.92	29.2
	<i>T</i> (s)	8.21	2.46	8.55	2.02	0.87	1.27

model wind and wave climate variability against ERA-40 in this section.

#### a. North Atlantic Oscillation

The annular modes in the North Hemisphere, also known as the North Atlantic Oscillation (NAO) is the most important pattern of hemispheric-scale climate variability in the mid- and high latitudes of North Atlantic where it modulates the strength and direction of westerly winds and storm tracks, and is strongly coupled with surface wind variability.

The NAO index for both ERA-40 (Fig. 10a) and the coupled model (Fig. 10b) are derived using the first principal component of a principal component analysis (PCA). SLPs in the Atlantic region (20°–90°N, 100°W–  $0^{\circ}$ ) were used for the PCA analysis. For both ERA-40 and our model results, the regression of SLP anomaly and 10-m wind vector with the NAO index demonstrated the NAO characteristics with a dipole SLP pattern between the Icelandic low and the Azores high (Figs. 11a,b). Positive NAO is associated with a negative SLP anomaly at high latitudes and a positive SLP anomaly in the midlatitudes. The positive SLP anomaly in the midlatitudes is stronger in the model. The corresponding 10-m wind vectors are roughly in geostrophic balance around these two SLP anomalies, strengthening the westerlies at high latitudes and weakening it at midlatitudes.

The regression coefficients of the SWH anomaly with the NAO index (Figs. 11c,d) also show a well-organized dipole pattern in the North Atlantic corresponding to the atmospheric responses (Figs. 11a,b). Positive SWH anomalies are found at high latitudes and the maximum increase coincide with the strongest increase of westerlies, while negative SWH anomalies are found at lowto midlatitudes and the maximum decrease coincides with the strongest decrease of westerlies. Corresponding to the atmospheric responses, the model-positive SWH anomalies are restrained to higher latitudes and relatively weaker, while the negative SWH anomalies at the low to midlatitudes are much stronger. The strongest positive SWH response is found at the west coast of the United Kingdom for both our model results and ERA-40 due to the increase of westerlies in this region. The strongest negative response is found along the eastern coast of North America, which is caused by the weaker waves propagating into this region from the Atlantic Ocean.

The mean wave direction anomaly response to the NAO index are shown as black arrows in Figs. 11c,d. We can see that, during positive NAO, the waves are directed more toward the midlatitudes from both the high latitudes and the subtropical region. Also, the waves at the high latitudes are directed more to the east toward Europe and the waves at the midlatitudes are directed more to the west toward the U.S. coast in the subtropics.



FIG. 10. Winter (JFM) NAO index time series calculated for (a) ERA-40 and (b) the coupled model, winter (JFM) SOI time series calculated for (c) ERA-40 and (d) the coupled model, and summer (JAS) SOI time series calculated for (e) ERA-40 and (f) the coupled model.

Unlike the SWH response, the strongest direction response is found to the east of Canada.

The positive mean wave period response to NAO is observed in the northeast sector of the North Atlantic with the maximums found along the western coast of Europe, while the negative response is found in the southern and western North Atlantic (Figs. 11e,f). This is because the mean wave period is dominated by swells.



FIG. 11. Regression coefficient against the NAO index during the JFM season for SLP pressure (color) and 10-m wind vector from (a) ERA-40 and (b) the model results, SWH (color) and Dir (vector) from (c) ERA-40 and (d) the model results, and mean wave period (color) from (c) ERA-40 and (d) the model results.

During positive NAO, the larger waves generated at high latitudes propagate toward the northeast and southeast as longer swells and increase the mean wave period in these regions, while the smaller waves generated at midlatitudes propagate northwest and southwest as shorter swells and decrease the mean wave period in the western North Atlantic.

# b. Southern Oscillation index

El Niño–Southern Oscillation (ENSO) is the most important pattern of large-scale climate variability in the tropics. The SOI index for both ERA-40 (Figs. 10c,e) and the coupled model (Figs. 10d,f) are derived using the standardized anomaly of the mean sea level pressure difference between Tahiti and Darwin.

## 1) BOREAL WINTER (JFM)

Figures 12a,b exhibits the regression coefficients of the SLP anomaly and 10-m wind vector with the SOI index for ERA-40 and the model results during the boreal winter (JFM). For both models, the positive SOI index is corresponding to positive SLP anomalies in the North Pacific centered near the Aleutian Island where the Aleutian low is. The positive SLP anomaly is stronger in the model and the center of this anomaly is shifted  $\sim 5^{\circ}$  eastward in the model compared to ERA-40. The corresponding 10-m wind vectors are roughly in geostrophic balance with the SLP anomaly, weakening the westerlies at midlatitudes, and steering it northward over the Bering Sea and southward along the western coast of North America.



FIG. 12. As in Fig. 11, but for the regression coefficient against the SOI during the JFM season.

The regression coefficients of the SWH (color) and wave direction (vectors) anomaly with the JFM SOI are given in Figs. 12c,d. We can see that, during positive SOI, the waves are directed more toward the north and northeast in most part of the North Pacific except the subtropical western Pacific region, where waves are directed more toward the southeast. Mild positive SWH anomalies are found at high latitudes in and around the Bering Sea. When the westerlies are steered more toward the north in this region (Figs. 12a,b), the propagation direction of the strong waves generated by the winter storms in this region are directed more toward the north and resulted in the increase of SWH in the Bering Sea region. Consequently, the amount of strong swells propagating southward is reduced. The reduction of swells from the north together with the weakening of the westerlies at the midlatitudes have resulted in a strong negative SWH anomalies in the central North Pacific between 30° and 45°N. Like the NAO response,

even though the atmospheric responses to the SOI variation are stronger over the deep ocean, the wave field responses appear to be stronger in the coastal areas. The accumulation of large waves at the northern Bering Sea has resulted in a very strong increase of SWH and similarly a strong increase of the SWH along the east coast of Asia.

Unlike the SWH, the negative mean wave period response to SOI is observed in the entire eastern North Pacific with the minimum found around 15°N. In the North Pacific, the large waves generated at the high latitudes propagate both northeast and southeast as swells. During positive SOI, there are more waves propagating northeast and less propagating southeast. Since the mean wave period at the midlatitudes and subtropical regions are dominated by these large swells (Chen et al. 2002), the reduction of swells from the high latitudes has resulted in the reduction of mean wave period in these regions.



FIG. 13. As in Fig. 11, but for the regression coefficient against the SOI during the JAS season.

#### 2) BOREAL SUMMER (JAS)

For both ERA-40 and our model results, the regression of the SLP anomaly with the JAS SOI index shows a dipole SLP pattern over the South Pacific (Figs. 13a,b). Positive SOI is associated with a negative SLP anomaly at high latitudes and a positive SLP anomaly at the midlatitudes. This pattern is shifted 10°–15° westward in our model compared to ERA-40, and the positive SLP anomaly in the midlatitudes is weaker in our model. The corresponding 10-m wind vectors are roughly in geostrophic balance around these two SLP anomalies, strengthening the westerlies at the high latitudes.

The regression coefficients of the SWH anomaly with the JAS SOI index (Figs. 13c,d) show a well-organized tripole pattern in the South Pacific corresponding to the atmospheric responses (Figs. 13a,b). The stronger positive SWH anomalies are found at high latitudes and the maximum increase coincide with the strongest increase of westerlies, while the weaker positive SWH anomalies are found in the subtropical region for both our model and ERA-40. Mild negative SWH anomalies are observed at the midlatitudes. Same as the atmospheric response, the SWH responses in our model are shifted  $\sim 10^{\circ}$ -15° to the west compared to ERA-40. As we can see from Figs. 9a,b, swells from the Southern Ocean propagate northeast into the South Pacific. Since the strong positive SWH anomalies from ERA-40 are found closer to the western coast of South America, swells from this region cannot propagate far into the South Pacific, while the SWH anomalies from our model are locate 10°-15° away from the South America coast and thus can easily propagate all the way through the South Pacific. This has resulted in the very different wave direction and period responses (Figs. 13e,f) between our model results and ERA-40, since the waves in the South Ocean dominate the wave responses in the South Pacific (Hemer et al. 2010b).

Also notice that the strongest positive and negative SWH and wave period responses are found along the Antarctic coast for both models.

#### 5. Summary

This study describes and evaluates a 29-yr (1981–2009) global surface gravity wave simulation generated by a new coupled atmosphere-wave model, with no atmospheric data assimilation, but running over observed SSTs and sea ice. The atmospheric model used in the coupled system is the NOAA/GFDL High-Resolution Atmosphere Model (HiRAM), and the wave model used here is WAVEWATCH III, the operational wave model developed and used at NOAA/NCEP. Prescribed SSTs and sea ice from the Hadley Centre Sea Ice and SST version 1.1 model (HadISST 1.1; Rayner et al. 2003) is used as the lower boundary condition for the atmospheric model. A new momentum roughness  $(z_0)$  parameterization as a function of both the wind speed and the sea state is also developed based on recent field observations. The new roughness is similar to Beljaars (1995) when the wind speed is less than 20 m s<sup>-1</sup>, but become significantly lower as the wind speed increases.

Extensive evaluation of monthly mean SWH against in situ buoys, satellite altimeter measurements, and ERA-40 showed very good agreements in terms of magnitude, spatial distribution, and scatter. The comparisons with satellite altimeter measurements indicate that the low bias in ERA-40 has been improved in our model simulations, partly due to the presence of tropical storms in our model. The model SWH, mean wave direction, and mean wave period have a strong response to the NAO in the North Atlantic during the boreal wintertime (JFM) and to the SOI in the Pacific Ocean during the boreal winter (JFM) and summer (JAS). But unlike the atmospheric subpolar responses, which are larger in the center of the ocean basins, the SWH responses to NAO and SOI are larger in the coastal regions.

This study has demonstrated that the coupled atmospherewave model that we have developed at GFDL does a reasonably good job in simulating the wave climatology for the past. Thus, we have gained some confidence in applying this model in the attribution of wave and wave climate projections for the future. In a companion paper, we will investigate the wave climate change to the SST/ sea ice anomalies in the late twenty-first century. Of course, the model behavior may differ considerably under climate forcing. Demonstrated high model skill for present-day simulations does not necessarily guarantee high model skill in climate projections for the future.

This new coupled system has made it possible for the atmosphere model to use a more physically based momentum roughness length based on sea states, and opened a new door for research on sea spray parameterization based on wave breaking, which will lead to improvements of the latent heat flux and albedo parameterizations in the atmosphere models. The success of this coupled system also gets us one step closer toward building a fully coupled atmosphere–ocean–wave model aimed to better understanding the fluxes of the atmospheric and oceanic boundary layers in the climate models.

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#### APPENDIX

#### Momentum Roughness Length Parameterization

The default WWIII drag coefficient parameterization is given in details in Tolman and Chalikov (1996). We are only going to go through the main points of their approach. This approach assumes the mean wind profile is close to logarithmic:

$$u_z = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_0}\right),\tag{A1}$$

where  $u_z$  is the wind speed at the reference height  $z, u_*$  is the friction velocity,  $\kappa = 0.4$  is the Von Kármán constant, and  $z_0$  is the momentum roughness length.

Equation (A1) is then rewritten in terms of the drag coefficient  $C_d$  following Chalikov (1995):

$$C_d = 10^{-3} \left( 0.021 + \frac{10.4}{R^{1.23} + 1.85} \right),$$
 (A2)

where, 
$$R = \ln\left(\frac{zg}{\chi\sqrt{\alpha}u_z}\right)$$
, (A3)

where g is the gravitational acceleration,  $\chi$  is a constant value of 0.2, and  $\alpha$  is the high-frequency energy level estimated parametrically from the Joint North Sea Wave Project (JONSWAP) data (Janssen 1989):



FIG. A1. Drag coefficient  $C_d$  comparison between our model parameterization (red dots) and different observations. The black circles and dots in the background are field measurements presented in Fig. 4 in Drennan et al. (2003). The green circles are field measurements from Drennan et al. (1996). The blue open squares are CBLAST field measurements from French et al. (2007). The four black symbols (see key in figure) shows the  $C_d$  estimated by Powell et al. (2003) based on different depth of surface layers, and the vertical bars indicate the 95% confidence limits on the estimates. The gray dashed line shows the Large and Pond (1981) wind speed–dependent bulk formula.

$$\alpha = 0.57 \left(\frac{c_p}{u_*}\right)^{-1.5},\tag{A4}$$

in which  $c_p$  is the phase speed at the peak frequency. Equations (A2)–(A4) yield a higher drag coefficient for higher wind speeds and younger waves, and therefore significantly overestimate the wind stress at high wind speeds (Moon et al. 2004). Moon et al. (2004) have shown that at a given wind speed, the relationship between the Charnock coefficient  $z_{ch}$  and the input wave age  $c_{pi}/u_*$ (the phase speed of dominant wind-forced waves calculated within the WWIII divided by the wind friction velocity) can be fit by the following form:

$$z_{\rm ch} = \frac{z_0 g}{u_*^2} = a \left(\frac{c_{pi}}{u_*}\right)^b,$$
 (A5)

where, *a* and *b* are fitting constants at nine different wind speeds ranging from 10 to 50 m s<sup>-1</sup> at a 5 m s<sup>-1</sup> interval. To use such a type of parameterization in numerical models, we have derived new empirical functions for the constants *a* and *b*, which can change continuously with wind speed by fitting the field data available in Figure A1:

$$a = \frac{0.023}{1.0568^{U_{10}}}, \quad b = 0.012U_{10},$$
 (A6)

where  $U_{10}$  is the 10-m wind speed in m s<sup>-1</sup>. We solve (A5) and (A6) together with (A1) at z = 10 m to obtain  $z_0$ .

Idealized experiments as described in Fan et al. (2010) are used for this empirical study. The new parameterizations are tested using the stand-alone WWIII under hurricane conditions following Fan et al. (2009b) and proved to be able to produce excellent wave fields compare to observations (not shown), very similar to the more complex scheme of Fan et al. (2009b).

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