ABSTRACT

GONG, YANLIN. Numerical Investigation of Coastal Circulation Dynamics near Cape Hatteras, North Carolina in January, 2005. (Under the direction of Dr. Ruoying He.)

A realistic regional ocean model is used to hindcast and diagnose coastal circulation variability near Cape Hatteras (CH), North Carolina in January, 2005. Strong extratropical winter storms passed through the area during the 2nd half of this month (January 15 ~ 31), leading to significantly different circulation conditions compared to those during the 1st half of this month (January 1 ~ 14). Model results are firstly validated against sea level, temperature, salinity and velocity observations. Analyses of along-shelf and cross-shelf transport, momentum, vorticity, and energy balances are further performed to investigate the circulation dynamics near CH. Our results show that during the strong winter storm period, both along-shelf (alongshore to the southward) and cross-shelf transport (seaward) increase significantly, mainly due to increases of geostrophic velocity associated with coastal sea level setup. In terms of momentum balance, the wind stress is mainly balanced by the bottom friction. Similarly, in the vorticity balance, the bottom stress curl offsets much of wind stress curl. The JEBAR (Joint Effect of Baroclinicity and Relief) effect is the most significant contributor of APV (Advection of Potential Vorticity). During the first half of month, the dominant kinetic energy balance on the shelf is between the time rate of kinetic energy change and the pressure work (PW) and dissipation, whereas during the stormy 2nd half of month, the main shelf energy balance is achieved between wind stress work and dissipation.

Numerical Investigation of Coastal Circulation Dynamics near
Cape Hatteras, North Carolina in January, 2005

by
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Chapter 1: Introduction

1. Oceanographic Setting near Cape Hatteras (CH)

Located on the east coast of North Carolina, Cape Hatteras (CH) extends offshore, separating two oceanographically different regions: the Middle Atlantic Bight (MAB) and the South Atlantic Bight (SAB) (**Figure 1**). The southwestward mean MAB shelf flow and northeastward mean SAB shelf flow converge off CH. In the mean time, the Gulf Stream travels through the area, turning northeastward seaward of CH. While MAB shelf waters make occasional excursions into the SAB, most of them are entrained into the Gulf Stream (*Gawarkiewicz et al., 2009*). Hydrographic properties and circulation dynamics near CH are highly complex due to the presence of different water masses. Based on the temperature and salinity observations in August 2004, *Savidge and Austin* (2007) identified four kinds of water masses in the area: cold fresh MAB water, warm salty SAB water, fresh Chesapeake Bay Plume water, and highly saline Gulf Stream water. Consequently, several ocean fronts co-exist near CH, including (1) Hatteras Front — the water mass boundary between the cold, fresh MAB shelf water (including Chesapeake Bay river plumes) and warm salty SAB shelf water; (2) MAB shelfbreak front — the water mass boundary between the cold fresh MAB shelf water and warm salty slope water; (3) Gulf Stream front — the water mass boundary between the warm salty Gulf Stream water and cold fresh shelf water.

Distinctive regional temperature and salinity characteristics are seen in other seasons as well (e.g., *Castelao et al., 2010*). In winter, the water column is largely well mixed due to

surface cooling and strong wind mixing. In summer, as a result of strong surface heating, the water column is thermally stratified, with a sharp thermocline developed at about 20-m depth (*Beardsley et al., 1985*). Below the thermocline is a band of cold water often referred as the “cold pool” (e.g., *Houghton et al., 1982*). Salinity increases from about 32 near the coast to about 34 at the shelfbreak (*Mountain, 2003*). In contrast to water temperature, seasonal variations of cross-shelf salinity gradients are relatively smaller. The exceptions are found at the inner shelf and near major estuaries, where the salinity gradients are enhanced due to increased river discharge during the spring (*Chant et al., 2008*). Shelf velocity observations from drifters (*Bumpus, 1973*) and moored current meter arrays (*Beardsley et al., 1985*) shows the southward depth-averaged mean flow is on the order of 5 cm/s. Lentz (2008) recently showed along-shelf current increased with water depth, changing from 3 cm/s at the 15 m isobath to 10 cm/s at the 100 m isobath. The along-isobath subtidal flow on the New England shelf (the northern part of MAB) is primarily geostrophic and barotropic, and correlated with large-scale wind stress fluctuations, especially under the strong wind in winter (*Shearman and Lentz, 2003*). Depth-averaged ageostrophic transport is consistent with Ekman transport induced by wind and bottom stress (*Shearman and Lentz, 2003*). Current oscillation amplitudes are higher on the slope than on the shelf in the synoptic band (periods between 8 and 4.8 days), but the phase is nearly constant across the shelf and slope (*Shaw et al., 1994*). South of CH, the SAB shelf circulation system can be dynamically separated into the inner shelf (~0- 20 m), middle shelf (~20- 40 m), outer shelf (~40- 70 m) and shelf slope (~70- 200 m) (*Atkinson et al., 1983*). The inner shelf is forced by river runoff and local winds, the middle shelf by local winds and Gulf Stream intrusions, and the outer shelf and

shelf slope by Gulf Stream and its frontal activities (*Lee et al., 1991*). Topographic effects such as isobaths divergence in the Straits of Florida, the Charleston Bump, and isobaths convergence off the CH all lead to complex circulation gyre and variability in the SAB.

One unique shelf circulation feature in our study area is the shelfbreak jet/front, which separates warm, saline slope water from cool, fresh shelf water. The MAB shelfbreak front and shelfbreak jet have been described in numerous studies utilizing intensive hydrographic surveys (*e.g., Beardsley and Flagg, 1976*), long-term mooring arrays such as Nantucket Shoals Flux Experiment (NSFE) (*Beardsley et al., 1985*) and the Shelf Edge Exchange Processes experiments (SEEP I and SEEP II) (*Aikman et al., 1988; Houghton et al., 1994*), climatology field (*Linder and Gawarkiewicz, 1998*) and numerical models (*Chen and He, 2010*). The temperature contrast across the front varies seasonally between 2 and 6 degree. The salinity contrast is relatively small, ranging by 1.5-2 psu. In the winter, the shelf and slope regions are separated by the density front extending to the surface. In the summer, isopycnal surfaces are relatively flat, allowing for diffusive exchange of shelf and slope water (*Houghton et al., 1994*). Associated with the cross-shelf density gradients is a strong (20-30 cm/s) baroclinic jet (*Linder and Gawarkiewicz, 1998*). Two major EOF modes of shelfbreak current are identified from a recent shelfbreak circulation modeling study (*Chen and He, 2010*). The first EOF mode accounts for 61% variance, confirming that the shelfbreak jet is persistent year-round circulation feature. The second mode accounts from 13% variance, representing baroclinic eddy passages across the shelfbreak. Indeed, an important process of water exchange across the shelfbreak front is through eddies formed near the shelfbreak (*Garvine et al., 1988; Flagg et al., 1998*). Shelfbreak eddies are composed of a mix of water

masses including MAB shelf and slope water and the Gulf Stream water (*Churchill and Gawarkiewicz, 2009*).

Near CH, water is often found along the Gulf Stream's northern edge which is considerably cooler and less saline than adjacent Gulf Stream water. To explain this phenomenon, *Ford (1952)* ruled out the possibility of the upwelling based on the fact that the salinity was lower than deeper water and argued that this cold, fresh water came from MAB shelf water. A detailed examination of shelf water export near CH was reported by *Fisher (1972)*. A well defined entrainment was observed with bands of surface water 10-15 km wide and 40 m thick from the continental shelf to the northern edge of the Gulf Stream. *Kupferman and Garfield (1977)* observed both surface and subsurface entrained shelf water, located above and below the thermocline north of CH. The estimated entrained shelf water transport was about 40, 000 m³/s (for fresh water) or 9×10^5 m³/s (for 34 psu shelf water). *Lillibridge et al. (1990)* found that the entrained surface shelf water was also having a maximum in chlorophyll fluorescence and dissolved oxygen and contained a distinct diatom assemblage of nearshore species. This study also concluded that double-diffusive processes were more important than shear flow instability in governing cross-isopycnal mixing. *Wood et al. (1996)* pointed out that the ultimate fate of some of organic matter produced in the cold pool of the MAB was entrainment into the Gulf Stream at CH. *Churchill and Berger (1998)* examined the processes of MAB shelf water entrainment into the Gulf Stream and identified two export areas. The shelfwater export in the north zone (35.4° N-36.1° N) is due to seaward movement of shelf water into Gulf Stream meanders or water discharged from the Gulf Stream. The shelfwater export in the south zone (35.2° N-35.4° N) is due to seaward flow of

shelf water in a strong current (5~10 cm/s) at the edge of Hatteras Front. *Gawarkiewicz and Linder* (2005) examined the statistics of drifter trajectories crossing the 1000 m isobath and found that there were two primary modes of shelfbreak front/slope water entrained into the Gulf Stream: a slow passage across the slope region to the Gulf Stream well north of CH (20-100 km) with a large radius of curvature, and a rapid passage occurring in close proximity (within 20 km) of CH with a small radius of curvature.

The Gulf Stream may exert a tremendous influence on the dynamics of the outer shelf and slope of Cape Hatteras. Numerous investigations (*Churchill et al., 1991; Gawarkiewicz et al., 1992*) have found Gulf Stream water over the upper slope and outer shelf well north of Cape Hatteras, as far as 37°N. *Kumar et al.* (2006) found an inverse relationship between Gulf Stream position relative to the shelfbreak north of Cape Hatteras and events of shelf water “overrunning” into the slope sea, suggesting that the approach of the Gulf Stream to the shelfbreak has a “blocking effect” on shelf circulation. Lateral movement of the Gulf Stream front also appears to influence the strength of the flow over the shelfbreak and upper slope, although the manner of this influence is still open to question. From analysis of current meter data from the southern MAB, *Bane et al.* (1988) found that the southwestward flow over the upper slope tends to intensify as the Gulf Stream approaches the shelf edge. This is in conflict with the recent work of *Bohm et al.* (2006) who used satellite-derived sea surface temperature (SST) and sea surface elevation data to relate Gulf Stream position to the strength of the current over the southern MAB slope. Their analysis indicated the southwestward flow over the slope is accelerated when the Gulf Stream is relatively far from the shelfbreak, and is decelerated when the Gulf Stream is close to the shelfbreak.

2. Meteorological Setting near CH

The meteorological forcing near CH shows clear seasonality as a result of interactions between the Icelandic Low and Subtropical High. In winter, the Icelandic Low dominates, providing a strong counterclockwise gyre circulation in the middle/high latitudes of the North Atlantic basin. CH area is therefore under the impact of strong south-southeastward wind. In contrast, the Subtropical High dominates in the summer, producing anticyclone circulation over the North Atlantic basin. Prevailing winds near CH thus shift to north-northeastward. Both spring and fall are transition periods for such seasonal wind changes (*Weber and Blanton, 1980*).

On the synoptic time scale, the CH region is susceptible to influence of strong winter extratropical storms and summer tropical cyclones. In the summer, the warmer tropical waters in the southern North Atlantic Ocean are favorable to the formation and development of tropical cyclone systems. Some of them (such as Hurricane Isabel) move cross the CH region, causing devastating damages to property and life (*Bell and Montgomery, 2008*). In the winter, the relatively cold continent is bounded by relatively warmer shelf/slope water and by consistently warmer Gulf Stream water to the east. The amount of heat and moisture available to the atmosphere from the open sea surface are generally large due to the presence of warm waters. This prevailing meteorological condition is known to be a favorable setting for the development of the extratropical cyclones (*Zishka and Smith, 1980*). Major winter storms, characterized by “clipping” ice, heavy snow, and gale-force winds, often batter the U.S. east coast from the Carolinas northward. Extensive observational studies, such as the

Genesis of Atlantic Low Experiment (GALE) (*Dirks et al, 1988*), have been conducted to study the marine boundary layer dynamics and air-sea interactions associated with extratropical cyclones formation in this region. It is illustrated that both sensible and latent heat fluxes increase dramatically as the colder, drier air move out over the comparatively warm shelf waters, and marine boundary layer height increases steadily in the offshore direction in response to increasing convection (*Wayland and Raman, 1989*). Two-dimensional (*Chao, 1992; Xue et al., 2000*) and three-dimensional (*Li et al, 2002; Nelson and He, 2011*) coupled atmosphere-ocean models were used to understand the development of the atmospheric mesoscale front and local winds due to differential fluxes over the land, the cold shelf water and the warm Gulf Stream, and how the mesoscale front feed back to the ocean and modify the upper ocean temperature and current fields. Model results show that the heat flux in the coupled experiment is about 10% less than the total surface heat flux in the experiment with fixed SST, and the response of the upper-ocean velocity field to local winds is on the order of 20 cm/s, suggesting significant atmospheric feedback to the ocean through the heat-flux-enhanced surface winds.

3. Objectives and Structures of this Thesis Research

Indeed, winter storms and the passage of atmospheric frontal systems induce strong physical responses in the CH area, including significant changes in sea level, across-shelf and along-shelf currents, stratification and turbulent mixing. These responses to weather forcing often have significant socio-economic effects through storm surges, rip currents, beach erosion, nutrient transport and related primary productivity. Detailed understanding of the

coastal circulation near the CH and better quantification of its variations under the influence of the winter storm require high-resolution space and time continuous realizations of ocean state variables, from which quantitative dynamics can be gleaned.

In this study, January 2005 is chosen as a case study period. Strong extratropical winter storms passed through CH area during the 2nd half of this month, leading to significantly different circulation conditions compared to those during the 1st half of this month. CH coastal circulation response to different atmospheric forcing conditions in January, 2005 and the corresponding ocean dynamics (i.e., momentum, vorticity and energy) are simulated and analyzed with the regional coastal circulation model. Available in-situ observations provide valuable ground-truth to validate the numerical model solutions. The structure of this thesis is as follows: chapter 2 describes the ocean model configurations and model-observation comparisons. Detailed analyses on ocean transport and circulation dynamics are given in chapter 3, followed by discussions and conclusions in chapter 4.

Chapter 2: Ocean Model Hindcast

1. Model Configurations

A regional ocean circulation model is implemented for the MAB and SAB (hereafter MABSAB), covering the area between 81.89°W and 69.80°W, 28.41°N and 41.84°N (**Figure 1**). The model is based on the Regional Ocean Modeling System (ROMS) (*Shchepetkin and McWilliams, 2005*), a free-surface, terrain-following, primitive equations ocean model in widespread use for estuarine, coastal and regional ocean-wide applications.

ROMS solves Navier Stokes equations using the Boussinesq and hydrostatic approximation. Then horizontal momentum equations are:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - fv = -\frac{1}{\rho_0} \frac{\partial p}{\partial x} + \frac{\partial}{\partial z} (K_m \frac{\partial u}{\partial z}) + D_u \quad (1)$$

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + fu = -\frac{1}{\rho_0} \frac{\partial p}{\partial y} + \frac{\partial}{\partial z} (K_m \frac{\partial v}{\partial z}) + D_v$$

Surface and bottom boundary conditions are defined as following:

At surface $z = \zeta(x, y, t)$

$$\rho K_m \frac{\partial u}{\partial z} = \tau_s^x(x, y, t) \quad \rho K_m \frac{\partial v}{\partial z} = \tau_s^y(x, y, t)$$

At bottom $z = -h(x, y)$

$$\rho K_m \frac{\partial u}{\partial z} = \tau_b^x(x, y, t) \quad \rho K_m \frac{\partial v}{\partial z} = \tau_b^y(x, y, t) \quad (2)$$

Where D_u and D_v are horizontal mixing, τ_s^x and τ_s^y are surface wind stress, τ_b^x and τ_b^y are bottom stress, K_m is the vertical viscosity coefficient, $h(x, y)$ is bottom depth and $\zeta(x, y, t)$ is the surface elevation. The horizontal resolution of this model is 2 km. Model bathymetry is

interpolated from National Geophysical Data Center (NGDC) 2-Minute Gridded Global Relief Data, with the minimum water depth of 10 m and maximum water depth of 5645 m in the model domain. Vertically, there are 36 terrain-following layers spaced to better resolve surface and bottom boundary layers.

The vertical coordinate transformations are:

$$z(x, y, \sigma, t) = \zeta(x, y, t) + [\zeta(x, y, t) + h(x, y)] \left[\frac{h_c \sigma + h(x, y) C(\sigma)}{h_c + h(x, y)} \right] \quad (3)$$

$$C(\sigma) = (1 - \theta_b) \frac{\sinh(\theta_s \sigma)}{\sinh \theta_s} + \theta_b \left[\frac{\tanh[\theta_s (\sigma + \frac{1}{2})]}{2 \tanh(\frac{1}{2} \theta_s)} - \frac{1}{2} \right]$$

where σ is a fractional vertical stretching coordinate ranging from -1 to 0, θ_s and θ_b are the surface and bottom control parameters, and h_c is a positive thickness controlling the stretching. In our model, the vertical s-coordinate model parameters are: $\theta_s=5$, $\theta_b=0.4$ and $h_c=10$. Momentum advections are solved using a 3rd-order upstream bias scheme for 3D velocity, and a 4th-order centered scheme for 2D transport. Tracer (temperature and salinity) advections are solved with a 3rd-order upstream scheme in the horizontal direction, and a 4th-order centered scheme in vertical direction. The baroclinic time-step size is 180 seconds and the barotropic time-step size is 4.5 seconds. The horizontal mixing for both the momentum and tracer utilizes the harmonic formulation with 100 m²/s (20 m²/s) as the momentum (tracer) mixing coefficient. Turbulent mixings for both momentum and tracers are computed using Mellor/Yamada Level-2.5 closure scheme (*Mellor and Yamada, 1982*).

We utilize surface atmospheric conditions from North America Regional Reanalysis (NARR) provided by National Centers for Environmental Prediction (NCEP). The spatial and temporal resolutions of NARR are 35-km and 3 hourly, respectively. Air-sea fluxes of momentum and buoyancy are computed by applying the standard bulk formulae (Fairall et al., 2003) to NARR marine boundary-layer winds, air temperature, relative humidity, air pressure, along with ROMS generated surface currents. To further constrain the net surface heat flux, we implement a thermal relaxation term following *He and Weisberg* (2002), such that

$$K_H \frac{\partial T}{\partial Z} = \frac{Q}{\rho C_p} + c(T_{obs} - T_{mod}) \quad (4)$$

where $c = 0.125 \text{ day}^{-1}$, and T_{obs} is the daily, 0.1° resolution blended cloud-free surface temperature field generated by NOAA Coast Watch. In addition, fresh waters (salinity = 0 psu) outflow from major rivers in the MABSAB area are considered. These include Connecticut river, Hudson river, Delaware river, Susquehanna river, Potomac river, James river, Roanoke river, Cape Fear river, Pee Dee river, Santee river, Savannah river, Ogeechee river, Altamaha river, Satilla river and Johns river. For each of them, the transport is specified using the volume time series measured by United States Geological Survey (USGS) river gauges.

For open boundary conditions, the model is nested inside the 1/12 degree global data assimilative HYCOM/NCODA output (<http://www.hycom.org/>), plus harmonics of M_2 tide

from an ADCIRC tidal model simulation of the western Atlantic (<http://www.unc.edu/ims/ccats/tides>). Sponge layers are defined over 20 grid points from the eastern and southern boundaries, where the horizontal viscosity increases linearly from value used in the model interior to a factor of 10 of it at the boundaries. Within the sponge layer, 3-dimensional tracer and momentum fields are also nudged to corresponding HYCOM model fields.

With this regional model set up, we perform a MABSAB circulation hindcast from January 1, 2004 to February 28, 2005. The model initial conditions are interpolated from HYCOM fields on January 1, 2004. One caveat we identify during the course of MABSAB ROMS implementation is coastal salinity bias in HYCOM solution. When compared with $0.25^\circ \times 0.25^\circ$ HydroBase Hydrographic climatology (*Curry, 1996*), HYCOM/NCODA is found to overestimate the coastal salinity field due to the lack of fresh water river input. For instance, HYCOM surface mean (averaged between 2004/1/1 and 2007/12/31) salinity is up to 2 (6) psu higher on the shelf (at major river mouths) than the corresponding HydroBase salinity values. Surface temperature differences between the HYCOM and HydroBase are seen as well, but to a relatively smaller extent. Together, biases in salinity and temperature fields lead to a bias in the density field, which in turn results in biases in the alongshore and cross-shelf pressure gradients. To correct for such mean biases, we replace the HYCOM 3-dimensional annual mean salinity and temperature fields with the corresponding HydroBase annual means. That is:

$$\begin{aligned}
T_{coorection} &= T_{hycom} - \bar{T}_{hycom} + \bar{T}_{hydrobase} \\
S_{coorection} &= S_{hycom} - \bar{S}_{hycom} + \bar{S}_{hydrobase}
\end{aligned}
\tag{5}$$

The sea surface height slope is accordingly revised using the geostrophic relationship, resulting in an enhanced along-shelf velocity at the eastern boundary north of 40°N. The model therefore reproduces a mean equatorward alongshelf flow, which is consistent with the earlier finding that the mean southward depth-averaged alongshelf flow on the MAB increases with water depth from 3 cm/s at the 15 m isobath to 10 cm/s at the 100 m isobath (Lentz, 2008).

We will focus on the model hindcast solutions near CH (blue square in **figure 1**) in the January, 2005 to understand coastal circulation responses to two drastically different atmospheric forcing conditions. In situ observations from buoy 44014 (**Figure 3**) show between January 1 and January 14, the coastal wind is weak and generally northward (upwelling favorable). The mean east-west and north-south wind components during these two weeks are ~0.1 m/s and 1.8 m/s, respectively. Both air and sea surface temperatures are relatively constant, ranging around 11 °C. In contrast, from January 15 to January 31, there are four consecutive winter storms (northeasterly downwelling favorable) passing through the area with wind speed up to 15 m/s. The mean east-west and north-south wind components during these two weeks are 1.7 and -6.6 m/s, respectively. Compared to the first half of month, air temperature decreases by as much as 20 °C, whereas the sea surface temperature decreases by as much as 5 °C. In the following discussions, we refer the time

between January 1 and January 14 as *period I*, and the time between January 15 and January 31 as *period II* to contrast coastal ocean responses.

Indeed, comparisons of model simulated ocean surface temperature, salinity, sea surface height and surface velocity fields between January 7 and January 22 show very different ocean states during *period I* and *period II*. In *period II*, significant shelf-wide water temperature decreases are seen, especially in the southern MAB region. As a result of downwelling favorable storm wind forcing, low salinity coastal waters are pushed both southward and shoreward. In the mean time, large coastal sea levels rise occurs in both MAB and SAB, which corresponds to strong surface velocity variations. On January 7, we see prevailing cross-shelf flow, whereas on January 22, surface flow on both MAB and SAB are southward. Strong storm forcing in *period II* also affects the Gulf Stream. The speed of its surface velocity is significantly reduced during this period.

2. Model Data Comparisons

To better validate model hindcast skills in January 2005, we compare model solutions against observations from the following sources: (1) National Ocean Service (NOS) sea level data, which provide the hourly measurement of water levels; (2) National Data Buoy Center (NDBC) buoy data— NDBC moored buoys are deployed in the coastal and offshore waters measuring barometric pressure, wind, air and sea temperature etc; (3) FINCH ship survey data in January 2005, including high-resolution temperature, salinity and underway Acoustic Doppler Current Profiler (ADCP) velocity observations.

Point-by-point comparisons of coastal sea level are made at several stations in the domain for hourly time series as well as 36-hour low pass filtered rendition (**Figure 5**) to examine subtidal wind-driven Ekman and shelf water dynamics (*Gill, 1982; Brink, 1998*). Direct comparisons show the model is able to resolve hourly sea level variations reasonably well through the course of simulation. At the 95% confidence interval, the correlation coefficients between the two hourly time series are above 0.86 at all these stations. Reasonable comparisons are also seen in the subtidal sea level comparisons. Except for Beaufort, NC, correlation coefficients are all larger than 0.6. The model/data agreements in coastal sea levels suggest the model faithfully captures these dynamics. Beaufort station is in a complex bayou behind the Outer Banks and Atlantic Beach, we speculate that the less satisfying subtidal sea level comparison is due to poor quality of wind field, that is the 35-km resolution NARR product is insufficient to resolve complex air-land interaction in this sub-area, rendering ocean model skill in subtidal sea level prediction. Also the minimum depth in our model is set as 10 m, which may distort the nearshore setting of as well.

The sea surface temperature comparisons at NDBC buoy 44014 (**Figure 6**) shows that the model is able to reproduce SST temporal variations SST in January 2005. Both the buoy data and model simulation show ocean temperature decreases from 11°C to 8°C in response to winter storms. The modeled SST is generally 0.5-1.5 degree warmer than the observation. Such differences are due to a number of factors, including the spatial offsets between the buoy points measurements and the model 2-km footprint, the difference between the bulk ocean temperature measured 1-m below the surface by the buoy and temperature simulated by the top vertical model layer, and the quality of NARR heat flux forcing. The

present model experiment does not consider air-sea coupling, which has been shown to be an important element to improve ocean temperature and heat flux budget calculation in January 2005 (Nelson and He, personal communication).

A towed undulating vehicle system (ScanFish) was used in January 2005 survey of FINCH program (data courtesy, G. Gawarkiewicz, WHOI). This instrument provides high-resolution water column temperature and salinity measurements along the ship track. In order to compare with ScanFish T/S data collected on January 20, 25 and 29 (**Figure 7**), we interpolate model simulated temperature and salinity fields at the same sampling locations. We note that because the model output is archived every 12.42 hour (M_2 tidal period), whereas the ScanFish T/S data are instantaneous measure underway, we expect some temporal aliasing issues in this comparison. Nevertheless, comparisons show that the model is capable of capturing major spatial features in the along- and across-shelf temperature and salinity distributions, such as the warm salty Gulf Stream water seaward of the shelfbreak, and cold fresher MAB water on the shelf. The model underpredicts the stratification on the Diamonds Shoal on January 20, where there is warm and salt bottom water residing between 100-m and 1000-m isobaths, likely a result of Gulf Stream salty water intrusion (*Lentz, 2003*). Subsequent T/S comparisons between observations and model are significantly better on January 25 and 29 than on January 20. One model deficiency we see is that the simulated salinity at the offshore tip of the southern most cross-shelf transect is 1-2 psu fresher than observed, suggesting that more freshwater entrainment into the Gulf Stream.

FINCH program also has velocity observations from two Acoustic Doppler Current Meter moorings (Mooring 5004 and 5024, see locations on **Figure 2**), located at about 30m isobaths in the last week of January (data courtesy, D. Savidge, SKIO). A 36-hr low-pass filter is applied to both observed and simulation east-west and north-south velocity components at these two locations. Because the local isobaths change orientation moving from MAB to SAB, strong along-shelf flows are seen from the north-south (east-west) velocity component at mooring 5004 (5024), respectively on January 24 and 28-29 in response to storm forcing. Comparisons show such temporal variations are well resolved by the model.

Overall, the model-observation comparisons suggest our regional circulation model is in general capable of reproducing the observed circulation patterns in January 2005, lending confidence that the circulation dynamics analyses to be discussed next are based on realistic hydrodynamic realizations. We investigate the along- and across-shelf transports, momentum balance, vorticity and kinetic energy budgets to characterize ocean responses to drastically different atmospheric forcing between the 1st and 2nd half of January 2005.

Chapter 3: Ocean Circulation Dynamic Analyses

1. Transport Analysis

Quantifying CH shelf water volume transport is important for a variety of reasons. First, it provides a basis for evaluating the environmental impacts of potential offshore oil and gas exploration in CH regions (e.g., *Berger et al., 1994*). Second, it has biological implication as many local commercial fish and other organisms depend on the transport of eggs and larvae from spawning grounds to nursery areas (e.g., *Checkley et al., 1988*). Lastly, it plays an important role in determining the fate of dissolved organic and inorganic carbon and associated biogenic elements budget in nearby shelf and slope (e.g., *Bauer et al., 2002*).

1.1 Along-shelf Transport

To compute the along-shelf transport in January 2005, we select three across-shelf transects in the vicinity of CH (see their locations in **Figure 2**). Transect 1-3 are cross-shelf lines at the Virginia Beach (36.9° N), the CH (35.4 ° N) and the Cape Lookout (34.6 ° N), respectively. All three transects are between the coastline and the 100-m isobath (which is taken as the boundary between the shelf and slope). The cross-shelf distances of three transects are 112.2 km (transect 1), 27.3 km (transect 2) and 58.0 km (transect 3), respectively. The integrations corresponding subsurface area of these transects are 3.16×10^6 m² (transect1), 0.90×10^6 m² (transect 2) and 1.93×10^6 m² (transect 3), respectively.

Based on velocity observations during the Ocean Margins Program, *Kim et al. (2001)* estimated the mean along-shelf transport at a nearby 36.7° N transect is 0.17 Sv (southward).

Our simulated along-shelf transport (based on 36-hour low pass filtered velocity) at three selected transects are much larger (**Figure 9**) in the January, 2005. Their monthly mean transports are - 0.42 Sv (alongshore to the southward) ($1\text{Sv}=10^6 \text{ m}^3/\text{s}$), 0.35 Sv (alongshore to the northward) and 0.41 Sv (alongshore to the northward). We note significant increases in southward (or decreases in northward) transport occur at all three transects in *period II*, in response to strong downwelling favorable storm wind forcing. In *period I*, the mean alongshelf transports at three transects are -0.01 Sv, 0.59 Sv and 0.57 Sv, respectively, as opposed to -0.79 Sv, 0.14 Sv and 0.26 Sv in *period II*.

We can decompose the along-shelf velocity into geostrophic and ageostrophic components. At transect 1 for instance (**Figure 10**), the along-shelf geostrophic velocity component can be computed as $v_g = -\frac{1}{f\rho} \frac{\partial p}{\partial x}$ from the model, and the ageostrophic velocity component is the difference between the total along-shelf velocity and geostrophic velocity $v_a = v - v_g$. As expected, the along-shelf velocity is dominated by its geostrophic component in both *period I* and *period II*. The geostrophic velocity in *period I* has an interesting layered structure on the inner shelf due to the presence of freshwater plume, as such the surface flow moves southward, whereas the bottom flow moves northward. This is different on the mid-shelf and outer-shelf, where flows over the entire column flow southward and northward, respectively. The northward flow on the outer-shelf is presumably related to the influence of Gulf Stream meanders. The ageostrophic component is much weaker than its geostrophic counterpart. Located in the surface and bottom boundary layers, they are primarily generated by surface and bottom stresses by virtue of Ekman balance. In

period II, strong southward downwelling favorable wind significantly build up coastal sea level, the corresponding geostrophic currents become much stronger, and flow uniformly southward almost over the entire shelf. Ageostrophic component drastically increases too. The bottom ageostrophic currents now flow northward, counteracting the southward flowing geostrophic velocity. The contributions of geostrophic ageostrophic components will be further studied in the momentum analysis section below.

1.2 Cross-shelf Transport

Because of flow continuity, variations in along-shelf transport induce corresponding cross-shelf transport changes. Using the 100-m isobath as the shelf-slope boundary, we define the section connecting two offshore end points of cross-shelf transect 1 and 2 as the along-shelf transect N; and the section connecting the offshore end points of transect 2 and 3 as along-shelf transect S (see their locations on **Figure 2**). Time series of computed cross-shelf transport at transect N and S are quite different (**Figure 11**) in the January, 2005. The monthly mean cross-shelf transport value at transect N (S) is 0.77 (0.05) Sv, suggesting transect N is the major shelfwater export site during this site. We see the northward flow at transect 2 in *period I*, and the southward flow at transect 1 in *period II* are the two major contributors for the offshore transport at transect N, thus well correlated with the offshore transport at transect N.

Since the cross-shelf transport in January 2005 mainly occurs at transect N, the north part of 100 m isobath in our study domain, we can examine the across-shelf velocity structures along transect N (**Figure 12**). In *period I*, the cross-shelf velocity has both

shoreward and seaward velocity components, whereas in *period II*, the cross-shelf velocity is uniformly seaward, indicating significantly enhanced offshore transport in *period II*. The maximum cross shelf velocity is found at the 35.5 ° N in both periods, which happens to be the same major shelfwater entrainment site identified previously by other studies (Churchill, 1998; Gawarkiewicz, 2006).

2. Momentum Analysis

Lentz (1999) divided the North Carolina shelf into three dynamically distinct regions based on the momentum budget differences: the surf zone, the inner shelf (between surf zone and the 13 m isobath) and the mid shelf. This study found that for the mid-shelf, the along-shelf momentum balance was dominated by along-shelf wind stress, pressure gradient, and bottom stress. The cross-shelf momentum balance is predominantly geostrophic. In the following, we intend to perform a similar momentum analysis based on model hindcast solutions. Vertically integrating the horizontal momentum equation gives:

$$\int_{-h}^{\zeta} \frac{\partial u}{\partial t} dz + \int_{-h}^{\zeta} u \frac{\partial u}{\partial x} dz + \int_{-h}^{\zeta} v \frac{\partial u}{\partial y} dz + \int_{-h}^{\zeta} w \frac{\partial u}{\partial z} dz - \int_{-h}^{\zeta} f v dz = \int_{-h}^{\zeta} \left(-\frac{1}{\rho_0} \frac{\partial p}{\partial x} \right) dz + \int_{-h}^{\zeta} \frac{\partial}{\partial z} \left(K_m \frac{\partial u}{\partial z} \right) dz + \int_{-h}^{\zeta} D_u dz \quad (6)$$

$$\int_{-h}^{\zeta} \frac{\partial v}{\partial t} dz + \int_{-h}^{\zeta} u \frac{\partial v}{\partial x} dz + \int_{-h}^{\zeta} v \frac{\partial v}{\partial y} dz + \int_{-h}^{\zeta} w \frac{\partial v}{\partial z} dz + \int_{-h}^{\zeta} f u dz = \int_{-h}^{\zeta} \left(-\frac{1}{\rho_0} \frac{\partial p}{\partial y} \right) dz + \int_{-h}^{\zeta} \frac{\partial}{\partial z} \left(K_m \frac{\partial v}{\partial z} \right) dz + \int_{-h}^{\zeta} D_v dz$$

If we define the depth averaged velocity $\bar{u} = \frac{1}{h + \zeta} \int_{-h(x,y)}^{\zeta(x,y,t)} u dz$, $\bar{v} = \frac{1}{h + \zeta} \int_{-h(x,y)}^{\zeta(x,y,t)} v dz$, vertically

integrated nonlinear adv $A_x = \int_{-h(x,y)}^{\zeta(x,y,t)} \left(u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} \right) dz$ $A_y = \int_{-h(x,y)}^{\zeta(x,y,t)} \left(u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} \right) dz$;

vertically integrated pressure gradient $P_x = \int_{-h}^{\zeta} \left(-\frac{1}{\rho_0} \frac{\partial p}{\partial x} \right) dz$ $P_y = \int_{-h}^{\zeta} \left(-\frac{1}{\rho_0} \frac{\partial p}{\partial y} \right) dz$; vertically

integrated horizontal mixing $D_x = \int_{-h}^{\zeta} D_u dz$, $D_y = \int_{-h}^{\zeta} D_v dz$, and apply both the surface and bottom conditions, we can rewrite the depth averaged momentum equation as:

$$\begin{aligned} \frac{\partial \bar{u}}{\partial t} + \frac{A_x}{h + \zeta} - f \bar{v} &= \frac{P_x}{h + \zeta} + \frac{\tau^{sx}}{\rho_0(h + \zeta)} - \frac{\tau^{bx}}{\rho_0(h + \zeta)} + \frac{D_x}{h + \zeta} \\ \frac{\partial \bar{v}}{\partial t} + \frac{A_y}{h + \zeta} + f \bar{u} &= \frac{P_y}{h + \zeta} + \frac{\tau^{sy}}{\rho_0(h + \zeta)} - \frac{\tau^{by}}{\rho_0(h + \zeta)} + \frac{D_y}{h + \zeta} \end{aligned} \quad (7)$$

The left hand of the new equations includes the acceleration term, nonlinear advection, Coriolis force term, whereas the right hand of the equations has the pressure gradient term, the wind stress term, bottom stress term, and the horizontal mixing term. Each term can be calculated from ROMS diagnostics output. To highlight differences in momentum balances between the two periods, we average each term over *period I* and *period II*, respectively (**Figure 13**).

For both u- and v- momentum equations, the two largest terms are Coriolis term and pressure gradient term, and they are generally balanced each other. The sum of Coriolis force and pressure gradient force terms constitutes the ageostrophic momentum, which is balanced by local acceleration, nonlinear advection, surface and bottom stream and horizontal mixing. The depth-averaged momentum term is larger on the shelf than in the deep ocean. The temporal mean acceleration terms in *period I* and *period II* are small. The nonlinear advection terms are relatively noisy, containing many small-scale variations. Their largest values are seen in the SAB along the Gulf Stream and at the entrance of shallower estuaries.

The effects of wind stress and bottom stress are most clearly presented on the shelf. Note that the shelf-wide surface wind stress in *period II* increases up to 5 times of that in *period I*. Because of the quadratic law being used by the model, the bottom stress is proportional to the square of bottom velocity. The largest bottom stress is seen to be associated with Gulf Stream in the SAB. In *period II*, strong storms lead to much large bottom velocity in the alongshelf direction, and subsequently significantly enhanced bottom stress.

Generally speaking, the shelf-wide ageostrophic balance during period I is achieved between nonlinear advection and bottom stress, whereas the balance during *period II* is accomplished jointly by nonlinear advection, surface and bottom stress.

The vertical distributions of the 3-dimensional momentum budget (**Figure 14**) can be gleaned along transect 1 off the Virginia Beach. In both *period I* and *period II*, the nonlinear advection term shows some horizontal structures, but not significant vertical structures are seen. The vertical viscosity term is the major contributor of the ageostrophic term ($u\partial v/\partial y + v\partial u/\partial x$). In response to strong storm forcing, both terms increase by 4-5 times in *period II* than in period I.

3. Vorticity Analysis

Another useful means to quantify shelf circulation responses is to analyze the vorticity balance. If starting from the primitive equations, after vertically averaging the momentum equations, taking the curl of the depth averaged momentum equations, we get:

$$\begin{aligned} \frac{\partial}{\partial t} \left(\frac{\partial \bar{v}}{\partial x} - \frac{\partial \bar{u}}{\partial y} \right) + \left(\frac{\partial A_y / (h + \zeta)}{\partial x} - \frac{\partial A_x / (h + \zeta)}{\partial y} \right) + \left(\frac{\partial f \bar{u}}{\partial x} + \frac{\partial f \bar{v}}{\partial y} \right) = & \left(\frac{\partial P_y / (h + \zeta)}{\partial x} - \frac{\partial P_x / (h + \zeta)}{\partial y} \right) + \left(\frac{\partial D_y / (h + \zeta)}{\partial x} \right. \\ & \left. - \frac{\partial D_x / (h + \zeta)}{\partial y} \right) + \left(\frac{\partial \tau^{sy} / \rho_0 (h + \zeta)}{\partial x} - \frac{\partial \tau^{sx} / \rho_0 (h + \zeta)}{\partial y} \right) + \left(\frac{\partial \tau^{by} / \rho_0 (h + \zeta)}{\partial x} - \frac{\partial \tau^{bx} / \rho_0 (h + \zeta)}{\partial y} \right) \end{aligned} \quad (8)$$

If we define $H = h + \zeta$; $M = (\bar{u}\vec{i} + \bar{v}\vec{j}) * H$; $A = A_x \vec{i} + A_y \vec{j}$; $D = D_x \vec{i} + D_y \vec{j}$

The Coriolis force curl can be rewrite as

$$\begin{aligned} \frac{\partial f \bar{u}}{\partial x} + \frac{\partial f \bar{v}}{\partial y} = \frac{\partial (f/H)(\bar{u} * H)}{\partial x} + \frac{\partial (f/H)(\bar{v} * H)}{\partial y} = (\bar{u}H) * \frac{\partial (f/H)}{\partial x} + (\bar{v}H) * \frac{\partial (f/H)}{\partial y} + \\ (f/H) \frac{\partial (\bar{u} * H)}{\partial x} + (f/H) \frac{\partial (\bar{v} * H)}{\partial y} = (\bar{u}H) * \frac{\partial (f/H)}{\partial x} + (\bar{v}H) * \frac{\partial (f/H)}{\partial y} = M \cdot \nabla \left(\frac{f}{H} \right) \end{aligned} \quad (9)$$

We can then rewrite the depth-averaged vorticity equation as follows:

$$\frac{\partial}{\partial t} \text{curl}_z \left(\frac{M}{H} \right) + M \cdot \nabla \left(\frac{f}{H} \right) = \left(\frac{\partial (P_y / H)}{\partial x} - \frac{\partial (P_x / H)}{\partial y} \right) + \text{curl}_z \left(\frac{\tau_s - \tau_b}{\rho_o H} \right) + \text{curl}_z \left(\frac{D}{H} \right) - \text{curl}_z \left(\frac{A}{H} \right) \quad (10)$$

The first term on the left hand of equation represents the local rate of change of the vorticity of the depth averaged flow. The second term on the left hand equation is the advection of the geostrophic potential vorticity (APV), representing transport across the f/H contours. If $APV=0$, the transport is exactly along the f/H contour. If $APV > 0$, the transport is in the onshore direction and the slope water and Gulf Stream tend to move shoreward. If $APV < 0$, the transport is in the offshore direction, thus shelfwater entrainment would occur. On the right hand the equation are the pressure gradient force curl, wind stress curl, bottom stress curl and horizontal diffusion curl and curl of nonlinear advection.

The pressure gradient force curl is equivalent to the Joint Effect of Baroclinicity and Relief (JEBAR) (*Sakisyan, 2006*) term if we assume $\zeta=0$ (rigid lid approximation) (for

detailed derivation, see appendix A). The JEBAR is firstly derived as $J(\int_{-h}^0 \frac{\rho g z}{\rho_0} dz, 1/h)$,

where $J(A, B) = \frac{\partial A}{\partial x} \frac{\partial B}{\partial y} - \frac{\partial B}{\partial x} \frac{\partial A}{\partial y}$, and then can be calculated from the model output history

data. Computing from the history data induces larger errors than computing from the diagnostic data, so JEBAR is calculated as the curl the pressure gradient force. The pressure gradient curl can be calculated directly from the pressure gradient force, which can be obtained from the model output diagnostic data. JEBAR term has been shown to determine the shoreward motion of Kuroshio northeast of Taiwan (Oey, 2010) and the seasonal veering of Kuroshio pathway (Guo, 2002).

Temporal means of depth-averaged vorticity budget are computed for *period I* and *period II*, respectively (**Figure 15**). Being the spatial derivate of momentum, each term in vorticity equation is fairly noisy. The local rate of change in vorticity is relatively small in both periods. The APV term on the other hand shows large patch of negative values near the CH, representing seaward motion of the Gulf Stream. This large negative value of the APV near CH is mainly balanced by the JEBAR term. Therefore, the Gulf Stream's cross-isobath veering near CH is consistent with Kuroshio veering near Taiwan (Guo, 2002). Both wind stress curl and bottom stress curl significantly increase and expand to the entire study domain in *period II* compared to in *period I*. the wind stress curl is relatively small while in the *period II*, the wind curl is much larger. In *period II*, the wind curl is positive in the inner and outer MAB shelf and negative in the middle shelf. In contract, the bottom friction curl in

period II is positive in the middle shelf and negative in the inner and outer shelf. Note that the wind and bottom vorticity terms contain the curl of the bathymetry as well.

The relationship between the cross-shelf transport and APV is not evident from the rather noisy spatial maps of vorticity budget terms. To examine this, we zoom in to 100-m isobath, and compare the along-isobath distributions of the temporal means of cross-shelf velocity and APV (**Figure 16**). In general, the relationship between the cross-shelf transport and APV is valid. In *period I*, the cross-shelf velocity has both shoreward and seaward components. In *period II*, the cross shelf velocity is uniformly seaward. As shown in previous section, the maximum cross-shelf velocity occurs at the 35.5° N, and maximum velocity in *period I* is larger than that in *period II*. Correspondingly, in *period I* the APV term has both the positive and negative value, while in *period II* the APV term is general negative. The minimum of the APV is also along the 35.5° N, and the minimum of the APV in *period I* is smaller than that in *period II*.

To piece out the relative contribution of each vorticity budget terms in determining APV variation, we sample each term along the MAB portion of 100-m isobath, then spatially average them to come up a set of 1-month long time series for these terms (**Figure 17**). The reason we only choose the north part of the transect N is because in the southern part of transect N, the bottom stress curl is much larger than all the other terms, probably due to the Gulf Stream's effect. In *period I* all terms are relatively small. Note both local rate of change of vorticity and curl of diffusion remain small through the entire record. In *period II*, both the wind curl and vorticity nonlinear advection (ADV) are positive. They contribute to the

positive APV, providing the flow shoreward motion tendency. This is to some extent offset by both the bottom stress curl and JEBAR, which are negative, and favoring flow seaward motion. At any rate, *period II* is characterized with significant variation in these vorticity budget terms, highlighting very different coastal ocean responses during the *period II* compared to *period I*.

4. Kinetic Energy Analysis

The energy conservation is one of the most fundamental laws of the nature. Energy in the ocean can be classified into two major parts: the mechanical energy and the internal energy. The mechanical energy further consists of two parts: the kinetic energy and the gravitational energy. On the global scale, wind stress and tidal forcing are the two most important sources of mechanical energy that drive the ocean general circulation. Although mechanical energy flux is 1000 times smaller than the heat flux, it controls the strength of the general ocean circulation (Huang, 2004). MacCready *et al* (2008) showed the mechanical energy budget analysis provides an insightful way to explore the tidal and wind induced mixing in the estuaries and shelf areas. We intend to apply the same method to better characterize the nature of CH shelf circulation in *period I* and *period II* of January 2005.

Following MacCready *et al* (2008), we neglect the horizontal mixing term (which is negligibly small), and time u , v to the corresponding momentum equation, we have:

$$\begin{aligned}
u \frac{du}{dt} - fvu &= -\frac{u}{\rho_0} \frac{\partial p}{\partial x} + u \frac{\partial}{\partial z} (K_m \frac{\partial u}{\partial z}) \\
v \frac{dv}{dt} + fuv &= -\frac{v}{\rho_0} \frac{\partial p}{\partial x} + v \frac{\partial}{\partial z} (K_m \frac{\partial v}{\partial z})
\end{aligned} \tag{11}$$

The two equations can be combined as:

$$\frac{d(\frac{1}{2} \rho_0 u^2 + \frac{1}{2} \rho_0 v^2)}{dt} = (-u \frac{\partial p}{\partial x} - v \frac{\partial p}{\partial y}) + \rho_0 u \frac{\partial}{\partial z} (K_m \frac{\partial u}{\partial z}) + \rho_0 v \frac{\partial}{\partial z} (K_m \frac{\partial v}{\partial z}) \tag{12}$$

Define the horizontal kinetic energy (KE) and horizontal pressure work as:

$$\begin{aligned}
KE &= (\frac{1}{2} \rho_0 u^2 + \frac{1}{2} \rho_0 v^2) \\
PW &= (-u \frac{\partial p}{\partial x} - v \frac{\partial p}{\partial y})
\end{aligned} \tag{13}$$

Next, volume integrating the kinetic energy equation gives:

$$\int_v \frac{dKE}{dt} dV = \int_v (-u \frac{\partial p}{\partial x} - v \frac{\partial p}{\partial y}) dV + \int_v (\rho_0 u \frac{\partial}{\partial z} (K_m \frac{\partial u}{\partial z}) + \rho_0 v \frac{\partial}{\partial z} (K_m \frac{\partial v}{\partial z})) dV \tag{14}$$

Following the method of the *MacCready et al* (2008), the first term on the left hand of the equation can be derived as the KE storage and KE advection term, and the second and the third terms on the right hand of equation can be derived as wind stress work and dissipation terms. That is:

$$\int_v \frac{dKE}{dt} dV = \frac{\partial}{\partial t} (KE_v dV) + \int_{A_{open}} (KE_v) u_n dA \tag{15}$$

$$\int_v (\rho_0 u \frac{\partial}{\partial z} (K_m \frac{\partial u}{\partial z}) + \rho_0 v \frac{\partial}{\partial z} (K_m \frac{\partial v}{\partial z})) dV = \int_{A_0} (u \tau^x + v \tau^y) |_{z=\eta} dA - \int_v \rho_0 K_m (u_z^2 + v_z^2) dV \tag{16}$$

The vertically integrated mechanical energy equation can be rewritten as:

$$\frac{\partial}{\partial t}(KE_v dV) = - \int_{A_{open}} (KE_v) u_n dA - PW + \int_{A_0} (u \tau^x + v \tau^y) |_{z=\eta} dA - \int_V \rho_0 K_m (u_z^2 + v_z^2) dV \quad (17)$$

representing the balance between the kinetic energy (KE) storage (or the time rate of the KE change) on the left hand side, and the sum of the advection of the kinetic energy (first term on the right hand side), horizontal pressure work (PW, 2nd term on the right hand side), wind stress work (3rd term on the right hand side), and dissipation due to the bottom friction and vertical mixing (4th term on the right hand side). As a check of our calculation, we add all terms in this kinetic energy equation together. The residual is not exactly zero due to numerical round off error, but is indeed several order of magnitude smaller compared to each of KE budget terms.

Each of the kinetic energy budget terms are calculated from the model output history and diagnostic files. **Figure 18** shows the spatial distributions of time and depth-averaged mechanical energy budget terms. The largest values of advection work (ADV) and pressure work (PW) occur seaward in the Gulf Stream area, and they are generally balanced between each other. In both periods, the wind stress work is positive on the shelf, suggesting the large-scale mean wind flows in the same direction of mean shelf current. Winds are much stronger in the *period II*, resulting much larger wind stress work on the shelf. For the same argument, because the means wind flows in the opposite direction of the Gulf Stream, the wind stress work is negative in the Gulf Stream area. The dissipation work is negative in both

periods and largest values are seen in the Gulf Stream area. Similar to the wind stress work, the dissipation work is larger on the shelf during the *period II* than *period I*.

To understand how the shelf circulation kinetic energy budget varies over time, we now focus on the region shallower than 100 m isobath. This part of coastal ocean has an area (volume) of $5.3 \times 10^{10} \text{ m}^2$ ($8.9 \times 10^{11} \text{ m}^3$), respectively. Temporal variations of the shelf-area averaged kinetic energy budget are shown in Figure 19. Note the high frequency variations in both KE storage term and PW are related to the M2 tidal sea level change. In *period I*, the dominant terms are KE storage, PW and dissipation work, whereas in *period II* the wind work and dissipation work become the most dominant terms. As shown earlier, both wind stress work and dissipation work got amplified in *period II*, and as expected, they indeed constitute major energy balance during the stormy 2nd half of January 2005. Throughout the month, the area averaged advection work plays a rather small role in shelf KE budget.

Chapter 4: Discussions and Conclusions

We use a realistic regional ocean circulation model to hindcast the circulation near CH region in January, 2005. Model results are compared with the observed coastal sea level, temperature, salinity and velocity observations. Generally good agreement are found, indicting the model has reasonable intrinsic hindcasting skills. Given time and space continuous model solutions, we perform analyses of along- and cross-shelf transport, momentum, vorticity and energy balance diagnostic to quantify coastal ocean responses to different atmospheric forcing conditions in January 2005. The first half of month (*period I*: January 1-14) is characterized by light, upwelling favorable wind, whereas the 2nd half (*period II*: January 15 -31) is characterized by consecutive storms with strong, southward downwelling favorable winds. In terms of shelf water transport, during the stormy 2nd half of month (*period II*), both southward along-shelf transport and seaward cross-shelf transport significantly increase. The along-shelf transport change is mainly due to the geostrophic velocity change associated with the coastal sea level setup. In terms of momentum balance, the wind stress during this time is mainly balanced by the bottom friction and the residual of the Coriolis and pressure gradient force. Similarly in the vorticity balance, the bottom stress curl offsets much of wind stress curl. The JEBAR (Joint Effect of Baroclinicity and Relief) effect is the most significant contributor of APV (Advection of Potential Vorticity). The dominant kinetic energy balance on the shelf is between the time rate of kinetic energy change and the pressure work (PW) during *period I*, whereas during the stormy *period II*, the main shelf energy balance is achieved between wind stress work and dissipation work.

Comparisons with the FINCH ScanFish hydrographic data show that the model has missed warm and salty bottom water on the Diamond Shoal. This feature is likely related to Gulf Stream water subsurface intrusion. Indeed, hydrographic conditions near CH are sensitive to the exact locations of Gulf Stream and its meanders. One approach of future model improvement is to assimilate satellite altimetry sea surface height and other coastal in-situ observations (e.g., HF Radar, ship hydrographic observations) to better constrain model fields. The newly available 4-dimensional variational (4DVAR) data assimilation technique has been successfully used in MAB circulation study (e.g., Zhang et al., 2010; Chen et al, in preparation). Another way of model improvement is to refine surface forcing by considering air-sea coupling during strong winter extratropical cyclones as those occurred in *period II*. Nelson and He (2011, submitted) use a modeling system that couples ROMS with Weather Research and Forecast (WRF) model. Air-sea coupled model simulations are performed for the same January 2005 period, and shows the ocean is play an important role in shaping coastal wind and heat flux fields, which in turn affect the shelf water stratification and velocity fields.

Ford (1952) firstly reported that MAB shelf water can be entrained by the Gulf Stream east of the CH. Our model simulation clearly shows the presence of this kind of entrainment (**Figure 4**). These entrained shelf waters are mixed with the Gulf Steam, but detailed of mixing processes and the fate of these shelf waters deserve further study in the future. One other unanswered question is how often and how much of the MAB shelf waters can move into the SAB? Pietrafesa (1994) suggested the MAB waters “often” pass by Cape Hatters and enter the SAB. Our preliminary surface particle trajectory calculation (not

shown) in January 2005 suggest the MAB waters that entered this SAB were also entrained in the Gulf Stream, in a way similar to the surface circulation pattern suggested by Gray (1963). The 3-dimensional ocean Lagrangian transport pathway near CH is therefore another important question needing to be further explored.

REFERENCES

- Aikman, F., III, H.W. Ou, and R.W. Houghton (1988): Current variability across the New England continental shelf-break and slope. *Continental Shelf Research*, 8, 625-651.
- Atkinson, L., T. Lee, J. Blanton, and W. Chandler (1983): Climatology of the Southeastern United States Continental Shelf Waters, *J. Geophys. Res.*, 88(C8), 4705-4718.
- Bane, J. M., Jr., O. B. Brown, R. H. Evans, and P. Hamilton (1988): Gulf Stream remote forcing of shelfbreak currents in the Mid-Atlantic Bight, *Geophys. Res. Lett.*, 15(5), 405-407.
- Bauer, J.E., D.J. DeMaster, D.J. Repeta and P.G. Verity (2002): Biogeochemistry and cycling of carbon in the northwest Atlantic continental margin: findings of the Ocean Margins Program. *Deep-Sea Res. II*, 49, 4271-4709.
- Beardsley, Robert C., David C. Chapman, Kenneth H. Brink, Steven R. Ramp, Ronald Schlitz (1985): The Nantucket Shoals Flux Experiment (NSFE79). Part I: A Basic Description of the Current and Temperature Variability. *J. Phys. Oceanogr.*, 15, 713-748.
- Bell, Michael M., Michael T. Montgomery, (2008): Observed Structure, Evolution, and Potential Intensity of Category 5 Hurricane Isabel (2003) from 12 to 14 September. *Mon. Wea. Rev.*, 136, 2023-2046. doi: 10.1175/2007MWR1858.1
- Berger, T.J., W.C. Boicourt, J.H.Churchill, P.Hamilton, R.J.Wayland, and D.R.Watts (1994): A physical oceanographic field program offshore of North Carolina. Tech. Rep. MMS 94, Science Applications International Corporation, 463 pp.
- E. Bohm, T.S. Hopkins, L.J. Pietrafesa and J.H. Churchill (2006): Continental slope sea level and flow variability induced by lateral movements of the Gulf Stream in the Middle Atlantic Bight, *Progress in Oceanography*, 70, 196-212.
- Brink, K. H. (1998), Wind-driven currents over the continental shelf, in *The Sea*, 10, 3-20, K.H. Brink and A. R. Robinson eds., Wiley, N.Y.
- Brown, Wendell S., Neal R. Pettigrew, James D. Irish (1985): The Nantucket Shoals Flux Experiment (NSFE79). Part II: The Structure and Variability of Across-Shelf Pressure Gradients. *J. Phys. Oceanogr.*, 15, 749-771.
- Bumpus, D.F. (1973): A description of the circulation on the continental shelf of the east coast of the United States. *Prog. Oceanogr.*, 6, 111-157

- Castelao, R., S. Glenn, and O. Schofield (2010): Temperature, salinity, and density variability in the central Middle Atlantic Bight, *J. Geophys. Res.*, 115, C10005, doi:10.1029/2009JC006082.
- Chant, R., et al. (2008): Dispersal of the Hudson River plume in the New York Bight, *Oceanography*, 21, 148–161.
- Chao, S.-Y. (1992): An air-sea interaction model for cold air outbreaks, *J. Phys. Oceanogr.*, 22, 821–841.
- Checkley, D. M., S. Raman, G. L. Maillet and K. M. Mason (1988): Winter storm effects on the spawning and larval drift of a pelagic fish. *Nature*, 335, 346–348.
- Chen, K., He, R. (2010): Numerical Investigation of the Middle Atlantic Bight Frontal Circulation using a High Resolution Ocean Hindcast Model. *Journal of Physical Oceanography* DOI: 10.1175/2009JPO4262.1
- Chen K., He, R. et al, Data Assimilative Modeling Investigation on Gulf Stream Warm Ring Interaction with Continental Shelf and Slope, Part 1: Method, *Journal of Geophysical Research* (in prep.).
- Churchill, J. and P. Cornillon (1991): Gulf Stream water on the shelf and upper slope north of Cape Hatteras. *Continental Shelf Research*, 11, 409–431
- Churchill, J. H., and T. J. Berger (1998): Transport of Middle Atlantic Bight shelf water to the Gulf Stream near Cape Hatteras, *J. Geophys. Res.*, 103(C13), 30,605 - 30,621, doi:10.1029/98JC01628.
- Churchill, J. H., and G. Gawarkiewicz (2009): Shelfbreak frontal eddies over the continental slope north of Cape Hatteras, *J. Geophys. Res.*, 114, C02017, doi:10.1029/2007JC004642.
- Cione, J. J., S. Raman, and L. J. Pietrafesa (1993): The effect of Gulf Stream-induced baroclinicity on the U.S. east coast winter cyclones. *Mon. Wea. Rev.*, 121, 421–430.
- Curry, R.G. (1996): HydroBase: a database of hydrographic stations and tools for climatological analysis. Technical Report WHOI-96-01, Woods Hole Oceanographic Institution, 44 pp.
- Dirks, R. A., J. P. Kuettner, and J. A. Moore (1988): Genesis of Atlantic Lows Experiment (GALE): An experiment, *Bull Am. Meteorol. Soc.*, 69, 148–160.
- Fairall, C. W., E. F. Bradley, J. E. Hare, A. A. Garchev, and J. Edson (2003): Bulk parameterization of air-sea fluxes: Updates and verification for the COARE algorithm. *Journal of Climate*, 16, 571–591.

- Fisher, A., Jr. (1972): Entrainment of Shelf Water by the Gulf Stream Northeast of Cape Hatteras, *J. Geophys. Res.*, 77(18), 3248–3255, doi:10.1029/JC077i018p03248.
- Flagg, C. N., M. Dunn, D.-P. Wang, H. T. Rossby, and R. L. Benway (2006): A study of the currents of the outer shelf and upper slope from a decade of shipboard ADCP observations in the Middle Atlantic Bight, *J. Geophys. Res.*, 111, C06003, doi:10.1029/2005JC003116.
- Flagg, C. N., D. Wallace, and Z. Kolber (1998): Cold anticyclonic eddies formed from cold pool water in the southern Middle Atlantic Bight, *Cont. Shelf Res.*, 17, 1839–1867.
- Ford, W.L. (1952): On the nature, occurrence and origin of cold low salinity water along the edge of the Gulf Stream. *Journal of Marine Research*, 11, 281-293.
- Fratantoni, P. S., R. S. Pickart, D. J. Torres, A. Scotti (2001): Mean structure and dynamics of the shelfbreak Jet in the Middle Atlantic Bight during fall and winter. *J. Phys. Oceanogr.*, 31, 2135–2156.
- Garfield, N., III, and D. L. Evans (1987): Shelf Water Entrainment by Gulf Stream Warm-Core Rings, *J. Geophys. Res.*, 92(C12), 13,003–13,012, doi:10.1029/JC092iC12p13003.
- Garvine, R. W., K.-C. Wong, G. Gawarkiewicz, R. K. McCarthy, R. W. Houghton, and F. Aikman III (1988): The morphology of shelf break eddies, *J. Geophys. Res.*, 93, 15,593–15,607.
- Gawarkiewicz, G., T. M. Church, G. W. Luther III, T. G. Ferdelman, and M. Caruso (1992), Large-scale penetration of Gulf Stream water onto the Continental Shelf north of Cape Hatteras, *Geophys. Res. Lett.*, 19(4), 373-376
- Gawarkiewicz, G. G., J. H. Churchill, F. B. Bahr, C.A. Linder, C. D. Marquette (2009): Shelfbreak frontal structure and processes north of Cape Hatteras in winter, *Journal of Marine Research*, 66, 755-799.
- Gawarkiewicz, G. G., C. A. Linder (2005): Lagrangian flow patterns north of Cape Hatteras using near-surface drifters, *Progress In Oceanography*, 70, 181-195.
- Gill, A.E., (1982): *Atmosphere Ocean Dynamics*. Academic Press, 662 pp
- Gray, I. E. and M. J. Cerame-Vivas (1963): The Circulation of Surface Waters in Raleigh Bay, North Carolina, *Limnology and Oceanography*, 8, 330-337.
- Guo, Xinyu, Hisashi Hukuda, Yasumasa Miyazawa, Toshio Yamagata (2003): A Triply Nested Ocean Model for Simulating the Kuroshio—Roles of Horizontal Resolution on JEBAR. *J. Phys. Oceanogr.*, 33, 146–169.
- He, R., R. H. Weisberg (2002), West Florida Shelf Circulation and Heat Budget for 1999 Spring Transition. *Continental Shelf Research*, Vol. 22, 5, 719-748

- Houghton, R. W., R. Schlitz, R. C. Beardsley, B. Butman, and J. L. Chamberlin (1982): Middle Atlantic cold pool: Evolution of the temperature structure during summer 1979, *J. Phys. Oceanogr.*, 12, 1019–1029.
- Houghton, R. W., D. B. Olson, and P. J. Celone (1986): Observation of an anticyclonic eddy near the continental shelf break south of New England, *J. Phys. Oceanogr.*, 16, 60–71.
- Houghton, R. W., F. Aikman III, and H. W. Ou (1988): Shelf-slope water frontal structure, and cross-shelf exchange at the New England shelfbreak. *Cont. Shelf Res.*, 8, 687–710.
- Houghton, R. W., C. N. Flagg, and L. J. Pietrafesa (1994): Shelf-slope water frontal structure, motion, and eddy heat flux in the southern Middle Atlantic Bight. *Deep-Sea Res. II*, 41, 273–306.
- Huang, R. X. (2004): Ocean, energy flows in. *Encyclopedia of Energy*, C. J. Cleveland, Ed., Vol. 4, Elsevier, 497–509.
- Kim, Y. Y., G. L. Weatherly, and L. J. Pietrafesa (2001): On the mass and salt budgets for a region of the continental shelf in the southern Mid-Atlantic Bight, *J. Geophys. Res.*, 106, 31,263–31,282, doi:10.1029/2000JC000738.
- Kumar A., A. Valle-Levinson, L.P. Atkinson (2006): Overrunning of shelf water in the southern Mid-Atlantic Bight, *Progress In Oceanography*, 70, 213-232.
- Kupferman, S. L., and N. Garfield (1977): Transport of Low-Salinity Water at the Slope Water-Gulf Stream Boundary, *J. Geophys. Res.*, 82(24), 3481–3486, doi:10.1029/JC082i024p03481.
- Lee, T. N., J. A. Yoder, and L. P. Atkinson (1991): Gulf Stream Frontal Eddy Influence on Productivity of the Southeast U.S. Continental Shelf, *J. Geophys. Res.*, 96(C12), 22,191–22,205, doi:10.1029/91JC02450.
- Lentz, S., R. T. Guza, S. Elgar, F. Feddersen, and T. H. C. Herbers (1999): Momentum balances on the North Carolina inner shelf, *J. Geophys. Res.*, 104(C8), 18,205–18,226, doi:10.1029/1999JC900101.
- Lentz, S. J. (2003), A climatology of salty intrusions over the continental shelf from Georges Bank to Cape Hatteras, *J. Geophys. Res.*, 108(C10), 3326, doi:10.1029/2003JC001859
- Lentz, Steven J. (2008): Observations and a Model of the Mean Circulation over the Middle Atlantic Bight Continental Shelf. *J. Phys. Oceanogr.*, 38, 1203–1221.
- Li, Y., H. Xue, and J. M. Bane (2002), Air-sea interactions during the passage of a winter storm over the Gulf Stream: A three-dimensional coupled atmosphere-ocean model study, *J. Geophys. Res.*, 107(C11), 3200, doi:10.1029/2001JC001161.

- Lillibridge, J. L., III, G. Hitchcock, T. Rossby, E. Lessard, M. Mork, and L. Golmen (1990): Entrainment and mixing of shelf/slope waters in the near-surface Gulf Stream, *J. Geophys. Res.*, 95(C8), 13,065–13,087.
- Linder, C., and G. Gawarkiewicz (1998): A climatology of the shelfbreak front in the Middle Atlantic Bight, *J. Geophys. Res.*, 103(C9), 18405-18423.
- MacCready, P., N. S. Banas, B. M. Hickey, E. P. Dever, Y. Liu (2009): A model study of tide- and wind-induced mixing in the Columbia River Estuary and plume, *Continental Shelf Research*, 29, 278-291.
- Mellor, G. L., and T. Yamada (1982), Development of a turbulence closure model for geophysical fluid problems, *Rev. Geophys.*, 20(4), 851–875, doi:10.1029/RG020i004p00851.
- Mountain, D. G. (2003), Variability in the properties of Shelf Water in the Middle Atlantic Bight, 1977–1999, *J. Geophys. Res.*, 108(C1), 3014, doi:10.1029/2001JC001044
- Nelson, J. S, and R.He (2011): Effect of the Gulf Stream on winter extratropical cyclone outbreak. *J. Geophys. Res.*(submitted)
- Oey, L-Y., Y-C. Hsin, and C-R. Wu (2010): Why does the Kuroshio northeast of Taiwan shift shelfward in winter? *Ocean Dyn.*, 60, 413–426.
- Pietrafesa, L.J., J.M. Morrison, M.P. McCann, J. Churchill, E. Bohm, R.W. Houghton (1994): Water mass linkages between the Middle and South Atlantic bights, *Deep Sea Research*, 41, 365-389.
- Sarkisyan, A.S. (2006): Forty years of JEBAR -- the finding of the Joint Effect of Baroclinicity and Bottom Relief for the modeling of ocean climatic characteristics, *Atmos. Ocean. Phys.* 42, 534–554.
- Savidge, D. K., and J. M. Bane Jr. (2001): Wind and Gulf Stream influences on along-shelf transport and off-shelf export at Cape Hatteras, North Carolina, *J. Geophys. Res.*, 106, 11,505–11,527, doi:10.1029/2000JC000574.
- Savidge, D. K., and J. A. Austin (2007): The Hatteras Front: August 2004 velocity and density structure, *J. Geophys. Res.*, 112, C07006, doi:10.1029/2006JC003933.
- Shaw, P.T, L, J. Pietrafesa, C.N. Flagg, R. W. Houghton, and K. H. Su (1994): Low-frequency oscillations on the outer shelf in the southern Mid-Atlantic bight. *Deep-Sea Research II*, 41, 253-271.
- Shchepetkin, A. F., and J. C. McWilliams (2005): The Regional Ocean Modeling System: A split-explicit, free-surface, topography following coordinates ocean model, *Ocean Modelling*, 9, 347-404.

- Shearman, R. K., and S. J. Lentz (2003): Dynamics of mean and subtidal flow on the New England shelf, *J. Geophys. Res.*, 108(C8), 3281, doi:10.1029/2002JC001417.
- Voorhis, A. D., D. C. Webb, and R. C. Millard (1976): Current structure and mixing in the shelf/slope water south of New England. *J. Geophys. Res.*, 81, 3695–3708.
- Wayland, R., and S. Raman (1989): Mean and turbulent structure of a baroclinic marine boundary layer during the 28 January 1986 cold air outbreak (GALE 86), *Boundary Layer Meteorol.*, 48, 227–254.
- Weber, A. H., J. O. Blanton (1980): Monthly Mean Wind Fields for the South Atlantic Bight. *J. Phys. Oceanogr.*, 10, 1256–1263.
- Wood, A. M., N. D. Sherry, and A. Huyer (1996): Mixing of chlorophyll from the Middle Atlantic Bight cold pool into the Gulf Stream at Cape Hatteras in July 1993, *J. Geophys. Res.*, 101(C9), 20,579–20,593.
- Xue, H., Z. Pan, and J. M. Bane Jr. (2000): A 2D coupled atmosphere-ocean model study of air-sea interactions during a cold air outbreak over Gulf Stream, *Mon. Weather Rev.*, 128, 973–996
- Zhang, W.G., J.L. Wilkin, H.G. Arango (2010): Towards building an integrated observation and modeling system in the New York Bight using variational methods, Part I: 4DVAR data assimilation, *Ocean Modelling*, 35. 119-133.
- Zishka, K. M., and Smith, P. J. (1980): The climatology of cyclones and anticyclones over North-America and surrounding ocean environs for January and July, 1950-77, *Monthly Weather Review*, 108, 387-401.

Figure 1: The MABSAB circulation model domain, bathymetry and locations of rivers considered in the simulation. Analyses in this study focus on the CH region are highlighted by the blue square box.

Figure 2: A schematic adopted from *Churchill* [2009] showing ocean circulation setting near CH [left]. SF represents the shelfbreak frontal jet. HF represents the Hatteras Front. [Right] Locations of NDBC buoy 44014, and FINCH current meter stations (5004 and 5042). Also shown are the locations of three cross-shelf transects and one (green line) along-shelf transect at 100-m isobath.

Figure 3: Time series of surface wind, surface air temperature and sea surface temperature measured by NDBC buoy 44014 in January, 2005.

Figure 4: Snapshots of simulated sea surface temperature (SST), sea surface salinity (SSS), sea surface height (SSH) and surface velocity on January 7 (a) and January 22 (b), 2005.

Figure 5: Comparisons of observed (blue) and simulated (red) sea levels at four stations near Cape Hatteras. The left column shows the hourly time series comparison, whereas the right column shows comparisons of their 36-hour low pass filter renditions. R value represents the correlation coefficient (at 95% confidence interval) between observation and simulation in given each panel.

Figure 6: SST comparison in January 2005. Observed SST is measured by NDBC buoy 44014.

Figure 7: 3-D temperature and salinity comparisons between the model hindcast and FINCH ScanFish survey data on January 20th (top panels), 25th (middle panels) and 29th (lower panels). Locations of observations (ship tracks) are shown in the left column.

Figure 8: Time series comparisons of simulated and observed north-south, and east-west velocity components. Observations are taken by FINCH ADCP moorings in the last week of January, 2005. R values represent the correlation coefficient (at 95% confidence interval) between observation and simulation.

Figure 9: Time series of simulated along-shelf transport (unit: Sv) at transect 1 (upper), transect 2 (middle) and transect 3 (lower) in January, 2005. Positive represents the alongshore northward flow and negative represents the alongshore southward flow.

Figure 10. Sectional views of time-averaged along-shelf velocity, along-shelf geostrophic velocity and along-shelf ageostrophic velocity at transect 1 in *period I* (a) and *period II* (b).

Figure 11: Time series of cross-shelf transport (unit: Sv) at transect N (upper) and transect S (lower) in January, 2005. Positive (negative) values represent the seaward (shoreward) transport across the transect.

Figure 12: Time-averaged cross-isobath velocity (m/s) profiles at transect N in *period I* (a) and *period II* (b).

Figure 13: Spatial views of temporal mean depth-averaged u momentum balance in *period I* (a) and *period II* (b) and v momentum balance in *period I* (c) and *period II* (d).

Figure 14. Sectional view of temporal mean u momentum balance in *period I* (a) and *period II* (b) and temporal mean v momentum budget in *period I* (c) and *period II* (d)

$$\frac{\partial}{\partial t} \text{curl}_z \left(\frac{M}{H} \right) + \text{curl}_z \left(\frac{A}{H} \right) + M \cdot \nabla \left(\frac{f}{H} \right) = \left(\frac{\partial P_y}{\partial x} - \frac{\partial P_x}{\partial y} \right) + \text{curl}_z \left(\frac{\tau_s - \tau_b}{\rho_o H} \right)$$

Figure 15: Spatial view of time and depth-averaged vorticity budgets in *period I* (a) and in *period II* (b). Shown from left to right are KE storage term (KE change), nonlinear advection work (ADV), pressure work (PW), wind stress work (wind) and dissipation work (dissipation).

Figure 16: Spatial view of of time- and depth-averaged cross-shelf velocity and APV sampled along transect N in *period I* (a) and *period II* (b)

Figure 17: Time series of time- and depth-averaged vorticity term budget along the 100 m isobath (the portions shown by the green line) in January, 2005.

Figure 18. Spatial view of the time-averaged, depth-averaged kinetic energy term budgets in *period I* (a) and *period II* (b)

Figure 19: Time series of the shelf (shallower than 100-m) area averaged depth-integrated kinetic energy term budgets in January, 2005. Term notations are the same as in Figure 18. Also shown is the time series of residual term, which is several order of magnitude smaller than other terms, indicting the KE budget analysis is robust.

Appdenix

Appendix A: Derivation of JEBAR term

(1) First we prove the following pressure gradient term's expression:

$$\begin{aligned}
 \int_{-h}^0 \frac{\partial p}{\partial x} dz &= h \frac{\partial}{\partial x} [p(z=0) + g \int_{-h}^0 \rho dz] + \frac{\partial}{\partial x} (g \int_{-h}^0 \rho z dz) \\
 &= h \frac{\partial}{\partial x} [p(z=0) + g \int_{-h}^0 \rho dz] + \frac{\partial}{\partial x} (g \int_{-h}^0 \rho z dz) - \int_{-h}^0 \frac{\partial p}{\partial x} dz \\
 &= h \frac{\partial}{\partial x} p(z=0) + h \frac{\partial}{\partial x} [g \int_{-h}^0 \frac{\partial p}{\partial z} (-\frac{1}{g}) dz] + \frac{\partial}{\partial x} [g \int_{-h}^0 \frac{\partial p}{\partial z} (-\frac{1}{g}) z dz] - \int_{-h}^0 \frac{\partial p}{\partial x} dz \\
 &= h \frac{\partial}{\partial x} p(z=0) - h \frac{\partial}{\partial x} (\int_{-h}^0 \frac{\partial p}{\partial z} dz) - \frac{\partial}{\partial x} (\int_{-h}^0 \frac{\partial p}{\partial z} z dz) - [\frac{\partial}{\partial x} \int_{-h}^0 p dz + \frac{\partial(-h)}{\partial x} p(z=-h)] \\
 &= h \frac{\partial}{\partial x} p(z=0) - h \frac{\partial}{\partial x} (\int_{-h}^0 \frac{\partial p}{\partial z} dz) - (\frac{\partial}{\partial x} \int_{-h}^0 \frac{\partial p}{\partial z} z dz + \frac{\partial}{\partial x} \int_{-h}^0 p \frac{\partial z}{\partial z} dz) + \frac{\partial(h)}{\partial x} p(z=-h) \\
 &= h \frac{\partial}{\partial x} p(z=0) - h \frac{\partial}{\partial x} (\int_{-h}^0 \frac{\partial p}{\partial z} dz) - \frac{\partial}{\partial x} (\int_{-h}^0 \frac{\partial p z}{\partial z} dz) + \frac{\partial(h)}{\partial x} p(z=-h) \\
 &= h \frac{\partial}{\partial x} p(z=0) - h \frac{\partial}{\partial x} (\int_{-h}^0 \frac{\partial p}{\partial z} dz) + \frac{\partial}{\partial x} [-h p(z=-h)] + \frac{\partial(h)}{\partial x} p(z=-h) \\
 &= h \frac{\partial}{\partial x} p(z=0) - h \frac{\partial}{\partial x} (\int_{-h}^0 \frac{\partial p}{\partial z} dz) - h \frac{\partial}{\partial x} [p(z=-h)] \\
 &= h \frac{\partial}{\partial x} [p(z=0) - \int_{-h}^0 \frac{\partial p}{\partial z} dz - p(z=-h)] \\
 &= h \frac{\partial}{\partial x} \{p(z=0) - [p(z=0) - p(z=-h)] - p(z=-h)\} \\
 &= 0 \\
 \int_{-h}^0 \frac{\partial p}{\partial y} dz &= h \frac{\partial}{\partial y} (p(z=0) + g \int_{-h}^0 \rho dz) + \frac{\partial}{\partial y} (g \int_{-h}^0 \rho z dz)
 \end{aligned}$$

(2) Then we prove that the curl of depth averaged pressure gradient is equivalent to JEBAR:

$$\frac{\partial}{\partial x} (\frac{1}{h} \int_{-h}^0 -\frac{1}{\rho_0} \frac{\partial p}{\partial y} dz) - \frac{\partial}{\partial y} (\frac{1}{h} \int_{-h}^0 -\frac{1}{\rho_0} \frac{\partial p}{\partial x} dz) = J (\int_{-h}^0 \frac{\rho g z}{\rho_0} dz, 1/h)$$

$$\begin{aligned}
& \frac{\partial}{\partial x} \left(\frac{1}{h} \int_{-h}^0 \frac{\partial p}{\partial y} dz \right) - \frac{\partial}{\partial y} \left(\frac{1}{h} \int_{-h}^0 \frac{\partial p}{\partial x} dz \right) \\
&= \frac{\partial}{\partial x} \left[\frac{\partial}{\partial y} (p(z=0) + g \int_{-h}^0 \rho dz) + \frac{1}{h} \frac{\partial}{\partial y} (g \int_{-h}^0 \rho z dz) \right] - \frac{\partial}{\partial y} \left[\frac{\partial}{\partial x} (p(z=0) + g \int_{-h}^0 \rho dz) + \frac{1}{h} \frac{\partial}{\partial x} (g \int_{-h}^0 \rho z dz) \right] \\
&= \frac{\partial}{\partial x} \left[\frac{1}{h} \frac{\partial}{\partial y} (g \int_{-h}^0 \rho z dz) \right] - \frac{\partial}{\partial y} \left[\frac{1}{h} \frac{\partial}{\partial x} (g \int_{-h}^0 \rho z dz) \right] \\
&= \frac{\partial}{\partial x} \left(\frac{1}{h} \right) \frac{\partial}{\partial y} (g \int_{-h}^0 \rho z dz) - \frac{\partial}{\partial y} \left(\frac{1}{h} \right) \frac{\partial}{\partial x} (g \int_{-h}^0 \rho z dz) \\
&= J(1/h, \int_{-h}^0 \rho g z dz)
\end{aligned}$$