Thesis submitted for the degree of Doctor of Philosophy

MODELLING OF THE CASPIAN SEA

JAMES FARLEY NICHOLLS

Space and Atmospheric Physics Group
Department of Physics
Imperial College

January 2013
The work in this thesis is my own and all exceptions are properly referenced.
Abstract

More advanced models of climate systems are needed for use in present day weather forecasting and climate projection, and there is a drive towards the use of coupled modelling of various processes to achieve this goal. This thesis seeks to investigate coupled ocean-atmosphere-wave modelling using the latest generation of models. The test basin for this investigation is the Caspian Sea, where accurate representation of the water budget is vital for prediction of water level changes, which have historically seen trends of up to 15 cm/year.

The individual models of atmosphere, waves and ocean are first run separately to investigate their skill in predicting observed conditions in the Caspian. These models capture the behaviour of the basin when model results are compared with observed wind speeds, currents, wave heights, sea-surface temperatures and precipitation.

The coupling of the ROMS ocean and WRF atmosphere models is seen to improve sea-surface temperature prediction, but, under the Janjic Eta surface layer scheme used here, increases evaporation above the level expected. The additional inclusion of wave coupling from the SWAN model decreases strong winds through wave dependent surface roughness, reduces sea-surface temperatures and increases precipitation; all leading to better agreement with measurements. Wave prediction is best when wave-atmosphere coupling is included, but not current-wave coupling - this is believed to be because of the “double counting” of currents, where they are included both implicitly in the model formulation and then explicitly through coupling.

The final part of this study considers near-inertial oscillations, which are frequently observed in the measured current records. The model is able to accurately represent the observations, and sees significant near-inertial oscillations over most of the basin. The amplitude of the oscillations in the model is found to increase with distance from the coastline. This agrees with the mechanism of barotropic and baroclinic waves, which are generated by the no flow condition at the coast, controlling inertial oscillations.
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Chapter 1

Introduction

More advanced models of environmental systems are needed for present day weather forecasting and climate projection. In order to achieve this goal a complete dynamical model of the whole environment is required. For this reason there is a movement towards coupled modelling of the ocean and atmosphere. Within the field of global climate modelling, coupling between ocean and atmosphere models has been widely implemented, however, the extension to higher spatial resolution and shorter timescales is in its infancy, and is investigated in this thesis. The further inclusion of wave model coupling with the atmosphere has yet to become widespread and one goal of this thesis is to test this coupling. This thesis aims to investigate coupled high-resolution ocean-atmosphere-wave modelling using the Caspian Sea as a test bed. The Caspian Sea is chosen because of the poor state of existing models, while the enclosed nature of the basin means that oceanic boundary conditions are not required and their impact on the model results can then be ignored. Near-inertial oscillations are an important wind-driven process in the oceanic mixed layer and their prevalence and magnitude have previously been characterised across the world’s oceans. However, to date there have been no studies of inertial oscillations on the Caspian Sea, and as such, this thesis takes a modelling approach to characterise their spatial structure on the Caspian and the mechanisms involved.

This section of the thesis introduces the concept of coupled modelling and some relevant studies, followed by an introduction to inertial oscillations and some previous modelling studies, finally important details on the Caspian Sea environment and the factors affecting sea-level change are presented.
1.1 Coupled Modelling

Coupled modelling involves combining models of separate processes. For example, this can involve using an ocean model in combination with a model of the atmosphere. The key is that coupling allows for the exchange of prognostic variables between the separate components, which are run in parallel.

Coupling of individual component models allows for the inclusion of increased model physics, representing coupled processes. While output from one model can be used to force another, this doesn’t allow for the feedback between the respective fields. Therefore by coupling the models these feedbacks can be explicitly represented. In addition to this, the model coupling allows for increased consistency between models rather than applying boundary forcing from another product.

Additionally coupling of models can allow for increased resolution of high frequency forcing fields. It can be computationally expensive to store forcing data at, say, hourly intervals, however, when coupling models, the data can be exchanged internally and so doesn’t need to be stored.

In this section, the physics involved with coupled processes between ocean, atmosphere and waves is discussed. Previous studies of these processes are considered.

1.1.1 Wave-Current Interactions

When considering waves on a moving body of water we need to consider the wave frequency relative to the flow of water ($\sigma$), as distinct from that which can be observed from a fixed point ($\omega$), where that measured is the frequency relative to a fixed point. The frequency can be given for waves on a moving body of water by the relation from Peregrine [1976]:

\[ \sigma = \omega - k:U \]  \hspace{1cm} (1.1)

where $k$ is the wave number and $U$ is the current velocity. From this it is obvious that when waves are travelling in the same direction as a current, the observed frequency ($\omega$) is greater than that relative to the water ($\sigma$). Therefore the relative frequency is greater when no currents are present, compared to when currents are parallel to the waves.

Bretherton and Garrett [1968] showed that when currents are present, wave action density ($N$), rather than energy density ($E$), is conserved. Action den-
Density is given by energy density divided by the relative frequency of the waves, \( N(\sigma, \theta) = \frac{E(\sigma, \theta)}{\sigma} \). This means that if the relative frequency is reduced, then the energy density is similarly decreased, and wave heights will be reduced. Hence when waves travel in the direction of currents, \( \sigma \) is reduced, and so are the wave heights.

When the current varies in space, and waves propagate onto a stronger (weaker) following current, the relative frequency of the waves remains unchanged, while the wave length is increased (reduced).

An additional effect is that if the current follows the wind direction, then the wind velocity relative to the water is decreased, leading to smaller waves. While, when the current follows the wind, the wave speed relative to a fixed point is increased, which, as noted by Vincent [1975], effectively decreases the fetch, and means that waves have less time to grow before reaching land, and so don’t reach the same heights.

The currents near the surface affect the wave field more than those near the bottom. However, given that wave motion penetrates below the surface, currents here also have an impact on the waves. Kirby and Chen [1989] studied the interaction and derived a method for calculating the depth averaged currents which affect the wave field, where the currents are weighted by the wave orbital velocity at that depth. This is shown by Elias et al. [2012] to provide better results, when used in the SWAN wave model, than either top-layer, or simple depth averaged, currents.

Generally wave model equations are expressed in terms of wave action density\(^1\) and so these effects can be implicitly represented. This however requires currents to be provided as forcing for the wave model or for the wave model to be coupled to an ocean circulation model.

\begin{itemize}
\item [1.1.2] Ocean-Atmosphere Interaction
\end{itemize}

When ocean and atmosphere models are run separately, each generally calculates energy fluxes given its own equation set. This means, for example, the atmosphere could receive more energy from evaporation than is lost from the ocean. This lack of consistency between models can be easily resolved through model coupling, where energy fluxes are calculated in one model and used as a boundary condition for the other.

More realistic energy fluxes can also be computed when ocean and atmosphere models are coupled. These bulk fluxes depend on both oceanic and atmospheric

\(^1\)As is the case with the SWAN model used in this study
parameters, and as such having a consistent set of near surface variables, where the ocean affects atmosphere and vice-versa, allows more realistic flux calculation.

Nelson and He [2012] use the coupled ocean-atmosphere model of Warner et al. [2010] to explore the impact of wind convergence on extra-tropical cyclones around the Gulf Stream. They show that the strong air-sea interactions associated with large SST gradients have to be accurately represented to simulate these cyclones. This would not be possible without the feedback between models and the high-resolution SST fields that can be obtained from a regional ocean model.

1.1.3 Wave-Atmosphere Interaction

Within atmospheric modelling, the computation of wind speed profiles near the ocean surface depends on the surface roughness length, which is a measure of the resistance to flow of air over the surface. The surface roughness is also used in the calculation of latent and sensible heat fluxes from the ocean.

It has long been known that waves influence the lower atmosphere through the surface roughness length. Charnock [1955] implicitly included the effect of waves in a widely used surface roughness relation, where roughness length depends on friction velocity squared. The direct impact of waves on surface roughness was then explored by Donelan et al. [1993], who showed roughness decreases with wave age\(^2\), but were unable to determine the relationship. Young seas are generated locally and the wave speed is much less than the wind speed, whilst older waves are more developed and travel more quickly. The wave age can also give an indication of the wave steepness as young seas tend to be choppy while fully developed seas are more regular and have smoother variations in surface elevation.

More recently a number of studies have determined relationships between wave parameters and surface roughness length. Of these, Taylor and Yelland [2001] find that roughness depends on wave steepness (wave height divided by wavelength):

\[ z_0 = 1200h_s \left( \frac{h_s}{\lambda_p} \right)^{4.5} + 0.11 \frac{\nu}{u_s} \]  

(1.2)

where \(h_s\) is the significant wave height\(^3\), \(\lambda_p\) is the wavelength of the peak of the wave

\(^2\)Wave age is defined as friction velocity divided by wave speed (\(\frac{u^*}{C_p}\)), and as the name would suggest is an indicator of how long ago waves were generated.

\(^3\)Significant wave height is the most widely used quantity to describe the height of the wave field. It is defined as the average height of the highest third of waves in a record.
spectrum, $\nu$ is the kinematic viscosity and $u_*$ is the friction velocity. The relation of Oost et al. [2002] depends on wave age:

$$z_0 = \frac{50}{2\pi \lambda_p} \left( \frac{u_*}{C_p} \right)^{4.5} + 0.11 \frac{\nu}{u_*}$$  

(1.3)

where $C_p$ is the phase speed of the peak of the wave spectrum.

The relationship between waves and surface roughness tends to mean that stronger winds lead to an increase in roughness and hence a feedback to reduction of surface wind speeds. However, studies of extreme winds, particularly in hurricanes, have shown that surface roughness might actually decrease with increasing wind speed beyond a certain point (e.g. Powell et al., 2003). This is believed to be because, at extreme winds wave tops become sheared off, creating sea-spray, which can then decrease roughness and lead to ‘skipping’, where an air packet passes from wave-peak to wave-peak avoiding the troughs.

As well as having an impact on surface roughness length, sea-spray from breaking waves also directly impacts energy fluxes at the ocean surface (e.g. Andreas et al., 1995). Spray throws water droplets into the air, which therefore increases the total water surface area, with an associated increase in evaporation. Equally, if spray is thrown into the air it might escape the saturated layer directly above the ocean, again increasing evaporation. Andreas and Emanuel [2001] suggests that sea-spray effects might account for an increase in energy transfer of 4% and be even more significant at wind speeds above 20 m/s.

Therefore, through these processes, waves have an effect on the near-surface wind field and surface bulk energy fluxes which it is not possible to represent in models without coupling of waves and atmosphere. In addition, the feedback effect of wave dependent roughness on the wave field, through its effect on the surface winds, can only be accurately modelled through coupling.

### 1.1.4 Fully Coupled Modelling

A number of studies with models involving coupling of atmosphere, ocean and waves have been reported in the literature.

The importance of dynamically consistent momentum fluxes is noted in Fan et al. [2009], where coupled wind-wave-current dynamics are studied during a tropical cyclone using an air-sea momentum exchange model. Generally momentum fluxes
are assumed to be independent of sea-state, however it is shown through the model of Fan et al. [2009] that momentum flux into the ocean currents could be significantly less than that from the atmosphere when waves are growing, and hence extracting momentum. This is because the wave field stores momentum, and therefore not all of that lost by the atmosphere is transferred to ocean currents immediately.

Warner et al. [2010] use their coupled model (see Section 2.5) to investigate the development and passage of a hurricane. They show that hurricane intensity is very sensitive to sea-surface temperature, and coupling of atmosphere and ocean reduced intensity and SSTs: better in line with measurements. Inclusion of wave coupling reduces wind speeds by increasing surface roughness above that calculated from the Charnock relation, further reducing hurricane intensity below that observed.

Liu et al. [2011] developed a coupled atmosphere-ocean-wave model using WRF, POM and SWAN. This was the first modelling study to consider the effects of both wave state and sea spray on the atmosphere-wave interaction; and was later extended by Liu et al. [2012]. Under idealised tropical cyclone conditions, the authors find that wave-atmosphere coupling strengthens the tropical cyclone system through spray decreasing roughness and increasing heat flux to the atmosphere; while ocean-atmosphere coupling reduces intensity through cooler SSTs leading to a decreased heat flux from the ocean.

Both Liu et al. [2011] and Warner et al. [2010] find ocean-atmosphere interactions reduce storm intensity. However they find contrasting results when the wave-atmosphere interaction is considered, where Liu et al. [2011] find an increase in intensity and Warner et al. [2010] find a decrease. This difference is directly attributable to the inclusion of sea-spray processes which increase ocean to atmosphere energy fluxes and decrease surface drag, both of which increase storm intensity. There is however agreement between the models that without sea-spray, waves lead to an increase in roughness and an associated decrease in near surface wind speeds.

All of the literature on coupled atmosphere-wave-ocean modelling is focused on extreme events such as hurricanes and tropical cyclones. While the coupled processes are likely to have the largest impact on behaviour in extreme conditions, the effect on mean conditions is understudied. For this reason one of the main goals of this study is to investigate the impact of coupled processes over the course of a longer simulation for the Caspian Sea.
1.2 Inertial Oscillations

Near-inertial waves are the most energetic form of internal waves (Kunze, 1985), comprising up to half of the near surface kinetic energy (Pollard and Millard, 1970). These waves are considered to have a vital role in mixing at the base of the mixed layer (D’Asaro, 1985) and in maintaining meridional overturning circulation (Munk and Wunsch, 1998). Despite the obvious importance of inertial oscillations, there have to date been no studies of their impact or presence on the Caspian Sea.

Inertial oscillations (or inertial waves) are caused by the Coriolis acceleration acting to restore geostrophic equilibrium, after a perturbation at the surface. Once a flow has been driven at the surface of the ocean by a wind impulse, it is subject to a Coriolis acceleration perpendicular to the direction of the flow. This results in the flow deviating to the right of its initial path (in the Northern Hemisphere). Again this flow is subject to a Coriolis acceleration, and deviation, to the right. The result is a circular flow, the sense of which is clockwise in the Northern Hemisphere and anti-clockwise in the Southern Hemisphere. This can therefore be identified by oscillations in the east-west ($u$) and north-south ($v$) current components, with the $v$ current leading by a quarter of a period in the Northern Hemisphere. The waves travel through the ocean at the local Coriolis frequency ($f$).

The exact frequency of the propagating wave can differ from the local Coriolis frequency. This can be because waves created at a certain latitude, with the expected frequency, propagate north or south to a latitude whose local Coriolis frequency is different from that of the wave (e.g. Garrett, 2001; Alford, 2003; Elipot and Lumpkin, 2008). Kunze [1985] notes that waves are actually generated at an effective frequency which is not equal to the local Coriolis frequency, but rather $f_{\text{eff}} = f + \zeta/2$, where $\zeta$ is the vorticity; an effect seen in the global study of Elipot et al. [2010]. Because both of these effects can lead to waves at a frequency slightly different to the Coriolis frequency, they are generally referred to as near-inertial oscillations (NIOs).

Near-inertial oscillations are forced by wind events, and the duration of the wind event is a key factor in the magnitude of the response. Pollard and Millard [1970] say that for unidirectional wind events lasting longer than half an inertial period, there is destructive interference, leading to a decrease in inertial amplitudes, therefore the largest inertial responses occur after short, high wind, events. Mortimer [2006] compares the inertial response due to wind events of similar magnitude, but different duration, and found that the shorter wind events led to larger NIOs. However, as shown in Dohan and Davis [2011], when winds rotate in phase with an inertial
current, at the same frequency, a stronger NIO event can be generated through resonance.

As inertial waves are forced by high frequency winds, Klein et al. [2004] uses a 1-D model to show the importance of using high temporal resolution wind forcing for the simulated inertial response. Inertial energy is reduced by a factor of 7 when daily wind forcing is applied when compared to 3 hourly forcing. Given this, we note the importance of having sufficiently high resolution winds for any study aiming to accurately model NIOs.

Because inertial oscillations are wind forced, they are strongest near the ocean surface in the mixed layer. It can generally be assumed (e.g. Pollard and Millard, 1970) that the inertial energy input from the wind is evenly distributed across the surface mixed layer, therefore, for a given wind forcing, the inertial amplitudes will be greatest for shallower mixed layers. This means that inertial oscillations tend to be stronger during the summer (when shallow mixed layers dominate) than in the well mixed winter months. Chaigneau et al. [2008] see amplitudes around 20% higher in the summer and autumn than winter, Park et al. [2005] find in a similar study that inertial amplitudes are between 15 and 30% higher in the summer than winter (depending on the ocean), with higher values still in the early autumn months.

There have been a number of studies of the global distribution of near-inertial waves from measurements (e.g., Park et al., 2005; Chaigneau et al., 2008; Elipot and Lumpkin, 2008). These have shown a large variability in the spatial structure and importance of these motions. Chaigneau et al. [2008] conduct a study of NIOs by examining the tracks of 8500 near-surface drifters between July 1999 and December 2006. They find that the global mean magnitude of near-inertial currents is 10 cm/s at 15 m depth, while local mean amplitudes vary from a few cm/s to nearly 25 cm/s. To date, however, there have been no studies of NIOs on the Caspian Sea, which is not included in the global studies.

### 1.2.1 Previous Modelling Studies of NIOs

Over the last few years realistic three-dimensional models have begun to be used to study near-inertial oscillations.

Furiuci et al. [2008] use the Princeton Ocean Model (POM) with 6 hourly reanalysis wind forcing to study the global distribution of wind energy input to near-inertial motions. They find that the total energy available for deep-ocean mixing is an order of magnitude less than previously assumed. POM is also used in
the study of Jordi and Wang [2008], who use high temporal-resolution atmospheric forcing, from a limited area model, to study NIO response to a storm in the western Mediterranean. While the model underestimates the amplitude of NIOs, the general behaviour is in good agreement with the measurements.

A coupled global ocean-atmosphere model is used by Komori et al. [2008] to study deep ocean vertical motions associated with inertial waves. Gierich et al. [2009] notice an inertial response to the passage of Hurricane Katrina in their high resolution HYCOM simulation.

ROMS has previously been used to study near-inertial oscillations by Zhang et al. [2010]. They conduct two idealised experiments of the effect of sea-breeze forcing in the Gulf of Mexico where the Coriolis period is 24 hours. A resonance occurs between NIOs and the sea-breeze, leading to a much stronger response than at other latitudes. A final experiment with a realistic setup of ROMS is performed using analytical wind and surface heat fluxes, where qualitative agreement is found with observations of wave propagation, vertical mixing and NIO magnitude.

These models have been shown to capture inertial behaviour, and so can be used to investigate the underlying dynamics. The ROMS model has until now not been used to study NIOs in a fully realistic simulation.

\section{1.3 Caspian Sea}

The Caspian Sea is the largest enclosed body of water in the world and is situated between Russia, Azerbaijan, Kazakhstan, Iran and Turkmenistan. The Caspian runs north-south and extends approximately 1000km from 37 to 47°N and, at its widest, 600km from 47 to 54°E. To the west lie the Caucasus mountains, to the south are the smaller Elburz mountains and to the east is largely desert (Figure 1.1).

The inflows of water into the Caspian are rainfall over the Sea, groundwater flow and river runoff. According to Klige and Myagkov [1992] the inflow is 79\% river runoff, 20\% rainfall and 1\% groundwater inflow. The Caspian is an enclosed basin and thus outflow is almost exclusively by evaporation from the surface (97\%), but is supplemented by a small flow into the Kara Bogaz Gol (KBG) which is a smaller, shallow, lagoon connected by a narrow channel to the Caspian and whose outflow is solely evaporative. River inflow is dominated by the Volga river which enters in the north of the basin and accounts for about 80\% of the runoff (Rodionov, 1994). The Ural, Kura, Terek and Emba rivers contribute almost all of the additional 20\% and flow into the north and west of the basin.
The Caspian Sea can be considered in three distinct basins. The North is very shallow with depths not exceeding 20 meters, the Central Caspian is much deeper with depths of up to 800 meters, while the South Caspian is deeper again (up to 1000m) (see Figure 2.4) and accounts for around 66% of the total water volume of the basin (Rodionov, 1994). The South and Central Caspian are separated by a sill, with depths up to about 180 m, running across the basin from the Aspheron peninsula.

The salinity of the Caspian Sea, at around 13 ppm, is approximately a third of that in world’s oceans. The salinity can be much lower than this in the North Caspian where the fresh water inflow from the Volga river dominates.

The Caspian is non-tidal, the circulation pattern is generally cyclonic and was characterised by Lednev [1943] (Figure 1.2). In winter, the pattern of southward currents along the west coast, and northward currents along the east coast, of the Central and South Caspian, leads to cold sea-surface temperatures (SSTs) along the west coast and warmer water on the east (Rodionov, 1994). In summer, wind-induced upwelling on the east coast of the Central Caspian leads to a region of cold SSTs (Rodionov, 1994 and Ibrayev et al., 2009). Ibrayev et al. [2009] find that evaporation along the east coast of the Central Caspian is high in winter when SSTs are high, and cold dry air from the east intrudes over the water; and low in summer due to the cold SSTs. They also find that evaporation in the North Caspian in
summer is almost twice that in the deeper basins.

During the winter months, the North Caspian and the KBG often completely freeze over. Rodionov [1994] notes that freezing begins in the north-east of the Caspian from mid-November and that sea-ice disappears by late-March, early-April.

Winds over the Caspian Sea are mostly northerly, occurring 41% of the time, and more frequently than this in summer (Rodionov, 1994).

1.3.1 Caspian Sea Level Change

As the Caspian is enclosed and has no in/outflow from/to the global oceans, its water level is very sensitive to changes in the water budget. Over the last century, the Caspian Sea level (CSL) has undergone dramatic changes, of up to 15 cm/year (Klige, 1992), which can be seen in Figure 1.3. More recent CSL trends, and the seasonal cycle, can be seen from monthly average sea levels as shown in Figure 1.4, where the impact of the recent droughts over Russia on the water level can be seen in the drop in CSL from 2010 onwards. The largest historical trends in CSL are the dramatic drop of around 20 cm/year between 1933-1940 and then a rise of around 13 cm/year from 1977. The drop in the 1930’s is attributed mainly to an unusually dry climate in the drainage basin resulting in decreased runoff to the Caspian, but was
exacerbated by higher than average evaporation (Rodionov, 1994). The recovery in CSL from 1977 is mainly explained by decreased evaporation, believed to be caused by a change in atmospheric circulation (Panin and Dzyuba, 2003), but also relied upon increased river runoff.

The change in CSL can be expressed as:

\[
\Delta CSL = R + P + G - E - F_{KBG}
\]

where \( R \) and \( G \) are the river runoff and groundwater flow into the Caspian, \( P \) and \( E \) are the precipitation and evaporation over the Caspian Sea, and \( F_{KBG} \) is the flow into the KBG (here \( R, G \) and \( F_{KBG} \) are normalised by the area of the Caspian and so are in terms of contribution to water level). The flow into the KBG is dependent on the CSL, but tends to be small, and of the order of a few cm/year. The volume of groundwater flow is uncertain, although this is believed to contribute no more than 1 cm/year. Given that the precipitation, evaporation and runoff terms are of the order of 100’s mm/year, these are the controlling factors in the balance.
For this reason modelled evaporation from the Caspian, and precipitation over the surrounding area, are considered in Chapters 3 and 4.

*Elguindi and Giorgi* [2006] use a hydrological balance equation to calculate CSL changes, using an ensemble of general circulation models (GCMs) to provide estimates of evaporation and precipitation rates over the sea and the drainage basin. Their model includes a term representing the fraction of precipitation over land which is lost by processes other than soil evaporation before it reaches the Caspian as runoff. This factor is used to tune the model to account for biases in precipitation and evaporation of each individual GCM and thus is not physically based. The broad trends in CSL change over the 20th century are represented by this model although the rapid variations are underestimated.

Forecasts of the GCMs performed for the A1b emission scenario from the Intergovernmental Panel on Climate Change (IPCC) were then used to provide projections of future CSL change. 6 of the 7 models predict a decline in CSL by 2100. Two of the models have widely diverging predictions (of +5 m and -20 m) while the spread of the others is less than 5 m with the ensemble average projecting a drop of 7 m over the 21st century.

*Golitsyn* [1995] studies the ability of a number of GCMs to simulate the water balance of the Caspian. They find that the higher resolution models performed best
and as such concluded that the balance could not be properly simulated without sufficient representation of the sea and surrounding areas. For this reason a move towards using regional climate models to study CSL change might be expected to provide better results.

1.3.2 Rainfall and Evaporation

There are rarely any measurements of rainfall over the Caspian Sea itself. The literature contains a number of different estimates of the mean total precipitation over the Caspian Sea. Rodionov [1994] gives a value of 195 mm/year over the period of 1900-1990, while yearly totals vary between 110 and 310 mm/year. Ibrayev et al. [2009] quotes three values from other studies (all published in Russian) of 243, 257 and 230 mm/year for the periods of 1970-1977, 1978-1982 and 1978-1990 respectively.

The recent availability of satellite observations allows the spatial structure of precipitation to be observed. For instance Figure 1.5 shows the mean climatology of precipitation from a combined product from the TRMM and GPCC datasets (see Section 2.6.3). From this, the total precipitation over the Caspian is 275 mm/year for the years 1999-2008, which is close to the values in Ibrayev et al. [2009] but higher than that in Rodionov [1994]. From the climatology we can see that the largest amount of precipitation falls over the Caucasus mountain range and over the south-western corner of the Caspian Sea. Given that the most intense precipitation over the Caspian Sea falls near to the coast, the exact location of this rainfall is of great importance for the Caspian’s water budget. A slight shift in the location of the precipitation might lead to the water falling over land. While this water should eventually reach the Caspian (as the drainage basin extends slightly to the south of the Sea) some will be either being evaporated or stored as groundwater, and therefore not reach the Caspian (at least immediately). This means that to get a good idea of the contribution to the water budget of rainfall in this location, a high resolution rainfall product, which accurately resolves the area and the precipitation, is required.

Evaporation can be estimated from humidity, water temperature and wind speed using bulk formulae, however it is more frequently calculated as a residual from the known trend in Caspian Sea level. When the CSL change is combined with measurements of the Volga discharge and rainfall, a value of the total evaporation can be inferred. Elguindi and Giorgi [2006] note that there are no reliable observations
of evaporation over the Caspian drainage basin, while Rodionov [1994] estimate the error in evaporation estimates at 30%.

The literature contains a range of values for the total evaporation over the Caspian. Rodionov [1994] calculates a mean value of 972 mm/year as the residual in the water balance over the period of 1900-1990, with annual totals from 790-1230 mm/year. Ibrayev et al. [2009] quotes values of 1039, 979 and 918 mm/yr.

The controlling factors for change in the CSL are the balance between evaporation and precipitation, which is noted as \( E - P \), and river inflow. \( E - P \) is the net amount of water lost from the Caspian to the atmosphere. Values of \( E - P \) can be calculated as the residual of CSL change and total runoff (assuming transport to the KBG and groundwater inflow are negligible). The CSL change can be precisely known from the many water level measurements. The river discharge for the major rivers is recorded, is seen to have a standard deviation of about 15% of the total and, according to Rodionov [1994], is known to 3-4% accuracy, although this seems optimistic. This means that \( E - P \) can be known with much more accuracy and ease than either evaporation or precipitation. \( E - P \) values in the literature range from 688 mm/year (Elguindi et al., 2011) to 796 mm/year (Ibrayev et al., 2009).
1.3.3 Models of the Caspian Sea

Three-dimensional models incorporating measured data covering the world’s oceans (e.g. HYCOM and SODA) are widely available. These products can either be used directly to study ocean circulation, or as boundary forcing for a regional ocean model. However, these are not available for the Caspian Sea as it is a separate, enclosed, basin so tends not to be included in global models. This makes studying the Caspian Sea less accessible than other oceans, and so it remains an understudied region.

The same applies to other large lakes such as the Great Lakes in North America, however these have been the focus of extensive research over the last 20 or so years, meaning that the Caspian has been left behind as probably the largest, most understudied body of water on the planet.

Ibrayev et al. [2009] use a three-dimensional primitive equation model to study the Caspian Sea. Atmospheric forcing is monthly mean re-analysis data, and repeats the same year of forcing perpetually. The focus is on the seasonal circulation and heat and moisture fluxes, and the model is able to represent the mean climatology of these processes. The model however does not seek to represent the shorter timescale processes involved in the dynamics of the Caspian Sea, and this remains largely unstudied.

1.4 Thesis Goals and Layout

The major aim of this thesis is to use advanced regional models of the atmosphere, ocean and waves to study the Caspian Sea. Within this goal, the coupling of these regional models is investigated on smaller scales and at higher resolution than widely employed in coupled climate models, and over longer simulation periods than previously tested. Further to this, an additional goal is to perform the first characterisation of near-inertial oscillations on the Caspian.

The Caspian Sea level has shown dramatic historical changes and is a sensitive balance between the net water loss from evaporation minus precipitation and river runoff, as such this thesis aims to investigate the model skill in representing this balance. These changes have resulted in huge human impact and so understanding the ability to model the key factors could be of great benefit. In order to better represent the balance of evaporation and precipitation, different planetary boundary layer and microphysics schemes in the atmospheric model are tested to find the
combination with the best skill. It is then investigated whether the coupling with an ocean model can improve the representation of evaporation and precipitation. By inclusion of high-resolution sea-surface temperature fields from the ocean model, and the feedback processes on temperatures, the coupling may improve the bulk flux representation which controls evaporation, and hence could give better water budget prediction. This is investigated by comparison with observed precipitation totals and evaporation estimates.

Presently there are no advanced models of the ocean or waves in the Caspian. These models are required by the oil and gas industry, among others, in particular for calculations of stresses on platforms. Here, models of waves and ocean are presented and validated against measurements. The impact of coupling, between these models, and with an atmospheric model, is then investigated. Particular emphasis is placed upon the coupling of wave and atmosphere models, which has not been widely applied elsewhere.

A further goal of this thesis is to study near-inertial oscillations on the Caspian Sea. The oscillations are an important component of the surface currents, and as such they need to be represented and captured in modelling the ocean. While there have been a number of studies of the distribution of NIOs over the global oceans, the Caspian Sea, as an enclosed body, has previously been ignored. For that reason, the spatial pattern of oscillations is characterised and the drivers of the distribution of inertial amplitudes are investigated.

The second chapter introduces the models used in this study along with the setups used on the Caspian region and the major datasets used for validation. Model results from year-long simulations are presented for the atmosphere, ocean and wave models individually in Chapter 3. The results of the coupling of these models are shown in Chapter 4, where particular emphasis is placed upon the impact of coupled processes on the results. The final results chapter discusses near-inertial oscillations, where modelled oscillations are compared with measurements, and the spatial structure over the basin and its causes are investigated. In Chapter 6 the major conclusions and future work are presented.
Chapter 2

Models and Data

2.1 Regional Modelling

Regional Circulation Models (RCMs) are frequently used to dynamically downscale fields from General Circulation Models (GCMs). By reducing the size of the area studied, RCMs can be run with higher resolution than GCMs, allowing more processes to be explicitly resolved. Those sub-gridscale processes which cannot be resolved are parameterised.

This study focuses on the use of regional models of the atmosphere, ocean and waves. The models used in this work and the setup, of each, is described in this chapter. Finally, a coupled model, allowing concurrent simulations, with exchanges of variables, is also presented.

2.2 WRF Atmosphere Model

This study uses the Advanced Research WRF (ARW) version of the Weather Research and Forecasting (WRF) model (Skamarock et al., 2008), which was developed for high-resolution applications (Done et al., 2004), and is intended for both research and operational use.

WRF uses a terrain following vertical coordinate system near the surface, which merges to follow pressure levels near the top of the atmosphere. The model solves the fully compressible, non-hydrostatic Euler equations and contains numerous options of surface layer, microphysics, radiation schemes and others.

The momentum equations in WRF are written as:
\[ F_U = \frac{\partial U}{\partial t} + m_x \left[ \frac{\partial}{\partial x} (Uu) + \frac{\partial}{\partial y} (Vu) \right] + \frac{\partial}{\partial \eta} (\Omega u) + \frac{m_x}{m_y \alpha_d} \left[ \mu_d \left( \frac{\partial \phi'}{\partial x} + \alpha_d \frac{\partial \psi'}{\partial x} \right) + \frac{\partial \phi}{\partial x} \left( \frac{\partial \psi'}{\partial \eta} - \mu_d' \right) \right] \]  

(2.1)

\[ F_V = \frac{\partial V}{\partial t} + m_y \left[ \frac{\partial}{\partial x} (Uv) + \frac{\partial}{\partial y} (Vv) \right] + \frac{m_y}{m_x \alpha_d} \left[ \mu_d \left( \frac{\partial \phi'}{\partial y} + \alpha_d \frac{\partial \psi'}{\partial y} \right) + \frac{\partial \phi}{\partial y} \left( \frac{\partial \psi'}{\partial \eta} - \mu_d' \right) \right] \]  

(2.2)

\[ F_W = \frac{\partial W}{\partial t} + \frac{m_x m_y}{m_y} \left[ \frac{\partial (Uw)}{\partial x} + \frac{\partial (Vw)}{\partial y} \right] + \frac{\partial (\Omega w)}{\partial \eta} - m_y^{-1} g \alpha \left[ \frac{\partial \psi'}{\partial \eta} - \mu_d' \right] \]  

(2.3)

with the mass conservation equation:

\[ \frac{\partial \mu_d'}{\partial t} + m_x m_y \left[ \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} \right] + m_y \frac{\partial \Omega}{\partial \eta} = 0 \]  

(2.4)

and geopotential equation:

\[ \frac{\partial \phi'}{\partial t} + \mu_d^{-1} \left[ m_x m_y \left( U \frac{\partial \phi}{\partial x} + V \frac{\partial \phi}{\partial y} \right) - m_y \Omega \frac{\partial \phi}{\partial \eta} \right] = 0 \]  

(2.5)

where the hydrostatic relation is:

\[ \frac{\partial \phi'}{\partial \eta} = -\mu_d \alpha_d' - \bar{\alpha_d} \mu_d' \]  

(2.6)

The terms \( F_{U,V,W} \) are forcing terms from model physics, mixing and Coriolis acceleration. In these equations the momentum variables in the horizontal \((U,V)\) and vertical \((\Omega)\) directions are written as \( U = \mu_d u / m_y \), \( V = \mu_d v / m_x \), \( \Omega = \mu_d \eta / m_y \), where \( u \), \( v \) and \( \eta \) are the covariant horizontal and vertical velocities. The map scale factors \( m_x \) and \( m_y \) are defined as the ratio of distance in computational space to
the distance on the Earth’s surface. In WRF, variables are expressed as a perturbation ($p'$) from the hydrostatically balanced reference state ($\overline{p(z)}$). This notation is used for pressure $p$, geopotential $\phi$, inverse density $\alpha$, dry air mass $\mu_d$. The mixing ratios of water vapour, cloud and rain are given by $q_v$, $q_c$ and $q_r$, the gravitational acceleration, $g$, and the inverse density of dry air is $\mu_d$.

### 2.2.1 Physics Schemes

Given the number of different options for various physics schemes available within WRF, it is not feasible to test each combination. In this study investigation is restricted to two planetary boundary layer, and two microphysics, schemes. This is because the emphasis in this study is on surface wind fields, evaporation and precipitation, which are controlled by these modules.

The microphysics schemes represent processes involved with water vapour, clouds and precipitation. The main differences between the available schemes are the number of hydrometeors included and whether the scheme is single or double moment. The possible hydrometeors represented are water vapour, cloud water, rain, cloud ice, snow and graupel, where many schemes don’t include graupel and some don’t include any in the solid phase. It is recommended by Skamarock et al. [2008] that for high resolution applications (of greater than 10 km) a scheme representing mixed-phase interactions should be included. While single-moment schemes calculate a mixing ratio of the variables, double-moment schemes additionally model individual concentrations.

The Thompson et al. [2008] microphysics scheme is tested. This uses 5 hydrometeors (cloud water, rain, cloud ice, snow and graupel), and is a single-moment scheme, except for cloud ice, where number concentration is explicitly predicted. This scheme was written from scratch rather than as an incremental update to an existing scheme, and as such is designed to act like double-moment schemes but with increased efficiency through the use of lookup tables.

The WRF Single Moment 6-class (WSM6) scheme (Hong and Lim, 2006) adds graupel to the underlying WSM3/5 schemes which include water vapour, cloud water, rain, cloud ice and snow.

Jankov et al. [2009] looks at the difference between microphysics schemes for a number of events over the California coast. It is noted that the concentrations of each hydrometeor varies widely between schemes, even when the total cloud matter is similar. They also find that rainfall is overestimated with all schemes, and that the
WSM6 has more precipitation than the Thompson scheme. The study of Otkin and Greenwald [2008] investigates a number of microphysics and planetary boundary layer schemes within WRF, where they find that cloud cover through the whole column is dependent on the boundary layer scheme as well as the microphysics.

Planetary boundary layer (PBL) schemes represent eddies, which are responsible for the transport of heat, momentum and moisture within the well mixed near surface layer. This layer is directly influenced by the Earth’s surface and varies greatly in height from around 50 m to a few kilometers. The PBL schemes are often tied to a particular surface layer scheme which acts in a much thinner (∼50 m) layer at the surface. This calculates friction velocities and exchange coefficients which allow the computation of fluxes of heat and moisture from the surface into the atmosphere. These fluxes are calculated in the land surface model (LSM) over land, and directly in the surface layer scheme over water. The most obvious difference between PBL schemes is whether the closure is local or non-local. Local closure schemes estimate turbulent fluxes at each point from the local variables and their gradients, whereas non-local closures include non-local terms which can either be parameterised or handled explicitly. Local closures are least appropriate for convective situations when fluxes are dominated by large eddies, which involve transport over longer distances.

The Mellor-Yamada-Janjic (MYJ) planetary boundary layer scheme (Janjic, 1994) is a one-dimensional turbulent kinetic energy, Eta scheme which is used operationally by NCEP including a local closure. This scheme is tied to the Monin-Obukhov (Janjic Eta) surface layer scheme (Janjic, 2002) based on the similarity theory of Monin and Obukhov [1954].

The Yonsei University (YSU) planetary boundary layer (Hong et al., 2006) is tied to the MM5 Monin-Obukhov similarity scheme adapted from the MM5 mesoscale model of Pennsylvania State University (PSU) and National Centre for Atmospheric Research (NCAR). The scheme uses a parameterised non-local closure scheme.

Hu et al. [2010] investigates two PBL schemes in WRF and finds that YSU is better able to represent surface temperature and humidity. It is found that the increased vertical mixing in the YSU scheme leads to a hotter, drier, boundary layer, which is in closer in agreement with the observations over the southern United States.
2.2.2 Bulk Fluxes

As mentioned previously, the fluxes of momentum and energy are calculated in the surface layer scheme in WRF. Both of the surface layer schemes employed in this study are based on Monin-Obukhov similarity theory (Monin and Obukhov, 1954), which is valid over land or water, and uses the relation for latent heat flux:

\[ E_q = L_e \rho M C_q U (q_s - q_a) \]  

where \( L_e \) is the latent heat of evaporation, \( \rho \) is the density of air, \( M \) is the moisture availability (between 0 and 1), \( U \) is the wind speed and \( q_s \) and \( q_a \) are the specific humidity of the surface and air respectively. The term \( C_q \) is the bulk transfer coefficient given by:

\[ C_q = \frac{\kappa^2}{\left[ \ln \left( \frac{z}{z_0} \right) - \psi \left( \frac{z}{L} \right) \right]^2} \]  

(2.8)

where \( \kappa \) is the von Karman constant and \( \psi(z/L) \) is a stability function, which depends on the Obukhov length \( L \).

These schemes use Kansas-type stability functions, which were formulated over land and with a limited range of atmospheric stabilities. The applicability to freely convective situations over water is therefore less certain.

2.2.3 Model Setup

WRF requires three-dimensional boundary forcing of temperature, relative humidity, geopotential height and horizontal winds; as well as 2-d forcing of surface pressure, skin temperature, surface temperature, sea-surface temperature, relative humidity at the surface and surface winds. This boundary forcing and model initialisation is taken from 6-hourly ERA-Interim reanalysis fields (see Section 2.2.4).

The setup used throughout this study includes three one-way nested grids of 36, 12 and 4 km horizontal resolutions, with 38 vertical levels on each. The outer domain extends out west from the Caspian area to the Eastern Mediterranean in order to capture the prevailing atmospheric conditions, while the inner grid covers the entirety of the Caspian Sea area (Figure 2.1). The simulations are performed with a time-step of 24 s for the inner domain, and a factor of 3 longer step for each progressively lower resolution grid.

At a resolution of 4 km, cumulus precipitation is permitted within the model.
(e.g. Skamarock et al., 2008) and as such no cumulus scheme is employed on the inner grid. The Noah land surface model (LSM) is selected, as described in Chen and Dudhia [2001]. The Rapid Radiative Transfer Model (RRTM) and Dudhia (Dudhia, 1989) schemes are used for longwave and shortwave radiation respectively.

Three different setups of WRF are tested in the first part of this study, using combinations of the two surface layer, and two microphysics, schemes described earlier in this section. These schemes represent an example of both local and non-local closure in the PBL; and single and double moment microphysics. Every other aspect of the WRF setup is left unchanged between simulations.

2.2.4 ERA Interim

ERA-Interim is a re-analysis dataset produced by the European Centre for Medium-Range Weather Forecasting (ECMWF) (Dee et al., 2011).

Re-analysis data fills the need for a set of observations of conditions over the whole of the Earth’s atmosphere. Clearly there is no such observational dataset and as such the closest equivalent is to use a combination of a model with observations. Re-analysis uses data-assimilation within a short period forecast to provide a best estimate of the state of the atmosphere at any given point in time.
Many satellite and in situ observations are included in the Interim re-analysis. For instance, satellite measurements of surface wind fields from Quikscat as well as ERS (European Remote Sensing Satellite) are used. This of course means that the wind speeds in ERA-Interim should be expected to correspond closely with those in the Quikscat data. Sea-surface temperatures are required to force the model component of ERA-Interim and these are obtained from the NCEP (National Centers for Environmental Prediction) Real-Time Global sea-surface temperature database for the period considered in this study. This is a 0.5° gridded product for daily SSTs retrieved from an AVHRR satellite combined with in situ observations.

ERA-Interim has a resolution of 1.5° over the whole of the Earth’s surface, which equates to around 150 km in the Caspian region. Atmospheric fields are available at 6 hour intervals on 37 pressure levels on this grid.

2.3 ROMS Ocean Model

The Regional Ocean Modeling System (ROMS) is a primitive equation model with a free-surface and terrain-following vertical s-coordinate (Haidvogel et al., 2000; Shchepetkin and McWilliams, 2005). ROMS is one of a number of three-dimensional models to solve the Reynolds-averaged Navier-Stokes equations using the hydrostatic and Boussinesq assumptions. The hydrostatic momentum equations are solved using a split-explicit time-stepping routine, where a number of barotropic time-steps occur for every baroclinic step. This means that the vertically integrated momentum is calculated on a shorter time-step than vertical component of the flow.

The momentum equations of ROMS are:

\[
\frac{\partial (H_z u)}{\partial t} + \frac{\partial (u H_z u)}{\partial x} + \frac{\partial (v H_z u)}{\partial y} + \frac{\partial (\Omega H_z u)}{\partial s} - f H_z v = - \frac{H_z}{\rho_0} \frac{\partial p}{\partial x} - H_z g \frac{\partial \eta}{\partial x} - \frac{\partial}{\partial s} \left( \frac{\partial^2 u'}{\partial H_z \partial s} - \frac{v}{H_z} \frac{\partial u}{\partial s} \right) - \frac{\partial (H_z S_{xx})}{\partial x} + \frac{\partial (H_z S_{xy})}{\partial y} + \frac{\partial S_{px}}{\partial s} \tag{2.9}
\]

\[
\frac{\partial (H_z v)}{\partial t} + \frac{\partial (u H_z v)}{\partial x} + \frac{\partial (v H_z v)}{\partial y} + \frac{\partial (\Omega H_z v)}{\partial s} + f H_z u = - \frac{H_z}{\rho_0} \frac{\partial p}{\partial y} - H_z g \frac{\partial \eta}{\partial y} - \frac{\partial}{\partial s} \left( \frac{\partial^2 v'}{\partial H_z \partial s} - \frac{v'}{H_z} \frac{\partial v}{\partial s} \right) - \frac{\partial (H_z S_{yx})}{\partial x} + \frac{\partial (H_z S_{yy})}{\partial y} + \frac{\partial S_{py}}{\partial s} \tag{2.10}
\]
\[ 0 = -\frac{1}{\rho_0} \frac{\partial p}{\partial s} - \frac{g}{\rho_0} H_z \rho \]  

(2.11)

with continuity equation:

\[ \frac{\partial \eta}{\partial t} + \frac{\partial (H_z u)}{\partial x} + \frac{\partial (H_z v)}{\partial y} + \frac{\partial (H_z \Omega)}{\partial s} = 0 \]  

(2.12)

and scalar transport:

\[ \frac{\partial (H_z C)}{\partial t} + \frac{\partial (u H_z C)}{\partial x} + \frac{\partial (v H_z C)}{\partial y} + \frac{\partial (\Omega H_z C)}{\partial s} = -\frac{\partial}{\partial s} \left( \rho' w' - \frac{v_0}{H_z} \frac{\partial C}{\partial s} \right) + C_{source}. \]  

(2.13)

Here, \( u, v \) and \( \Omega \) are the velocities in the horizontal \((x, y)\) and vertical \((s)\) directions; \( \eta \) is the free-surface elevation; \( H_z \) is the vertical stretching factor; \( f \) is the Coriolis parameter. \( S_{xx} \) etc. are radiation stress terms; \( \rho \) is the density of sea-water; \( p \) is the pressure, \( v \) and \( v_0 \) are the molecular viscosity and diffusivity; \( C \) represents any tracer quantity; \( C_{source} \) are sinks and sources of the tracers; while an overbar represents time-average and prime indicates turbulent fluctuation. The density relation is closed by a function \( \rho = f(C, p) \). These equations are closed by parameterising the Reynolds stresses and turbulent tracer fluxes as:

\[ \overline{u'w'} = -K_M \frac{\partial u}{\partial z}; \overline{v'w'} = -K_M \frac{\partial v}{\partial z}; \rho'w' = -K_H \frac{\partial \rho}{\partial z} \]  

(2.14)

where \( K_M \) is the eddy viscosity for momentum and \( K_H \) is the eddy diffusivity.

The terrain-following \( s \)-coordinates (Song and Haidvogel, 1994) allow for increased resolution in areas of particular interest, for example near the surface, thermocline and bottom boundary layer. The structure of the vertical levels is user controlled through setting \( V \)-stretching, \( V \)-transform, \( \theta_s \), \( \theta_b \) and thermocline depth \((tcline)\) parameters.

ROMS incorporates several coupled modules for processes such as sediment transport (Warner et al., 2008), biological processes (Fennel et al., 2006 and 2008) and sea-ice (Budgel, 2005). The inclusion of these modules within a simulation is optional and depends on the processes to be studied.
Vertical mixing within ROMS can be solved with a number of closure schemes. Generic Length Scale (GLS) schemes (Umlauf and Burchard, 2003) were implemented in ROMS by Warner et al. [2005] to represent a number of existing two-equation turbulence closures. The Mellor-Yamada level 2.5 scheme (Mellor and Yamada, 1974) is used widely in many models and has been implemented as the k-kl scheme within the GLS closure, while in Umlauf and Burchard [2003] a generic model was proposed which is also adopted here as GLS gen. Large et al. [1994] propose a competing LMD vertical mixing scheme which is based on a k-profile parameterisation. Here a boundary layer depth is determined at each grid cell where mixing is strongly enhanced with a polynomial profile in this layer, while matching the interior mixing at the boundary.

2.3.1 Bulk fluxes

In ROMS the heat fluxes are calculated using version 3.0 of the COARE algorithm (Fairall et al., 2003). This is based upon Monin-Obukhov similarity theory as described in Section 2.2.2, however here the stability functions used are a blend of the Kansas functions, at near neutral stability, with the improved representation of Fairall et al. [1996] for more convective situations. The scheme was specifically designed and tested over oceans, and is found to be accurate to within 5% for wind speeds up to 10m/s and within 10% for winds between 10 and 20m/s. The algorithm is largely untested for high wind speeds above 20m/s as there is insufficient measured data above this range. In the standard COARE 3.0, the surface roughness depends on a Charnock parameter that is constant for wind speeds up to 10m/s and above 20m/s, but increases linearly between these values. While this was based on results of Yelland and Taylor [1996], these results are no longer supported by the authors and as such this linear dependence (rather than a constant value) is controversial.

In the latest version of the COARE algorithm there are optional parameterisations which include wave conditions in the computation of the bulk fluxes. These are based on two different methods for the calculation of surface roughness, from Taylor and Yelland [2001] and Oost et al. [2002].

As suggested by Andreas et al. [1995], sea-spray may also play an important part in the bulk energy fluxes. Algorithms including the effects of sea spray were deemed insufficiently accurate for inclusion in the COARE 3.0 algorithm, but further attempts have been made (e.g. Andreas, 2008) which could warrant future inclusion.
2.3.2 Model setup

ROMS requires atmospheric forcing fields of surface winds, temperature, pressure and humidity along with incoming shortwave and outgoing longwave radiation and precipitation rate. These can be provided from observations, an atmospheric model or re-analysis data. For the purpose of this study atmospheric fields are taken from the WRF model and ERA-Interim reanalysis.

Model initialisation is taken from an average of May temperature and salinity profiles from the World Ocean Circulation Experiment. This mean profile is then applied across the basin with zero-velocity fields. From this, the model is spun-up with ERA-40 reanalysis forcing for 40 years, then with 1 year of WRF forcing.

The model bathymetry is a composite of data from the Caspian Environmental Program, Azerbaijan naval navigation charts, Turkmenistan hydrographic charts and side scan sonar data taken in the region around the measurement stations.

River forcing is provided for the five largest rivers flowing into the Caspian Sea (Volga, Ural, Kura, Terek and Emba). Values for monthly runoff for 2001 are taken from the Global River Discharge Database (RivDIS v1.1), where this is the most recent data available. This provides an additional source of fresh water into the upper layer of the water column, which is of particular importance in representing the dynamics of the North Caspian where the Volga inflow leads to much reduced salinity. The sea-ice module of Budgell [2005] is required given the freezing in the North Caspian in winter. This dynamical sea-ice model is based on ice thermodynamics and calculates prognostic ice concentrations and thicknesses, while using two layers to allow for temperature gradients within the ice.

As the Caspian Sea is an enclosed body of water no oceanic boundary forcing is required in this work. The KBG however must be considered as a sink of water. The channel between the Caspian Sea and KBG however is less than a kilometer wide and so cannot be accurately represented in the model setup employed here. Given this, the channel width is set to 8 km (two grid cells) and the permeability is reduced to ensure the flow is in line with historical observations. This however means that the KBG will not be accurately represented in the model and as such results for this area should not be trusted.

All simulations in this work are performed on a 4 km grid with 32 vertical levels. The vertical level structure is formed with $V$-stretching and $V$-transform set to 2; $\theta_s = 5$ and $\theta_b = 0.4$; and $tcline = 40$ m. The baroclinic time-step is 1 minute within which there are 20 barotropic time-steps. In Chapter 3 the effect of changing
the spacing of the vertical levels is briefly investigated, while both LMD and GLS vertical mixing schemes are employed.

## 2.4 SWAN Wave Model

The Simulating Waves Nearshore (SWAN) model (Booij et al., 1999) is a third-generation wave model which is used to simulate gravity wave fields in water depths ranging from the open ocean to coastal areas and shallow lakes. SWAN is based on the techniques used in the WAM model (WAMDI group, 1988) to model waves in the open ocean, but additionally includes processes associated with shallow waters to give a model that is more widely applicable. The model explicitly represents the main processes governing transfer of wave energy: wind generation, white-capping, bottom dissipation, depth induced breaking and both quadruplet and triad wave-wave interactions.

The model uses an unconditionally stable propagation scheme which doesn’t require the Courant stability condition to be satisfied (Holthuijsen, 2007) allowing free choice of temporal and spatial resolutions.

SWAN describes waves with a two-dimensional wave action density spectrum which implicitly allows the presence of currents to be considered. In Cartesian coordinates this is written as:

\[
\frac{\partial}{\partial t} N + \frac{\partial}{\partial x} c_x N + \frac{\partial}{\partial y} c_y N + \frac{\partial}{\partial \sigma} c_{\sigma} N + \frac{\partial}{\partial \theta} c_{\theta} N = \frac{S}{\sigma} \tag{2.15}
\]

where \( c_{x,y,\sigma,\theta} \) are the propagation velocities in \( x \) and \( y \) directions, and in frequency and directional space. The first term represents the local rate of change of action density in time, the second and third are the propagation of action density in the \( x \) and \( y \) directions respectively, the fourth term is the shift in relative frequency and the fifth term represents diffractive and refractive processes. The term on the right hand side of Equation 2.15 is the source term and includes representations of all of the processes involved in generation, dissipation and wave-wave processes. This equation can be formulated in terms of spherical co-ordinates for use when computations are to be performed over large areas.

SWAN was tuned and validated including the use of currents as input forcing (Ris et al., 1999). However, in that study, the currents considered are tidal rather than wind driven. Given that in the open ocean waves will usually be in the same
direction as the wind driven currents (when tides are removed), some of the effect of a following current will be implicitly included in the model tuning. It is therefore possible that including currents in the wave model forcing in the Caspian (where tides are negligible) will effectively mean the currents are considered twice, and so not improve wave prediction. This would be counter-intuitive as any additional information should improve model skill, assuming the relevant process is properly represented.

2.4.1 Model setup

For the computations in this study, the grid used is identical to that in ROMS, with 4 km resolution. The timestep is 5 mins, while directional resolution is $10^\circ$ and frequency resolution is logarithmic with 31 frequencies between 0.04 and 1 Hz.

As noted by Komen et al. [1996], the ability of a wave model to accurately reproduce observed conditions is heavily dependent on the meteorological forcing used, and generally the wave model errors are smaller than those of the wind models. Signell et al. [2005] use four different atmospheric models to force SWAN and find drastically different wave fields are produced depending on the forcing (Figure 2.2). They conclude that the advantages gained by using limited area atmospheric models, rather than general circulation models, are dependent on the region in question, but that when orographic effects are important (as with the Caspian), limited area models offer significant improvement in the modelled wave fields.

Given that, the simulations in this study are all forced by three-hourly wind output from WRF simulations. In each case, SWAN is initialised from rest, where the wave field is seen to take only one or two days to spin-up to a realistic situation.

2.5 COAWST model

The Coupled Ocean Atmosphere Wave Sediment Transport (COAWST) model was developed by Warner et al. [2010] to couple the individual WRF, ROMS and SWAN models. The Model Coupling Toolkit (MCT) is used to allow exchange of variables between the individual models, which are run concurrently, as described in Warner et al. [2008]. Where the individual models are run on different grids, interpolation is performed using the Spherical Coordinate Remapping and Interpolation Package (SCRIP: Jones, 1998).

Within this modelling system, any combination of wave, ocean and atmosphere
Figure 2.2: Wind forcing and resulting wave fields from four simulations of the SWAN wave model, using different wind products, from Signell et al. [2005].
models can exchange variables at user defined intervals. This allows the models to be run with higher resolution input fields, where, for example, ROMS can be run with atmospheric forcing from WRF at 10 minute intervals, which would be computationally expensive otherwise.

Figure 2.3 shows the transfer of variables between the model components of COAWST. The ocean model gives SST fields to the atmosphere, while it receives its required surface forcing fields, as well as latent and sensible heat fluxes which are computed within WRF. The atmosphere model gives surface wind forcing as required by the wave model, and in turn receives wave height, length and period which can be used to calculate surface roughness (as described in Section 1.1.3) which affects wind speeds. Between ocean and waves models the bathymetry, free-surface height and currents are fed to the wave model, while various wave parameters can be used in ROMS to calculate wave-driven flows, particularly near-bottom processes involved in sediment transport.
2.5.1 Model Setup

Where COAWST is used in Chapter 4 of this study, each of the individual models are used as previously described. The coupling occurs between each pair of models at 10 minute intervals.

As WRF is run on three successively smaller, higher resolution, grids, the coupling is performed on the inner nest. This means that it is the output from the 4 km grid which is used to force SWAN and ROMS, while these models influence the inner WRF grid. In future releases of COAWST it is planned to allow coupling to take place on each nested grid concurrently, but in the present model, communication is limited to one grid.

At present COAWST does not include sea-ice and as such simulations should not be performed when ice is present. For this reason simulations are limited to the period of 1st April to 1st December 2008.

2.6 Measurements

Through the course of this work, results from simulations are compared with observations. In this section the measurement methods are described.

2.6.1 Surface Wind Speeds - Quikscat

The Quikscat satellite employs scatterometry to measure wind speed and direction across the global oceans (JPL, 2006). The satellite follows a polar orbit, performing twice-daily passes over each point on the Earth’s surface with a 1800 km wide swath.

The scatterometry works by emitting microwave radar pulses and determining the surface wind behaviour based on the reflected energy from the sea-surface. Given that the scatterometry relies on measuring the roughness of the sea-surface, it can’t provide wind speed measurements over the land surface, or in coastal areas where land within the viewing area contaminates results. The study of Pickett et al. [2003] finds that accuracy of Quikscat is improved offshore compared with a range of distances about 20-50 km from the coast.

Given that the wind speed measurements depend on sea-state, it is clear that the retrieval will be affected by sea-ice. The satellite includes sea-ice detection, and this then is used to remove invalid data. However in the Caspian Sea the winter sea-ice goes undetected, and so the winds speeds provided are not removed, and do
not represent the reality. The winds in Quikscat affected by sea-ice appear too high and so are removed from part of the analysis.

In order to provide a gridded product, the data is interpolated both temporally and spatially which results in a smoothing beyond that of the 25 km resolution. The resulting product is a twice daily estimate of 10 m wind speeds on a 0.25° grid over the ocean surfaces. The operational range of Quikscat is 3-30 m/s, while accuracy is found to be between 1 and 2 m/s.

Given the wind speed measurements are not valid below 3 m/s, records at these values will be removed from some of the analysis. Where this is done, comparison is not made when the simulated wind is less than 3 m/s.

2.6.2 Sea Surface Temperatures - AMSR-AVHRR

Sea-surface temperatures (SSTs) are measured by a number of satellites, however in this study only two datasets are considered.

Advanced Microwave Scanning Radiometer (AMSR) gives through-cloud SST retrieval which allows for near 100% coverage. The microwave sensors have resolution of around 56 km, which causes problems near land, where data are not available.

Advanced Very High Resolution Radiometer (AVHRR) uses infrared, which means that cloud coverage is a barrier to SST retrieval. However in clear sky conditions the satellite provides twice daily observations on a 4.6 km grid.

These two satellite datasets have been combined as described in Reynolds et al. [2007]. This product is available on a 0.25° by 0.25° grid, providing daily SST values, which are bias adjusted with in situ measurements. By combining AMSR and AVHRR, coverage is available both in cloudy conditions as well as near land, while systematic biases are reduced.

2.6.3 Precipitation - TRMM-GPCC

Tropical Rainfall Measuring Mission (TRMM) is a satellite combining different measurements of rainfall (Huffman and Bolvin, 2007). The main instrument is the TRMM Microwave Imager (TMI) which measures microwave energy emitted by raindrops to quantify rainfall volumes. The data from TRMM is at 3 hour intervals between 40°S and 40°N, however products are also available, further from the equator, incorporating monthly station data.

The Global Precipitation Climatological Center (GPCC) rain gauge analysis (Schneider et al., 2008) has been combined with TRMM satellite measurements to
improve accuracy. The GPCC measurements are integral in time and so do not suffer inaccuracy of temporal resolution, although the geographic distribution can be patchy and involves lots of spatial smoothing, so this should not be used alone for spatial rainfall patterns. Therefore by combining GPCC with TRMM spatial coverage is added to local accuracy.

The combined GPCC-TRMM monthly mean rainfall is used in this study on a $0.25^\circ$ by $0.25^\circ$ grid which covers the Caspian Sea area.

It is noted that during the winter months there is often an intense peak in the precipitation from TRMM-GPCC in the North Caspian. This seems to be spurious as there is nothing in the literature suggesting strong precipitation in the area, and the monthly values can reach up to 700 mm in the satellite data. It seems likely that the anomalies are caused by the sea-ice affecting the satellite retrieval, as they only occur between December and March, and the extent of the area affected corresponds broadly with the sea-ice coverage. For this reason, these peaks are removed from the data before comparison is made with model results.

2.6.4 Station Measurements

In addition to the satellite observations, station measurements are available for surface wind speeds, significant wave heights and current velocities. These measurements provide a localised but high temporal resolution picture of the atmospheric and oceanic conditions in the Caspian. The location of station measurements is indicated in Figure 2.4.

Within the records, missing or erroneous data is noted and removed from the analyses. All of the measurements are taken on or near oil platforms, as such sheltering is a concern. The data have all been quality controlled to remove anomalous values.

Wind speeds

Wind speed measurements used in the study are taken at the Central Azeri platform located at 40.03N 51.35E. Measurements are available from 1st December 2007 to 1st December 2008.

Data is recorded by a cup and vane anemometer located on the top of an oil platform at 102 m above the mean sea level (MSL). 1 minute data records are subsampled to provide observations at 10 minutes intervals. Data are then quality controlled by ensuring values fall in an acceptable range; don’t contain spikes;
Figure 2.4: Bathymetry of the Caspian Sea (m). The stations where measurements are taken are indicated: Central Azeri (green), East Azeri (red), DWG (yellow) and Shah Deniz (magenta).

and are comparable with records at nearby sites. Generally the accuracy of cup anemometers is believed to be around 4%.

In order to provide surface wind speed values, those recorded at 102 m are converted to 10 m using the log-law relation:

$$\frac{U_{10}}{U_z} = \frac{\ln \left( \frac{U_{10}}{z_0} \right) - \psi \left( \frac{10}{L} \right)}{\ln \left( \frac{U_z}{z_0} \right) - \psi \left( \frac{1}{L} \right)}$$  \hspace{1cm} (2.16)

which gives wind speed at any height, $z$, given the wind speed at 10 m ($U_{10}$), the surface roughness length ($z_0$) and the stability function $\psi$ which depends on the Monin-Obukhov length ($L$). Conversion to 10 m values might introduce some error, particularly as variable roughness is not accounted for, however agreement with calibration measurements, taken nearer the surface, is good. All of the above mentioned post-processing is performed by the instrument suppliers prior to making the data available.

**Wave heights**

Wave height measurements are taken from three stations: East Azeri (40.02N 51.45E), Shah Deniz (39.90N 50.45E) and DWG (40.17N 51.17E). Measurements are avail-
able from 1st December 2007 to 1st December 2008 at Shah Deniz, and between 1st January and 1st December 2008 at East Azeri and DWG.

Wave measurements are taken over a 20 minute period, sampling twice a second, and then subsampled to provide a record every 10 minutes. The measurements are taken by a wave radar stationed around 30 m above MSL on an oil platform, which returns significant wave heights. The radar is downward looking and measures the distance to the sea-surface, such that over a number of samples the sea-surface profile can be determined to an accuracy of ±6 mm. The same quality control process is followed as with the wind speed measurements.

Currents

The current measurements used in this study are taken near the Shah Deniz station at 39.88N 50.37E. These measurements are available between 1st December 2007 and 13th September 2008 at 10 minute intervals.

Measurements are taken by two acoustic Doppler current profilers (ADCPs) working at 300 kHz. One ADCP is upward looking and one is downward looking, situated at approximately 46 m depth. Between the two profilers, measurements are taken at 37 depths between 6 m and 84 m below MSL at 2 m intervals (with the exception of between 42-50 m where none are taken). This provides near total coverage of the water column at the station, where the water depth is 92 m. The accuracy of measurements is believed to be of the order of 1 cm/s (Firing, 1991).
Chapter 3

Model Validation

The first aspect of this work is to ensure that the models used in the study are able to represent the observed environment, and be aware of their limitations. This is done by comparing in turn the atmospheric model, then the ocean and finally wave model with observations.

3.1 WRF Atmosphere Model

As the ocean and wave models both require atmospheric forcing it makes sense to first consider the atmosphere model used to provide this.

Within this section three different setups of WRF are employed. The first of these combines the MYJ PBL scheme with Thompson microphysics; the second uses the same microphysics but with the YSU PBL scheme; finally the WSM6 microphysics is used with YSU. ERA-Interim reanalysis data are also compared with observations as an example of the atmospheric data available without running a limited area model.

In this section simulations are performed from 1st December 2007 to 1st December 2008 with a three day model spinup before this.

3.1.1 Validation of surface winds

The key atmospheric driver of waves and oceanic circulation is the surface wind field. For that reason the WRF atmosphere model is validated against 10 m wind speed observations from station measurements and satellite data. Measured 10 m wind speeds are available from a station at Central Azeri at 10 minute intervals for the whole simulation period. Speeds are also compared with winds measured
by Quikscat. Here the two WRF setups using Thompson microphysics are used as a direct comparison of the MYJ and YSU planetary boundary layer schemes (and their associated surface layer schemes).

**Comparison of Quikscat with station measurements**

As there are two distinct sets of measurements of wind speeds it makes sense to first compare them to check their consistency. As Quikscat provides a spatial map of wind speeds, data is extracted corresponding to the location of the Central Azeri station.

Figure 3.1 shows a quantile-quantile plot (hereafter qq plot)\(^1\) of Quikscat winds against the measurements at Central Azeri between 1st Dec 2007 and 1st Dec 2008. Here, over the whole range of observed wind speeds, Quikscat shows larger values than the station measurements. The disparity between the measurements is greatest at low wind speeds; the station measurements often see winds below 3 m/s while Quikscat very rarely registers winds at these values. Overall Quikscat is biased 15% higher than the station measurements with a root mean square error (RMSE) of 2.0 m/s. From this it is clear that there is quite a disparity between the two sets of measurements, which is something to be aware of when comparing with the simulations.

The problem at low wind speeds occurs because Quikscat records very few instances of winds less than 3 m/s compared to the station measurements. This can possibly be explained by Quikscat winds being calculated from scatter from waves. Even when wind speeds are very low, waves will still propagate from elsewhere, and as such Quikscat might register winds when none are present. This is accounted for in that Quikscat’s operational range is from 3-30 m/s and the results are therefore not generally valid below this. This fits with the observation that there is not good agreement with the measurements below 3 m/s. The comparison between the measurement sets is then made ignoring all instances where the station measurements give wind speeds of less than 3 m/s. This results in Quikscat having a bias of +9.7% and an RMSE of 1.87 m/s. Therefore while this increases the agreement, there is still a significant difference between the measurements.

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\(^{1}\)A qq plot is a method of comparing the distribution of two datasets, and is made by plotting the quantiles of one dataset against those of another. By definition, the values for the predictor increase monotonically with the observations, and as such no idea of the correlation between them is given. For a predictor whose distribution is equal to that of the observations, all of the points would lie along the x=y line, while the divergence from this gives an idea of the tendencies of the predictor.
One possible explanation for the difference in the sets of measurements is that the station winds might suffer from sheltering. As the winds are recorded on an oil platform it is possible that winds are reduced by sheltering from the platform, which would explain why lower speeds were recorded than were seen by Quikscat.

**Comparison with station winds**

10 m wind speeds at the Central Azeri station are extracted at 3 hr intervals from the WRF simulations, and compared with the equivalent wind speed measurements. Where comparison is made with ERA-Interim it is done so at 6 hour intervals.

In Figure 3.2, qq plots of the simulated winds, from both WRF setups and from ERA-Interim, against the station measurements are shown. It is noticeable that WRF and ERA both have an overprediction of the wind speeds below 2 m/s. Above this, WRF generally has a consistent overestimation of wind speeds, which leads to biases of +11.6 and +10.0% from the MYJ and YSU setups respectively (Table 3.1). Given that the overprediction is constant through the range of observed wind speeds it would seem all winds tend to be overpredicted. At winds above 2 m/s ERA-Interim tends to fall below the measurements and has an overall bias of -6.3%. Here, Interim underpredicts the strongest winds by more than weaker events,
Figure 3.2: QQ plots of wind speeds from simulations of WRF using the MYJ and YSU planetary boundary layer schemes, and ERA-Interim, against station measurements at Central Azeri between 1st Dec 2007 - 1st Dec 2008

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Station Measurements</th>
<th>Quikscat</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Bias (%)</td>
<td>RMSE (m/s)</td>
</tr>
<tr>
<td>WRF with MYJ</td>
<td>+11.6</td>
<td>2.63</td>
</tr>
<tr>
<td>WRF with YSU</td>
<td>+10.0</td>
<td>2.62</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>-6.3</td>
<td>2.24</td>
</tr>
</tbody>
</table>

Table 3.1: Comparison of simulation and ERA-Interim surface wind speeds at Central Azeri between 1st Dec 2007 - 1st Dec 2008 with station measurements and Quikscat satellite data

therefore it performs worst at the extremes. The WRF simulations have RMSE values of 2.63 and 2.62 m/s while that for ERA-Interim is 2.24 m/s, this shows that while WRF is better at predicting the distribution of the magnitude of wind speeds (Figure 3.2), Interim tends to better represent the timing and presence of events. This is not entirely surprising as ERA-Interim is a reanalysis product which assimilates (other) measurements.

Comparison can be made for each season\(^2\) individually, and is shown in Figure 3.3. This shows a similar pattern for each season, where wind speeds are overesti-

\(^2\)Throughout this study the seasons are defined as winter (Dec-Jan-Feb), spring (Mar-Apr-May), summer (June-July-Aug) and autumn (Sep-Oct-Nov).
Figure 3.3: QQ plots of seasonal wind speed from simulations of WRF with MYJ (blue), YSU (green) PBL schemes and ERA-Interim (light blue) against station measurements at Central Azeri

...mated over the whole range. The exception to this is in spring where winds up to around 9 m/s are good, but show positive bias above this.

The directionality of the winds is displayed in Figure 3.4, where it can be seen that the differences between the two WRF simulations are minimal. The simulations and the measurements show that the winds are predominantly from the North. WRF predicts a larger spread of wind direction than is observed, where easterly winds are too frequent. The measurements show correspondence with the observation that winds tend to be northerly around 40% of the time over the Caspian, while the model has lower occurrence of northerlies, around 35%.

The seasonality of wind direction is compared between the MYJ simulation of WRF and the measurements in Figures 3.5 and 3.6. It is clear that northerly winds are more common in the summer than other seasons and this is captured in the model. Agreement is generally good, apart from in autumn where WRF has mainly easterly winds, which is not seen in the measurements. The differences in annual wind directions are mainly accounted for by the errors in the autumn, while the model has good skill through the other seasons.
Figure 3.4: Wind roses showing direction of travel of winds from WRF simulations (a and b) and measurements (c) at Central Azeri between 1st Dec 2007 - 1st Dec 2008
Figure 3.5: Wind roses showing direction of travel of winds for each season from the MYJ WRF simulation at Central Azeri between 1st Dec 2007 - 1st Dec 2008

Figure 3.6: Wind roses showing direction of travel of winds for each season from the measurements at Central Azeri between 1st Dec 2007 - 1st Dec 2008
Figure 3.7: QQ plots of wind speeds from simulations of WRF using the MYJ and YSU planetary boundary layer schemes, and ERA-Interim, against Quikscat measurements at Central Azeri between 1st Dec 2007 - 1st Dec 2008

Comparison with Quikscat winds

First the models are compared with Quikscat satellite measurements at Central Azeri, replicating the comparison with station measurements. A qq plot of both sets of simulated winds from WRF and the Interim reanalysis is shown in Figure 3.7. Here both WRF simulations and Interim underpredict the winds below 4 m/s, but above this the WRF results follow the Quikscat measurements very closely while ERA-Interim gives consistently lower wind speeds than Quikscat. Instances when the model predicts wind speeds of less than 3 m/s are removed from the quantitative analysis. The MYJ setup of WRF has a bias of +3.2% and RMSE of 2.33 m/s, the YSU setup has a bias of +1.0% and RMSE of 2.24 m/s while ERA-Interim is biased 12.3% low with an RMSE of 2.32 m/s. Here we can conclude that WRF is a good predictor of the Quikscat winds at Central Azeri. The winds are predicted with a low error and the whole range of observations is well matched (with the exception of the lowest winds where Quikscat is not valid), while the bias remains small.

The comparison is then extended to spatial validation of the simulations against Quikscat over the whole Caspian. In Figure 3.8 the annual mean bias and RMSE from the WRF simulation with MYJ PBL is shown. In the North Caspian the
Figure 3.8: Mean bias and RMSE (m/s) of wind speeds for the MYJ WRF simulation against Quikscat measurements between 1st Dec 2007 - 1st Dec 2008. This does not account for sea-ice and low wind errors in Quikscat.

The simulation has a large negative bias of more than 1 m/s, while everywhere else the bias is generally less than ±0.5 m/s. The RMSE in the North Caspian is also poor, with values exceeding 3.5 m/s.

To look in more detail at the comparison with Quikscat, seasonal mean bias and RMSE is shown in Figure 3.9 for the MYJ simulation. Here it is clear that the wind speeds are much too low in the North Caspian in winter and spring, causing large RMSEs. This occurs when sea-ice is present and so can be explained by error in the satellite retrieval in these areas, as discussed in Section 2.6.1. Through the rest of the year model skill in the North Caspian is much better, and is comparable to other areas. For this reason, the comparison is then made removing values where sea-ice affects the data, along with the low winds for which Quikscat is not valid.

In Figure 3.10 the annual mean bias from both WRF simulations is shown. It is obvious from the North Caspian the effect that the sea-ice had on the data retrieval,
Figure 3.9: Seasonal mean bias (m/s) of wind speeds for the MYJ WRF simulation against Quikscat measurements between 1st Dec 2007 - 1st Dec 2008. This does not account for sea-ice and low wind errors in Quikscat.

<table>
<thead>
<tr>
<th>Simulation setup</th>
<th>Mean bias (%)</th>
<th>Mean RMSE (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>WRF MYJ</td>
<td>+7.1</td>
<td>2.47</td>
</tr>
<tr>
<td>WRF YSU</td>
<td>+6.0</td>
<td>2.36</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>-7.8</td>
<td>2.29</td>
</tr>
</tbody>
</table>

Table 3.2: Comparison of surface wind speeds with Quikscat over the whole of the Caspian between Dec 2007 and Dec 2008

where now the bias here is positive. Both simulations have positive bias through the majority of the Caspian, which is highest in the North Caspian. The simulation with the MYJ PBL has slightly higher winds than the YSU setup throughout, leading to a mean bias of +7.1% compared to +6.0% (Table 3.2). Figure 3.11 shows RMSE of both simulations against the Quikscat measurements. Overall mean RMSEs of 2.47 (MYJ) and 2.36 (YSU) m/s are seen for the two WRF simulations, and the YSU setup has noticeably smaller errors. Here both simulations represent the spatial mean winds over the whole Caspian well, while the MYJ setup tends to predict slightly stronger winds everywhere.

Figure 3.12 shows the seasonal mean bias. Generally there is a positive bias through spring and summer and a negative bias in wind speeds in winter. This trend is not seen when comparing with the station data (Figure 3.3) and the reason
Figure 3.10: Mean bias (m/s) of wind speeds for the MYJ and YSU WRF simulations against Quikscat measurements between 1st Dec 2007 - 1st Dec 2008

Figure 3.11: RMSE (m/s) of wind speeds from the MYJ and YSU WRF simulations against Quikscat measurements between 1st Dec 2007 - 1st Dec 2008
Figure 3.12: Seasonal mean bias (m/s) of wind speeds for the MYJ WRF simulation against Quikscat measurements between 1st Dec 2007 - 1st Dec 2008.

for it is unclear.

ERA-Interim wind speeds are also compared with Quikscat where the bias and RMSE are shown in Figure 3.13. The ERA-Interim winds are too low over the whole of the Caspian, with negative biases of more than 0.5 m/s through most of the Caspian. The exception is in the North Caspian, where the winds through the winter were not included in the comparison. In the Central and Southern basins ERA-Interim has lower errors than WRF, with the exception of coastal areas where the errors tend to be larger. One might expect Interim to perform worse around the coasts as the resolution is low and as such the winds here may be contaminated with values over land. Wind speeds over land tend to be lower, so this could explain the negative bias in coastal areas in ERA-Interim. Overall the mean bias of ERA-Interim winds is -7.8% and the RMSE is 2.29 m/s.

3.1.2 Evaporation and Precipitation

To get an idea of the skill of the models in representing the water budget, evaporation and precipitation can be compared with measured values. Measured precipitation
Figure 3.13: Annual mean bias and RMSE (m/s) of wind speeds from ERA-Interim against Quikscat
from the combined TRMM-GPCC product can be compared with model output, while comparison is also made with Caspian wide climatological values of precipitation and evaporation from the literature.

Evaporation

Comparing evaporation from the WRF simulations, it can be seen from Figure 3.14 that the YSU setup has higher evaporation. There is 1360 mm/year of evaporation from the Caspian Sea in the YSU simulation using Thompson microphysics compared with 1098 mm/year in the equivalent MYJ run. The evaporation fields using the YSU PBL show differences when the microphysics scheme is changed: there is almost the same total evaporation using the Thompson and WSM6 microphysics, but there is less using the WSM6 scheme when the KBG is ignored. These differences likely come from the impact of clouds on the surface fields.

The increased evaporation in the YSU setup comes despite lower surface wind speeds and both setups receiving the same sea-surface temperatures. The near-surface air temperature is also an average of 0.25°C hotter, which should lead to a more stable boundary layer, and hence less evaporation. As such the difference in

Figure 3.14: Total evaporation (mm) from the WRF simulations between 1st Dec 2007 - 1st Dec 2008.
evaporation between the two schemes is likely due to increased vertical turbulence in the YSU scheme transporting moisture out of the surface boundary layer, as seen by Hu et al. [2010].

Evaporation from the MYJ simulation is shown in Figure 3.15 for each season. Here it can be seen that evaporation peaks in autumn, and is at its lowest in spring. The highest evaporation occurs when the SSTs are high relative to the surface air temperature: in autumn; in the deep Central and South Caspian in winter; and the shallow North Caspian in summer. It can be seen that evaporation in the summer in the North Caspian is around twice that elsewhere, as expected. The expected high evaporation along the east coast of the central Caspian in winter is seen in the model, however the low summer evaporation in the same region is not seen. This is likely because the low SSTs due to upwelling causing this effect in summer are not captured in ERA-Interim which is used as the forcing, while the high winter evaporation is caused both by high SSTs (which are not captured in ERA-Interim) and intrusion of cold dry air, which is captured by ERA-Interim.

While there are no satellite observations of evaporation, comparisons can be made with values quoted in the literature. This must be done with caution however, as evaporation measurements tend to be the product of simple bulk formulae.
or residuals in the Caspian sea level changes. The range of quoted annual mean evaporation totals is 918-1039 mm/year. In comparison, the WRF simulations give values of 1087, 1357 and 1342 mm/year. It should not necessarily be expected that the simulations fall in the range of climatological mean values, however the observed range in Rodionov [1994] is from 790-1230 mm/year. The simulations using the YSU planetary boundary layer scheme therefore predict more evaporation than was seen for any year in the last century, and therefore is seen to overestimate the evaporation. The simulation using the MYJ scheme lies within the range of observations and close to the climatological mean.

Precipitation

Precipitation from WRF is compared between the three setups. Figure 3.16 shows that the precipitation in each WRF simulation is mainly concentrated in two regions, over the Caucasus mountains and along the south western corner of the Caspian Sea. The pattern in rainfall in all setups is very similar, however there is more rainfall in the most intense regions using the WSM6 microphysics scheme. This results in the WSM6 run having mean precipitation over the model domain of 328 mm/year compared to 251 mm/year using Thompson microphysics with the YSU PBL (Table 3.3). The effect of different PBL schemes is smaller, with 269 mm/year using Thompson microphysics and MYJ PBL. The reason for this difference is not clear, as the YSU scheme evaporates more water from the surface, so might be expected to have more precipitation.

The precipitation fields from the one year WRF simulations can be compared with those observed by TRMM-GPCC data between 1st December 2007 and 1st December 2008 (Figure 3.17). The satellite observations show peaks of precipitation over the Caucasus mountains and in the south west corner of the Caspian, as seen in the model results, although the intensities are lower than predicted. The observations also show that the areas of lowest precipitation occur along the east coast of the Caspian Sea and to its east; as is the case in the model. Generally the observations have a smoother distribution of precipitation than the model as would be expected given the lower resolution and the spatial interpolation involved in their calculation. The spatial distribution of precipitation is very well captured by the WRF simulations.

Comparing the domain total precipitation is perhaps the best way to quantitatively compare the simulations with satellite observations. Given one of the areas
Figure 3.16: Total precipitation (mm) from the WRF simulations between 1st Dec 2007 - 1st Dec 2008.

Figure 3.17: Total precipitation (mm) from the TRMM-GPCC data between 1st Dec 2007 - 1st Dec 2008.
Table 3.3: Comparison of modelled precipitation ($P$) and evaporation ($E$) from 1st Dec 2007 - 1st Dec 2008 with TRMM-GPCC satellite precipitation measurements (all values are mm/year)

<table>
<thead>
<tr>
<th>Simulation setup</th>
<th>Domain P</th>
<th>Caspian and KBG E</th>
<th>P</th>
<th>E-P</th>
<th>Caspian E</th>
<th>P</th>
<th>E-P</th>
</tr>
</thead>
<tbody>
<tr>
<td>WRF MYJ and Thompson</td>
<td>269</td>
<td>1098</td>
<td>281</td>
<td>817</td>
<td>1087</td>
<td>289</td>
<td>798</td>
</tr>
<tr>
<td>WRF YSU and Thompson</td>
<td>251</td>
<td>1360</td>
<td>271</td>
<td>1089</td>
<td>1357</td>
<td>280</td>
<td>1077</td>
</tr>
<tr>
<td>WRF YSU and WSM6</td>
<td>328</td>
<td>1361</td>
<td>340</td>
<td>1021</td>
<td>1342</td>
<td>349</td>
<td>993</td>
</tr>
<tr>
<td>TRMM-GPCC</td>
<td>289</td>
<td>-</td>
<td>299</td>
<td>-</td>
<td>-</td>
<td>310</td>
<td>-</td>
</tr>
</tbody>
</table>

of peak precipitation lies so close to the edge of the Caspian Sea, a slight shift in location of precipitation might lead to a more dramatic change in total precipitation over the water surface. However it is important to look at the volume of precipitation over the water surface from the point of view of the water budget of the Caspian Sea. Many of the values quoted in the literature ignore precipitation over the KBG so equivalent values are also computed here.

Table 3.3 compares the observed precipitation with that simulated by the WRF simulations. From this it can be seen that both over the domain as a whole and over the Caspian Sea specifically the total volume of precipitation is well captured in the simulations. Over the whole domain TRMM-GPCC has 289 mm/year, while the simulations predict 269, 251 and 328 mm/year. Over the water surface (both with and without KBG) the WRF simulations with Thompson microphysics slightly underpredict the total precipitation although the values from the WSM6 simulation estimate more than the satellite observations.

The simulated precipitation totals can also be compared with climatological values quoted in the literature about the Caspian’s water budget (as presented in Section 1.3.2). Here the range of values for mean precipitation over the Caspian Sea was from 195-257 mm/year, while the annual range was 110-310 mm/year (Rodionov, 1994). The quoted range is slightly lower than the TRMM-GPCC climatology of 275 mm/year, which implies either that precipitation over the Caspian has increased between the period of the observations quoted in the literature (up to 1990) and the satellite observation period (1999-2008) or that there is a measurement bias. The totals for the simulations with Thompson microphysics lies close to, but above, the climatology, while the WSM6 scheme has a large overestimation compared to the climatology. The simulated precipitation from the Thompson microphysics scheme falls between the range of measurements when used with either MYJ or YSU planetary boundary layer schemes, and as such is seen to give a good representation of
the observed conditions. The WSM6 microphysics scheme however seems to predict slightly too much precipitation.

The seasonality is shown in Figure 3.18 for the MYJ-Thompson simulation and the TRMM-GPCC observations, while the other WRF runs gave similar results. From this a very distinct seasonality is clear. The rainfall peak over the south-west of the Caspian is mostly confined to autumn and winter. The temporal agreement with the satellite observations is again good.

Evaporation was seen to be highest in autumn and lowest in spring, while this is also the case with the rainfall in the south-west. This might indicate that the precipitation is directly driven by evaporation from the Caspian Sea. The rainfall could be caused by moist air travelling south over the Caspian and ascending over the coast, meaning that precipitation in the region might be driven by northerly winds. However it was seen in Figure 3.5 that the winds in autumn are not predominantly from the North, while they are in spring. This seems to suggest that the wind direction is not the driver for the seasonal cycle of rainfall in the region.

**Water Budget**

Through these comparisons it is clear that of the WRF simulations, that with the YSU setup has larger evaporation totals, and that the WSM6 scheme gives more precipitation. The important factor in Caspian Sea level change however is the balance between evaporation and precipitation (or $E - P$). The MYJ setup of WRF with Thompson microphysics gives an $E - P$ of 798 mm/year for the year simulation, while YSU with the same microphysics gives 1077 mm/year. Comparing the WSM6 microphysics scheme with that of Thompson using the same PBL shows that the $E - P$ is much higher with Thompson, at 1077 mm/year compared to 993 mm/year. This means that the YSU setup will lose nearly 300 mm of water per year relative to the MYJ setup, with the same microphysics scheme, which would lead to a large divergence in Caspian sea level if continued over a longer simulation.

These $E - P$ values can be compared with those quoted in the literature where the range of mean $E - P$ estimates is 688-800 mm/year. This shows that the MYJ simulation of WRF lies within the range of the observations, while those with the YSU PBL seem to be too high in their $E - P$ prediction. This is clearly because of the much higher evaporation in the YSU PBL scheme when compared with the MYJ scheme.

These values of $E - P$ in the literature should be a good representation of reality
Figure 3.18: Seasonal precipitation (mm) from the MYJ-Thompson WRF simulation (a) and the TRMM-GPCC data (b)
as they are generally calculated from observations of river inflow into the Caspian and change in the sea level, however, given the rapid changes in water level observed historically it can not be assumed that the simulation period used here corresponds to the mean conditions. From Figure 1.4 it can though be seen that there is little change in CSL over the simulation period, and as such the conditions during this year are more likely to correspond to the climatological mean.

Based upon this assumption it can be concluded that WRF is best able to represent the observed water balance over the Caspian Sea with the MYJ PBL and Thompson microphysics schemes, while this setup also gives good estimates of both evaporation and precipitation individually.

### 3.2 ROMS Ocean Model

In this section, results from simulations of the ROMS ocean model are discussed. Different vertical mixing schemes are first tested to determine which is best able to represent the observed current profiles. Following on from this, the effect of different atmospheric forcing on the currents and sea-surface temperatures (SSTs) is investigated. Simulations are forced by ERA-Interim as well as output from two WRF model simulations. These were using both MYJ and YSU setups with the Thompson microphysics; the results of both were discussed in Section 3.1. All of the simulations performed in this section are for one year between 1st December 2007 and 1st December 2008.

#### 3.2.1 Vertical mixing schemes

Simulations were performed using WRF forcing (with the MYJ setup) and employing 3 different vertical mixing schemes. These schemes are the LMD scheme, the k-kl GLS and generic GLS schemes (as discussed in Section 2.3). A further simulation was performed using LMD vertical mixing but with a different structure of the vertical levels within ROMS, in order to briefly investigate the effect this plays. This setup had the same number of vertical levels, with an increased resolution near the surface, and correspondingly lower resolution further down in the water column.

Simulated currents are compared with those measured at the Shah Deniz station to determine how well the models represent the observed vertical structure. The measurements are available from the beginning of the simulation until 13th

\[ \theta_s = 8, \theta_b = 0.4 \text{ and } tcline = 10. \]
September 2008, as such these comparisons are performed over a period of 9 and a half months. This is done first by looking at the mean current profile, and then by looking more closely at three depths: 6 m representing the near surface; 20 m which tends to be around the base of the mixed layer; and 70 m showing the conditions at depth near the bottom.

Current profiles

In Figure 3.19 the mean current profiles are compared with measurements, where currents tend to decrease with depth. All of the simulations have a ‘trough’ of lower currents between 20 and 70 m depth which is not seen in the measurements. However all of the simulations capture the near bottom reduction in currents from about 75 m downwards very well. In the upper layer the simulations seem to capture the observed current behaviour, but with a consistent underprediction from 20 m upwards. The measurements show a decrease in currents from the surface until around 20 m, then the currents plateau and are quite constant until nearly 30 m deep when they begin to drop off again. This plateau behaviour is generally not well captured by the models, although both simulations using LMD mixing do suggest similar behaviour, but nearer the surface between 12 and 20 m; this is not seen in either GLS simulation.

The current profiles for all three simulations using the same vertical level structure show similar results. Using the generic GLS mixing gives the strongest currents through most of the column (down to about 50 m) which therefore agrees most closely with the measured profile. However below this is has the weakest currents and so performs worst. The LMD and GLS k-kl simulations give very similar results, the only significant difference being that the LMD simulation shows a plateau in currents, which might suggest it is capturing the physical behaviour better. The comparison between the LMD simulations employing different vertical structures shows that with the surface intensified resolution, the currents are weaker in the top 50 m of the column and then slightly stronger below this. This means that the results are generally further away from the measurements when using the setup with increased near-surface resolution.

Surface, mid-depth and bottom currents

Figure 3.20 shows q-q plots of the simulated and measured current speeds at 6, 20 and 70 m depth.
Figure 3.19: Mean current magnitude profiles from ROMS simulations and measurements at Shah Deniz between 1st Dec 2008 - 13th Sep 2008

From Figure 3.20a it is obvious that the underprediction of currents at 6 m in all the simulations (as seen in Figure 3.19) is mainly due to an underprediction of high current events above 0.55 m/s. At lower speeds the simulations seem to accurately represent the distribution of current magnitudes, but they fail to capture the largest events. As seen in Figure 3.2 the high winds speeds were not underpredicted, which means that this problem is not a direct result of inaccurate wind forcing. Comparing the three simulations, it seems that the GLS schemes perform slightly better than LMD which has a small underprediction of all current speeds.

Contrary to the currents at 6 m, those at 20 m (Figure 3.20b) show a consistent underprediction throughout the whole range of speeds. The differences between the simulations are small, with the generic GLS scheme performing best for the lower currents and the k-kl GLS setup best for stronger currents.

In Figure 3.20c there is a less obvious pattern to the underprediction of the mean near bottom currents. The currents have a slight underprediction up to about 0.4 m/s, but then the larger events are better represented. This is where the most dramatic difference between the simulations is found, with the LMD scheme performing much better for the strongest events than either GLS simulation.

Table 3.4 shows values of the mean bias and RMSE from the simulations. The biases quantify the information in the vertical profiles, but generally show that the
Figure 3.20: QQ plots of currents at 6m (a), 20m (b) and 70m (c) from ROMS simulations, with different vertical mixing, against measurements at Shah Deniz between 1st Dec 2008 - 13th Sep 2008.

generic GLS scheme is best near the surface and at 20 m, while the k-kl GLS scheme is best at 70 m. However the RMSE provides additional information. The LMD simulation has the smallest RMSE of the four simulations at each of the three depths considered. The $\theta_s = 8$ simulation has larger RMSE than the other LMD simulation at all depths and also larger biases at 6 and 20 m, although it does perform better near the bottom.

From these comparisons, all three vertical mixing schemes and both vertical structures are seen to represent the observed currents at Shah Deniz. However the LMD scheme seems to perform best of the three (given its low RMSE values), and it provides better results when the model vertical levels are more evenly spread through the water column. This is therefore the setup which is adopted for all the subsequent simulations performed in this study.
### Table 3.4: Comparison of modelled currents from ROMS simulations with different vertical mixing schemes, with station measurements at depths of 6, 20 and 70 m between 1st Dec 2007 - 13th Sep 2008

<table>
<thead>
<tr>
<th>Vertical mixing</th>
<th>6m Bias (%)</th>
<th>RMSE (m/s)</th>
<th>20m Bias (%)</th>
<th>RMSE (m/s)</th>
<th>70m Bias (%)</th>
<th>RMSE (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LMD</td>
<td>-13.0</td>
<td>0.100</td>
<td>-11.3</td>
<td>0.081</td>
<td>-11.3</td>
<td>0.081</td>
</tr>
<tr>
<td>GLS k-kl</td>
<td>-10.2</td>
<td>0.107</td>
<td>-16.4</td>
<td>0.089</td>
<td>-6.5</td>
<td>0.086</td>
</tr>
<tr>
<td>GLS gen</td>
<td>-7.1</td>
<td>0.105</td>
<td>-11.0</td>
<td>0.089</td>
<td>-16.5</td>
<td>0.082</td>
</tr>
<tr>
<td>LMD ($\theta_s=8$ and $tcline=10$)</td>
<td>-16.0</td>
<td>0.104</td>
<td>-18.6</td>
<td>0.089</td>
<td>-7.1</td>
<td>0.085</td>
</tr>
</tbody>
</table>

### 3.2.2 Effect of Different Atmospheric Forcing

ROMS simulations were performed using atmospheric forcing from ERA-Interim and the two WRF simulations. Comparisons of these simulations are made with currents at the Shah Deniz station and sea-surface temperatures (SSTs) from the AMSR-A VHRR satellite product.

#### Currents

Mean current profiles from the simulations and measurements are shown in Figure 3.21. The simulated profiles with WRF forcing are both very similar, and as described in Section 3.2.1, where currents are underpredicted at all depths (especially between 20 and 70 m) until about 80 m where the near bottom currents are well represented. In the simulation forced by ERA-Interim however the underprediction of currents is much greater and extends to the near bottom currents. It is clear from the profiles that with Interim forcing, ROMS isn’t able to represent the current profile. Between the two sets of WRF forcing, the run forced by the YSU setup has slightly stronger currents above 50 m and slightly weaker below this, however the profiles seem to mirror each other and the differences are small.

Looking at the currents near the surface we see in Figure 3.22a that using either set of WRF forcing, ROMS underpredicts the strongest currents (above 0.5 m/s) but represents those at lower speeds well. The YSU forcing is better for the currents up to 0.5 m/s, but above this the underprediction is less pronounced with MYJ forcing. ERA-Interim forcing leads to a consistent underprediction of all current magnitudes. Table 3.5 shows that the bias in the Interim forced simulation is -51%, while those forced by WRF are -13 and -9% for MYJ and YSU respectively. The RMSE is almost unaffected by the choice of WRF forcing, but is 20% higher when ERA-Interim is used.

Figure 3.22b, at 20 m, shows again that ROMS underpredicts the range of cur-

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Figure 3.21: Mean current magnitude profiles from ROMS simulations, using different atmospheric forcing, and measurements at Shah Deniz between 1st Dec 2008 - 13th Sep 2008

<table>
<thead>
<tr>
<th>Forcing</th>
<th>6m Bias (%)</th>
<th>RMSE (m/s)</th>
<th>20m Bias (%)</th>
<th>RMSE (m/s)</th>
<th>70m Bias (%)</th>
<th>RMSE (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>WRF (MYJ)</td>
<td>-13.0</td>
<td>0.100</td>
<td>-15.2</td>
<td>0.088</td>
<td>-11.3</td>
<td>0.081</td>
</tr>
<tr>
<td>WRF (YSU)</td>
<td>-9.0</td>
<td>0.099</td>
<td>-12.8</td>
<td>0.084</td>
<td>-13.7</td>
<td>0.080</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>-50.7</td>
<td>0.120</td>
<td>-51.1</td>
<td>0.113</td>
<td>-34.2</td>
<td>0.092</td>
</tr>
</tbody>
</table>

Table 3.5: Comparison of modelled currents from ROMS, using different atmospheric forcing, with station measurements at depths of 6, 20 and 70 m between 1st Dec 2007 - 13th Sep 2008
Figure 3.22: QQ plots of currents at 6m (a), 20m (b) and 70m (c) from ROMS simulations, using different atmospheric forcing, against measurements at Shah Deniz between 1st Dec 2008 - 13th Sep 2008
rents when forced by ERA-Interim. Both sets of WRF forcing lead to a consistent underprediction of current magnitudes, where underprediction is smaller for YSU for most of the range, but MYJ does better for the extremes. Overall these simulations have biases of -15% and -13% from MYJ and YSU, and -51% from Interim. As near the surface, the RMSE values from both WRF forced simulations are similar (YSU is slightly better) and about 20% less than that of the ERA-Interim forced simulation.

Near the bottom, the currents produced by both WRF forced runs are very similar until 0.3 m/s, above which the MYJ setup gives stronger currents (Figure 3.22c). This means the simulation with the MYJ setup of WRF as forcing is able to represent the extreme bottom currents while YSU underpredicts them. Again ERA-Interim shows the most dramatic underprediction with a bias of -34% compared with -11% for MYJ and -14% for YSU WRF.

Overall, it is obvious that ROMS doesn’t give a good representation of the observed currents at Shah Deniz when ERA-Interim forcing is applied. However, when output from the WRF model is used to force simulations, good agreement with current measurements is observed. The main concern is a mean underprediction of current magnitudes at all depths down to 80 m, while the current speeds below this are good. This bias is due to an underprediction of strong currents near the surface, while further down the water column there is a more general underprediction of the range of current magnitudes. The YSU setup of WRF is generally better at predicting the mean current behaviour, while the MYJ forced run performs better for strong currents at all depths.

**Sea-surface temperatures**

Comparison of SSTs is made between output of the three ROMS simulations and AMSR-AVHRR data. The comparisons are made for the length of the one year simulation.

Figure 3.23 shows the mean bias of the WRF and ERA-Interim forced ROMS simulations against the measurements. All three model simulations are biased high over much of the basin, and the simulation forced by MYJ-WRF is hotter than YSU-WRF. This is evident in the Central Caspian where the MYJ setup leads to mean SSTs around 1 °C too high, whereas the YSU run has biases here of around +0.75 °C. In the North Caspian all the simulations have very low bias, as they do in the middle portion of the South Caspian, while in the shallow areas here SSTs are
overpredicted. All three simulations perform poorly in the KBG which should not be surprising given its crude representation within the model. For this reason SST comparisons are made with and without the KBG and shown in Table 3.6. From this we see that when the KBG is ignored, the simulations have warm biases of +0.7 (MYJ), +0.5 (YSU) and +0.67 (Interim) °C. The ERA-Interim forced simulation has similar biases to the WRF forced simulations, the difference is that Interim forced ROMS has strong positive biases down the west coast of the South and Central Caspian which are not obvious in the other simulations.

The RMSE of SSTs is shown in Figure 3.24. Here it can be seen that the

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Mean Bias (°C)</th>
<th>Mean RMSE (°C)</th>
<th>Mean Bias (°C)</th>
<th>Mean RMSE (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ROMS with MYJ WRF</td>
<td>+0.61</td>
<td>1.84</td>
<td>+0.70</td>
<td>1.75</td>
</tr>
<tr>
<td>ROMS with YSU WRF</td>
<td>+0.44</td>
<td>1.66</td>
<td>+0.52</td>
<td>1.55</td>
</tr>
<tr>
<td>ROMS with ERA-Interim</td>
<td>+0.61</td>
<td>1.58</td>
<td>+0.67</td>
<td>1.48</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>+0.25</td>
<td>1.45</td>
<td>+0.24</td>
<td>1.33</td>
</tr>
</tbody>
</table>

Table 3.6: Comparison of sea-surface temperatures from ROMS simulations and ERA-Interim with AMSR-AVHRR between 1st Dec 2007 - 1st Dec 2008
errors are large in the KBG (as expected) and around some parts of the coastline, which could be because of contamination of the satellite data. The RMSE tends to be higher in the shallow areas of the North Caspian and south-east Caspian than throughout the interior. Figure 3.23 this seems to show that ROMS is too warm in the shallow areas when compared to AMSR-AVHRR. Throughout the Caspian, the simulations have mean RMSE values of 1.75, 1.55 and 1.48 °C (WRF-MYJ and -YSU and ERA-Interim respectively).

The skill of ROMS in predicting SST can be determined in more detail by looking more closely at the evolution of SSTs throughout the year. This is done in detail for the simulation forced by output from the MYJ setup of WRF, and some snapshots are shown in Figure 3.25. In the winter months the simulations show a cold tongue of water down the west coast of the Central Caspian, which is not present in the satellite data (Figure 3.25a) although it is expected from the literature (Rodionov, 1994). This may be because the satellite data is not high-enough resolution to capture it, or because it flows too close to the coast to be seen without data contamination. The sea-ice extent over winter in the North Caspian is very well represented by the
simulations, but ROMS simulates freezing in the KBG, which is not seen in the satellite data. During the sea-ice melting season of March-April the simulations don’t melt the ice quickly enough, which leads to the SSTs being too cold in April (Figure 3.25b). Moving into summer, the warming is very well represented, as is a cool area due to upwelling on the east coast of the Central Caspian (Figure 3.25c). Throughout autumn the development of SSTs is again good, but happens around 5 days later than in the observations (Figure 3.25d).

It is also instructive to compare directly the ERA-Interim SST values with AMSR-AVHRR. In Figure 3.26 the bias of ERA-Interim SST is shown, and it is clear that the biases tend to be lower than that of any of the ROMS simulations, with a mean bias of +0.24 °C. Similarly the RMSE of Interim SSTs (also shown in Figure 3.26) tends to be lower than that of the ROMS simulations, particularly in

Figure 3.25: Snapshots of SST (°C) from a ROMS simulation (forced by MYJ WRF) and AMSR-AVHRR satellite data. From: 17th Dec 2007 (a), 5th Apr 2008 (b), 22nd June 2008 (c) and 21st Sep 2008 (d).
the shallow coastal areas, and the mean RMSE is 1.33 °C.

The SSTs provided by ERA-Interim match the satellite observations of AMSR-AVHRR more closely than any of the three ROMS simulations. This is to be expected as ERA-Interim incorporates satellite SST measurements within its reanalysis and as such should be expected to accurately represent the broad-scale features of the observations. Whilst the ROMS simulations don’t incorporate any SST observations, the predictions correspond well with the AMSR-AVHRR measurements. Of the simulations, ROMS performs best when it is forced by the YSU setup of WRF, with a mean bias of +0.52 °C and mean RMSE of 1.55 °C. ROMS performs worst in the shallow coastal areas, however in the interior of the basin predictions are very good.

### 3.2.3 Evaporation

Evaporation calculated in ROMS can be compared with the values found in the literature which have a range of 918 - 1039 mm/year. The simulation forced by the YSU setup of WRF has more evaporation than using the MYJ setup which gives...
<table>
<thead>
<tr>
<th>Simulation</th>
<th>Evaporation (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ROMS with MYJ WRF</td>
<td>1054</td>
</tr>
<tr>
<td>ROMS with YSU WRF</td>
<td>1136</td>
</tr>
<tr>
<td>ROMS with ERA-Interim</td>
<td>1053</td>
</tr>
</tbody>
</table>

Table 3.7: Total evaporation from the ROMS simulations over the Caspian (ignoring the KBG) between Dec 2007 and Dec 2008

the same amount as with ERA-Interim forcing (Table 3.7). The YSU WRF forced simulation predicts an evaporation total closest to the expected value (1160 mm), but all of the simulations fall in the range of historical totals.

These evaporation predictions can be compared to those from the WRF simulations in Table 3.3. It can be seen that ROMS and WRF predict similar evaporation under the MYJ setup, with ROMS giving 44 mm less during the year. However with the YSU setup, WRF has 225 mm more evaporation than ROMS. Given the difference in SST fields between the ROMS run and associated WRF simulation (using ERA-Interim SST), it cannot be concluded whether the difference in evaporation is due to SST differences or the bulk flux algorithm employed. Despite this, the ROMS simulations give values for total evaporation closer to what is expected from the literature than the WRF.

### 3.3 SWAN Wave Model

The SWAN wave model is validated by comparing simulated significant wave heights against measurements at three stations (Shah Deniz, East Azeri and DWG). A SWAN simulation is performed between 1st December 2007 and 1st December 2008 using wind forcing from the MYJ setup of WRF. Station data is missing for Dec 2007 at both DWG and East Azeri, so these comparisons are made for the first 11 months of 2008, while at Shah Deniz waves are compared over the whole year.

Figure 3.27 shows qq plots of significant wave height from each of the three stations. At each station the wave heights are higher in the model than observations with biases of +19.4% (DWG), +33.8% (SD) and +16.5% (EA). The higher bias seen at Shah Deniz than the other two stations can, at least partially, be explained by the winds. Comparing the forcing wind field with Quikscat at each of the station locations we see that at DWG and East Azeri the winds are overestimated by 1.5 and 3.3% respectively, while at Shah Deniz the winds are biased 7.5% high. This could therefore, explain why each of the stations sees a positive bias in the simulated wave.

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Figure 3.27: QQ plot of significant wave height from SWAN simulations against measurements at DWG, Shah Deniz and East Azeri between 1st Dec 2007 - 1st Dec 2008.

It is clear from the qq plots that the simulations perform worst for the low wave events. This is particularly noticeable at East Azeri, where below 1 m the waves are overpredicted by about 20%, while above this the simulated distribution corresponds quite well with that of the observations. This is especially important given that observed wave heights are less than 1 m 73% of the time, meaning the model performs worst where it has the largest number of observations. The same is true at the other two stations, where agreement is worst for the smaller wave heights. At DWG, as at East Azeri, the extreme waves are predicted better, however they still show an overprediction.

Figure 3.28 shows a time series of simulated and measured wave heights at East Azeri between 6th July and 24th August 2008. This time is chosen as it contains a few large wave events along with long calm periods. From this, it is evident, as surmised from Figure 3.27, that the low wave events are overpredicted in the model, while the storm events are well represented.

Another reason for the model overprediction of wave heights could be because of station sheltering. The wave heights are measured near oil rigs and therefore the wave field could be affected by the surrounding structures, which might lead to
increased blocking and dissipation of wave energy. This effect would not be captured in the simulations which would therefore predict larger waves than observed. It is not clear however that this would affect the small waves more than the large events, and so is not likely to provide an explanation for the model bias.

Whatever the cause, wave heights are overpredicted at all stations by the model. This positive bias is mainly due to an overprediction of low wave conditions, while the extremes are more accurately represented.

### 3.4 Conclusions

Within this Chapter models of the atmosphere, ocean and waves have been validated for simulations on the Caspian Sea by comparison with measurements. The results from each have been shown to accurately represent the observed behaviour around the Caspian Sea, while certain model short-comings are noted.

The use of two different planetary boundary layer schemes in WRF has been tested, and the resulting wind speeds showed only small differences. WRF is seen to overestimate wind speeds compared with both station measurements and satellite observations. This slight overestimation occurs in all seasons when comparing with
the station measurements, but is greatest in spring and summer when compared
with satellite observations. The direction of winds predicted in the model is in good
agreement with the measurements, except in autumn where the model predicts
too-frequent easterly winds. Wind speeds from WRF are higher than those from
ERA-Interim throughout, and generally provide better agreement with observations,
particularly in coastal areas.

The use of different planetary boundary layer schemes is seen to result in large
differences in evaporation from the Caspian. The YSU scheme predicts much more
evaporation, likely due to the non-local closure producing more vertical transport
in the boundary layer, resulting in a warmer drier surface layer, as seen previously
by Hu et al. [2010]. Comparison with climatological values found in the literature
suggests that this results in too much evaporation. Totals produced using the
MYJ scheme are in line with expectations, as are the qualitative seasonal, spatial,
patterns.

The two microphysics schemes tested in WRF both produce spatial patterns,
and totals, of precipitation which are in broad agreement with the satellite observa-
tions. Intense precipitation is predicted by the model over the south-west corner
of the Caspian in autumn and winter as observed. This precipitation would seem
to be driven by strong, seasonal, evaporation in the South Caspian, and cannot
be explained by increased occurrence of northerly winds pushing moist air from the
Caspian over the Elburz mountains to the south. The precipitation totals are higher
using the double-moment WSM6 microphysics scheme than that of Thompson et al.
[2008], as was seen in the study of Jankov et al. [2009]. The measured precipitation
totals fall in between those predicted by WRF using the two microphysics schemes,
however the values with the Thompson scheme lie closer to the climatology. Of the
combinations of PBL and microphysics schemes tested in WRF, only the MYJ and
Thompson schemes can represent the climatological evaporation minus precipitation
balance, which controls Caspian Sea level, and for this reason that setup of WRF is
generally favoured in the subsequent simulations.

In testing a number of different vertical mixing schemes within ROMS, it is not
clear that any one is able to best represent the observed currents at all depths. The
LMD scheme however has the smallest RMSE and is also the only scheme able to
reproduce the observed plateau in currents in the upper part of the water column.
This scheme is therefore chosen for other simulations of ROMS.

The ROMS ocean model is shown to represent the observed current profile much
more accurately when forcing is provided by WRF rather than ERA-Interim. ERA-
Interim forcing results in currents which are around 50% too weak through the water column, while using WRF leads to a smaller underprediction of around 10-15%. Near the surface, the underestimation of currents is confined to extreme events, the cause of which cannot be directly attributed to the atmospheric forcing, where no such problem is seen at the extremes. Further down the water column all current magnitudes are underestimated (by around 30% at 40 m), which perhaps suggests that the vertical mixing is not transporting sufficient energy down through the water column.

Sea-surface temperatures in ROMS represent the observations well, although there is a warm bias. The skill is worst around the coastlines, where satellite data is least accurate, and in the shallow coastal areas. Bias is similar whether forcing is provided by ERA-Interim or WRF, although ERA forcing reduces RMSE, likely because it uses SST observations as model forcing, which feeds back to more realistic surface temperatures. Cooler SSTs are seen when ROMS is forced by a WRF simulation using the YSU PBL scheme, due to increased evaporation as a result of the drier atmospheric boundary layer in this scheme. The evolution of SSTs throughout the ROMS simulation is in good agreement with the observations, including the cold water caused by summer upwelling on the east coast of the Central Caspian in summer. The ROMS simulations are however noted to heat and cool too slowly, in spring and autumn respectively, in the North Caspian. The cold tongue of water seen in the model along the west coast in winter does not appear in the satellite observations, but is expected from the literature and is likely an example of the satellite’s poor resolution near to coastlines.

Wave prediction with the SWAN model is compared with measurements at three stations, and the model is found to overestimate wave heights at each location. This might be explained by overestimation of the WRF winds used as forcing, particularly that the bias in wind speeds is highest at the station of highest wave height overestimation. Extreme wave events are generally very well represented while the main model deficiency is at low wave height conditions, which is not easily explained by the wind forcing.

The models have therefore been shown capable of simulating the observed conditions over the Caspian Sea, and can be used to further investigate the behaviour of the Caspian and the impact of coupled processes on modelled results.
Chapter 4

Coupled Modelling

In this chapter results of coupled simulations are presented and the effects of this coupling are discussed. Simulations using a combination of the three models previously discussed and validated (WRF, ROMS and SWAN) are run and compared with uncoupled simulations and measurements.

4.1 Coupled Simulations

Coupled simulations are performed using the COAWST system as discussed in Section 2.5. All of the simulations used in this chapter use the MYJ setup of WRF and the LMD mixing scheme within ROMS\(^1\). It is the output from this WRF only run which is used to force simulations of ROMS and SWAN when not coupled with the atmospheric model. All of the coupled simulations use the same initial conditions and, in the case of WRF, external boundary conditions as the uncoupled runs which allows for direct comparison between coupled and uncoupled simulations. The coupling timestep for communication between each set of coupled models is set to 10 minutes.

The first coupled simulation involves coupling ocean and atmosphere where, here, ROMS receives high resolution forcing from WRF, while the SST forcing fields for the atmosphere model are taken from ROMS.

Coupled wave-ocean model simulations are forced with the uncoupled WRF output. The simulations seek to determine the effect of currents on the wave field and as such are performed with different coupling setups. These include a run where no currents are used in the calculation of the wave field and a simulation where cur-

\(^1\)These model setups are discussed in Section 2.2, and the model results in Chapter 3.
rents, calculated with the *Kirby and Chen* [1989] formulation, are used into SWAN. The effect of waves on the current field is not considered here.

A fully coupled simulation with WRF, ROMS and SWAN is performed where there is communication between each set of models, although as with the ROMS-SWAN runs, wave driven currents are not considered. Currents are calculated using the *Kirby and Chen* [1989] formulation for inclusion in SWAN, while WRF uses the *Oost et al.* [2002] parameterisation to calculate wave dependent surface roughness. WRF here takes SST values from ROMS, while ROMS and SWAN both take their atmospheric forcing from WRF.

Further to this fully coupled simulation, a run is performed with the same setup aside from the impact of currents on waves. Here SWAN does not receive currents from ROMS, meaning that the simulation is effectively WRF coupled to ROMS and WRF coupled to SWAN without ROMS-SWAN coupling (denoted WRF-ROMS,WRF-SWAN).

### 4.2 Surface Winds

Within this section, 10 m wind speeds are compared between the uncoupled WRF simulation and the WRF output from the coupled simulations. Direct comparisons are made between the simulations, as well as comparing the output with station wind speed measurements and Quikscat satellite data, as in Section 3.1.1.

#### 4.2.1 Comparison with station winds

Figure 4.1 shows a qq plot of simulated wind speeds against the station measurements from simulations of WRF, coupled WRF-ROMS and fully coupled WRF-ROMS-SWAN. The coupling of WRF and ROMS looks to have little effect on the surface winds, while it can be seen from Table 4.1 that the winds are increased slightly with this coupling. The RMSE is also not greatly effected by coupling with the ocean model.

The fully coupled simulation however shows a clear divergence from the other two runs. Here, at wind speeds above 7 m/s the fully coupled run has much reduced wind speeds relative to either the WRF or WRF-ROMS simulations. This is a correction of the overprediction of strong winds, although the correction is too large leading to a slight underprediction of high wind speeds in the fully coupled run. The bias in simulated station winds is reduced from +10.3% in the coupled WRF-ROMS
Figure 4.1: QQ plot of wind speed from coupled simulations of WRF against station measurements at Central Azeri between 1st Apr - 1st Dec 2008.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Bias (%)</th>
<th>RMSE (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>WRF</td>
<td>+9.6</td>
<td>2.83</td>
</tr>
<tr>
<td>WRF-ROMS</td>
<td>+10.3</td>
<td>2.85</td>
</tr>
<tr>
<td>WRF-ROMS-SWAN</td>
<td>+5.7</td>
<td>2.72</td>
</tr>
<tr>
<td>WRF-ROMS,WRF-SWAN</td>
<td>+7.3</td>
<td>2.76</td>
</tr>
</tbody>
</table>

Table 4.1: Surface wind comparison between simulations and measurements at Central Azeri between 1st Apr - 1st Dec 2008.
Figure 4.2: QQ plot of wind speed from coupled simulations of WRF, ROMS and SWAN, with and without the effect of currents on waves, against station measurements at Central Azeri between 1st Apr - 1st Dec 2008.

run to +5.7\% in the fully coupled simulation. This can be attributed directly to the effect of waves, whereby the waves increase roughness at high wind speeds, which in turn reduces the surface winds. In addition to a reduced bias, the RMSE is also reduced in the fully coupled simulation, suggesting that the wave coupling improves surface wind prediction at the station.

Surface winds are compared between the fully coupled run and the equivalent where SWAN isn’t affected by currents, Figure 4.2. It is seen here that the winds above 7 m/s are stronger in WRF-ROMS,WRF-SWAN than in the fully coupled run. This is in the range where the wave dependent roughness has a big impact, so the difference in the waves must be considered. As will be seen later in Section 4.3 the wave heights are larger in the WRF-ROMS,WRF-SWAN run than the fully coupled simulation, from which we might expect increased roughness and hence decreased wind speeds. However as discussed in Section 1.1.1 when currents follow waves, the waves travel more quickly and lengthen. This means that in the fully coupled simulation, even though the waves are smaller, they might give larger roughnesses given the \textit{Oost et al.} [2002] roughness parameterisation is proportional to wavelength.

The high wind events are better predicted in the WRF-ROMS,WRF-SWAN simulation than WRF-ROMS-SWAN, and lie closer to the measured distribution.
However the mean bias is smaller in the fully coupled simulation at 5.7% compared with 7.3% due a cancellation of the positive bias at low wind speeds by the underprediction at large winds, which is not present in the WRF-ROMS,WRF-SWAN results.

4.2.2 Wind Field Comparison

The surface wind fields from the coupled simulations and WRF are compared in Figure 4.3, along with the differences between them. The winds are slightly increased through coupling of WRF and ROMS, where the increases are concentrated along the west coast and the south east corner of the Caspian. These differences are likely due to differences in the SST field, where increased shear in temperature leads to increased wind speeds. The mechanism for this process is that the SST gradient leads to a corresponding temperature and pressure gradient in the atmospheric boundary layer; geostrophic balance then leads to flow along isotherms. The coupled WRF-ROMS simulation would be expected to have larger shears in temperature given the higher resolution of the ROMS SST field compared to that of ERA-Interim, possibly explaining this result.

Winds in the fully coupled simulation are lower throughout the Caspian than when wave coupling is not included. The differences however are smallest furthest south, where wind speeds and wave heights are lower. In the North Caspian the winds are decreased by wave coupling despite this being the area of smallest waves, this is likely because the fetch here is short and so waves are ‘young’ giving high roughness relative to their wave height.

As seen in Figure 4.1 the biggest effect of the coupling with SWAN on surface winds in WRF is seen for high wind speed events. For this reason the 90th percentile of winds is plotted at each grid cell in Figure 4.4. It can be seen that coupling WRF with ROMS has little effect, although there is some increase with the coupling along the west coast particularly, as is seen in the mean winds. Coupling with the wave model however leads to large reductions in the strong winds, with decreases of the order of 1 m/s over much of the Caspian. Again the decrease in winds because of wave coupling is less in the South Caspian where the extreme winds are only about 7 m/s.

For the WRF-ROMS,WRF-SWAN simulation it is seen that mean winds are increased in parts of the basin and decreased in others relative to WRF-ROMS-SWAN by the inclusion of ocean-wave coupling. In the middle of the Central Caspian winds
Figure 4.3: Mean wind speed (m/s) from coupled WRF simulations (top) and differences between (bottom) them between 1st Apr - 1st Dec 2008.

Figure 4.4: 90th percentile wind speed (m/s) from coupled WRF simulations (top) and differences between them (bottom) from 1st Apr - 1st Dec 2008.
are higher in WRF-ROMS,WRF-SWAN. This is a region of strong currents relative to the surrounding areas and as such assuming the currents are in the same direction as the waves, wavelengths will be increased here when currents are considered (see Section 1.1.1). This means that even though the waves are higher in WRF-ROMS,WRF-SWAN, they might have lower roughnesses given their decreased wavelength increases smoothness in the Oost et al. [2002] parameterisation. The clearest area of higher winds in the fully coupled simulation is in the very south west corner of the Caspian, here the opposite argument applies as the currents are weak relative to nearby and so waves propagating here will decrease in wavelength, decreasing in steepness and roughness, when the effect of currents on waves is considered.

The 90th percentile wind speeds follow similar local differences, although here the WRF-ROMS,WRF-SWAN simulation has stronger winds over much of the Caspian. Here, as noted previously, this is despite the higher waves and likely because of the increased wavelength when the currents are following the waves, which leads to higher roughness. This effect will be most prominent for high wind speeds as this is when currents will more likely be wind driven and hence in the same direction as the wind, and waves.

4.2.3 Comparison with Quikscat winds

Figure 4.5 shows the mean bias against Quikscat between 1st April and 1st December 2008 of simulations from WRF, WRF-ROMS, WRF-ROMS-SWAN and WRF-ROMS,WRF-SWAN. From this we see that surface wind speed biases are slightly increased by the coupling of WRF with ROMS, which is seen in mean biases increasing from +11.3% to +12.9% (Table 4.2). The fully coupled simulation shows decreased bias against Quikscat throughout, the mean bias reduced to +8.6%. The WRF-ROMS,WRF-SWAN simulation shows slightly increased wind speeds over the fully coupled run, which is manifested in a slightly increased bias of +9.2%. The wave coupling in both WRF-ROMS-SWAN and WRF-ROMS,WRF-SWAN can be seen to decrease the bias in the regions of largest overprediction in WRF-ROMS, whereas elsewhere the reduction is less.

RMSE of the four simulations against Quikscat winds is presented in Figure 4.6. The RMSE is barely changed through the ocean-atmosphere coupling seen in mean values of 2.55 and 2.56 m/s (with and without coupling respectively). The errors however are noticeably reduced through coupling with the wave model, particularly
Figure 4.5: Wind speed bias (m/s) from coupled WRF simulations against Quikscat between 1st Apr - 1st Dec 2008.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Mean Bias (%)</th>
<th>Mean RMSE (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>WRF</td>
<td>+11.3</td>
<td>2.55</td>
</tr>
<tr>
<td>WRF-ROMS</td>
<td>+12.9</td>
<td>2.56</td>
</tr>
<tr>
<td>WRF-ROMS-SWAN</td>
<td>+8.6</td>
<td>2.48</td>
</tr>
<tr>
<td>WRF-ROMS, WRF-SWAN</td>
<td>+9.2</td>
<td>2.49</td>
</tr>
</tbody>
</table>

Table 4.2: Surface wind comparison between simulations and Quikscat data between 1st Apr - 1st Dec 2008.
in the eastern part of the Central Caspian. The mean RMSE is reduced to 2.48 m/s. A clear improvement is therefore seen in the surface winds speeds through coupling of the wave model with the wind model.

The seasonal bias of wind speeds against Quikscat is shown for the WRF and WRF-ROMS-SWAN simulations in Figure 4.7. The WRF simulation shows large positive bias through the basin in spring and the South Caspian in summer. These biases are decreased with the coupling, most clearly that in the summer. In autumn, the WRF simulation provides good agreement with the Quikscat winds, with mean bias of +1.8%. The winds here are not decreased through coupling by the same amount as they are during the other seasons, so the bias is improved to +1.1%. Overall the coupling improves the bias, throughout, in each season (with the exception of winter which hasn’t been considered).

4.3 Wave Heights

Here wave heights are compared between the coupled and uncoupled SWAN simulations. Simulated wave heights are first compared with the station measurements, then the Caspian-wide wave heights are contrasted between the simulations.
Figure 4.7: Seasonal bias in wind speeds from simulations of WRF (a) and WRF-ROMS-SWAN (b) against Quikscat measurements.
4.3.1 Comparison with station wave heights

As in Section 3.3 wave heights are compared with station measurements at Shah Deniz, East Azeri and DWG, between 1st April and 1st December 2008.

Figure 4.8 shows qq plots of significant wave heights at the three stations from simulations of SWAN; SWAN coupled with ROMS; fully coupled WRF-ROMS-SWAN; and the WRF-ROMS,WRF-SWAN run.

At each of the stations it is clear that coupling SWAN with ROMS leads to a dramatic reduction in wave heights. This is as expected because in the Caspian currents are generally wind driven and so in the same direction as waves, which leads to a reduction in wave heights (see Section 1.1.1). Bias and RMSE of all the simulations are presented in Table 4.3, where at each station the bias is decreased
<table>
<thead>
<tr>
<th>Simulation</th>
<th>DWG Bias (%)</th>
<th>RMSE (m)</th>
<th>East Azeri Bias (%)</th>
<th>RMSE (m)</th>
<th>Shah Deniz Bias (%)</th>
<th>RMSE (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SWAN</td>
<td>+25.8</td>
<td>0.44</td>
<td>+24.6</td>
<td>0.43</td>
<td>+42.7</td>
<td>0.44</td>
</tr>
<tr>
<td>ROMS-SWAN</td>
<td>+4.9</td>
<td>0.48</td>
<td>+5.0</td>
<td>0.47</td>
<td>+24.0</td>
<td>0.37</td>
</tr>
<tr>
<td>WRF-ROMS-SWAN</td>
<td>-12.4</td>
<td>0.50</td>
<td>-10.9</td>
<td>0.49</td>
<td>+10.0</td>
<td>0.34</td>
</tr>
<tr>
<td>WRF-ROMS,WRF-SWAN</td>
<td>+11.0</td>
<td>0.37</td>
<td>+9.8</td>
<td>0.37</td>
<td>+28.3</td>
<td>0.36</td>
</tr>
</tbody>
</table>

Table 4.3: Comparison of modelled and measured wave heights between 1st April and 1st December 2008 at DWG, East Azeri and Shah Deniz

by around 20% through this coupling, bringing the mean wave heights much closer to the observations. However from Figure 4.8 it is seen that this reduction of wave heights means that the larger waves are now heavily underpredicted at both DWG and East Azeri, which leads to an increased RMSE. The effect of the coupling is greatest for the large waves, as these events correspond to strong winds, driving the strongest currents, which will tend to be in the wave direction. At lower wind speeds (and hence wave heights) the currents may not be in the wave direction and so in some instances act to increase wave heights, and others decrease them.

When coupling with WRF is added into this, in the fully coupled simulation, wave heights are further reduced. This is what we would expect as the wind speeds are reduced by this coupling, which in turn leads to smaller waves. From Figure 4.8 it can be seen that this further reduces the wave heights at the extremes, such that the large wave events are underpredicted by around 50% at DWG and East Azeri and also underpredicted at Shah Deniz. At DWG and East Azeri the bias is now negative showing mean underprediction by the model, and the RMSE is greater than in either the SWAN or SWAN-ROMS runs. At Shah Deniz the RMSE and bias are both improved through the coupling of WRF, ROMS and SWAN.

The final coupled simulation is that of WRF-ROMS,WRF-SWAN which is as the fully coupled run, but without the effect of currents on waves. It can be seen in Figure 4.8 that the wave heights in this simulation are higher than in the fully coupled run. The reason for this is the same as that discussed for the difference in wave heights between SWAN and coupled ROMS-SWAN. It is also important to note that the wave heights are lower than in the SWAN run due to the reduction in wind speed due to the increased wave roughness. At each of the stations the distribution of wave heights predicted by the WRF-ROMS,WRF-SWAN simulation is closest to the observed at mid to high waves. At each station the bias is improved over that from SWAN, and slightly increased over that in SWAN-ROMS. At both DWG and East Azeri the RMSE is much smaller in the WRF-ROMS,WRF-SWAN run than any of the other simulations considered, while at Shah Deniz it is better.
than for SWAN but slightly higher than from WRF-ROMS-SWAN.

4.3.2 Wave Field Comparison

The wave fields over the whole of the Caspian are compared between the coupled simulations to give an idea of where the waves are most affected by coupled processes.

Figure 4.9 shows the mean wave height over the period 1st April to 1st December 2008. From this it is obvious, as at the stations, that the mean waves are reduced through coupling with ROMS, and further through addition of WRF coupling. With wave-ocean coupling, the wave heights are reduced throughout, although the reduction is less in the North Caspian and shallow areas where the currents are weakest. The fully coupled run has much smaller mean waves in the Central Caspian than ROMS-SWAN. This is in the area of highest mean waves and so can be attributed to the reduction in wind speeds because of the wave roughness. The WRF-ROMS,WRF-SWAN simulation has higher waves throughout than the fully coupled run, as the wave-current coupling which is shown to decrease wave heights, is not included. This simulation however has smaller waves over the whole of the Caspian than the SWAN simulation due to the decrease in wind speeds through wave-atmosphere coupling.

The high wave events are affected in much the same way as the mean wave heights as can be seen from Figure 4.10 which shows the 90\textsuperscript{th} percentile wave heights.
Figure 4.10: 90th percentile significant wave height (m) from coupled SWAN simulations (top) and differences between simulations (bottom) from 1st Apr - 1st Dec 2008.

4.4 Currents

4.4.1 Station currents

The mean current profiles from the coupled simulations over the period 1st April to 13th September 2008 are compared with station measurements in Figure 4.11. Each of the simulations produces a mean current profile following the same shape, where currents are underpredicted down to around 60 m depth, and below this currents are stronger than the measurements. All of the simulations have the dip in mean currents from 30 to 70 m which is not seen in the measurements (as noted in Section 3.2.1), and a plateau between 10 and 20 m which mimics that in the measurements slightly further down the water column.

The coupled WRF-ROMS output has slightly weakened currents in the top 5 m or so, but stronger currents from there to near the bottom. The mean currents from the fully coupled WRF-ROMS-SWAN are higher than WRF-ROMS throughout the column, particularly in the upper 20 m.

Table 4.4 shows the mean bias and RMSE of the currents with the station measurements at depths of 6, 20 and 70 m. This shows that the currents are highest in the fully coupled simulations as seen from the mean profiles. However it is also seen here that the RMSE is lower, at each of the depths considered, in the ROMS
Figure 4.11: Mean current magnitude profiles from coupled simulations and measurements at Shah Deniz between 1st Apr - 1st Dec 2008.

<table>
<thead>
<tr>
<th>Simulation setup</th>
<th>6m</th>
<th>20m</th>
<th>70m</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Bias (%)</td>
<td>RMSE (m/s)</td>
<td>Bias (%)</td>
</tr>
<tr>
<td>ROMS</td>
<td>-16.6</td>
<td>0.098</td>
<td>-13.2</td>
</tr>
<tr>
<td>WRF-ROMS</td>
<td>-15.5</td>
<td>0.099</td>
<td>-10.4</td>
</tr>
<tr>
<td>WRF-ROMS-SWAN</td>
<td>-10.6</td>
<td>0.107</td>
<td>-7.1</td>
</tr>
</tbody>
</table>

Table 4.4: Comparison of modelled currents with station measurements at depths of 6, 20 and 70 m between 1st Apr - 13th Sep 2008

simulation than any of the coupled simulations, while increasing the level of coupling increases the errors slightly.

At the Shah Deniz station coupling ROMS to WRF is seen to slightly increase currents through most of the water column, which is likely due to the increase in wind speeds in the coupled simulation. The inclusion of SWAN coupling further increases currents throughout the water column, despite the reduced wind speeds seen in this simulation. This is not due to wave driven currents as these have not been included in our simulation. Instead the cause is believed to be the increased roughness due to waves, leading to increased surface stress and therefore more energy being input to the ocean.
Sea-surface temperatures from ROMS and the coupled simulations are compared with satellite measurements from AMSR-A VHRR. The SSTs from the ocean model are compared between 1st April and 1st December 2008.

The bias between modelled and satellite SSTs is shown in Figure 4.12. The most obvious effect of coupling WRF to ROMS is that the SSTs are reduced in the North Caspian, where the bias here is much improved. The SSTs are also decreased across the middle of the Central Caspian, again correcting an overprediction in ROMS. Excluding the KBG, coupling WRF with ROMS reduces the mean bias from +1.00 to +0.72 °C (see Table 4.5).

Figure 4.13 shows the RMSE, and it can be seen that the coupled WRF-ROMS simulation has lower errors in the North Caspian and along the east coast of the Central Caspian. RMSE is however increased in the south east corner in the coupled run. The mean RMSE is slightly reduced by coupling ocean and atmosphere from 1.63 to 1.57 °C.

**4.5 Sea Surface Temperatures**

Sea-surface temperatures from ROMS and the coupled simulations are compared with satellite measurements from AMSR-A VHRR. The SSTs from the ocean model are compared between 1st April and 1st December 2008.

The bias between modelled and satellite SSTs is shown in Figure 4.12. The most obvious effect of coupling WRF to ROMS is that the SSTs are reduced in the North Caspian, where the bias here is much improved. The SSTs are also decreased across the middle of the Central Caspian, again correcting an overprediction in ROMS. Excluding the KBG, coupling WRF with ROMS reduces the mean bias from +1.00 to +0.72 °C (see Table 4.5).

Figure 4.13 shows the RMSE, and it can be seen that the coupled WRF-ROMS simulation has lower errors in the North Caspian and along the east coast of the Central Caspian. RMSE is however increased in the south east corner in the coupled run. The mean RMSE is slightly reduced by coupling ocean and atmosphere from 1.63 to 1.57 °C.
Table 4.5: Comparison of sea-surface temperatures from the coupled simulations with AMSR-A VHRR between 1st Apr - 1st Dec 2008

<table>
<thead>
<tr>
<th>Simulation</th>
<th>with KBG</th>
<th>without KBG</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean Bias (°C)</td>
<td>Mean RMSE (°C)</td>
</tr>
<tr>
<td>ROMS</td>
<td>+0.97</td>
<td>1.68</td>
</tr>
<tr>
<td>WRF-ROMS</td>
<td>+0.69</td>
<td>1.67</td>
</tr>
<tr>
<td>WRF-ROMS-SWAN</td>
<td>+0.54</td>
<td>1.64</td>
</tr>
</tbody>
</table>

Figure 4.13: RMSE of SST (°C) from coupled ROMS simulations against AMSR-AVHRR between 1st Apr - 1st Dec 2008.
Comparing the evolution of SST more closely between the ROMS and WRF-ROMS simulations, temperatures are shown for selected dates in Figure 4.14. During spring both simulations are too warm in the South Caspian, where WRF-ROMS is hotter (Figure 4.14a). Into summer, the differences through coupling are small and both simulations represent the observations accurately (Figure 4.14b). Through the cooling period at the end of the summer ROMS cools too slowly, in the North Caspian in particular, as seen in Section 3.2.2. This is improved by coupling, where the cooling is quicker and in better agreement with the satellite data (Figure 4.14c). Towards the end of the simulations, as winter begins, both simulations are close to the observed, while WRF-ROMS is slightly better in the South Caspian, where it has cooler SSTs (Figure 4.14d). Here it is clear that ROMS is better at predicting SST in the KBG, which is likely due to its poor representation in ROMS feeding back into the atmosphere when coupled to WRF.

One cause of improvement in SSTs through coupling is likely due to improvement in near-surface temperature forcing. For the ROMS simulation, surface forcing was taken from WRF output, however these surface fields will obviously differ from those in the coupled WRF-ROMS run. The key here is the difference between the SSTs in ROMS and ERA-Interim - when WRF is run uncoupled it takes SST fields from ERA-Interim, whereas in the coupled simulations SSTs come from ROMS. Higher sea-surface temperatures in the boundary forcing will lead to increased surface air temperatures, which when used as forcing for ROMS will give higher SSTs. We saw in Section 3.2.2 that the annual mean SST was lower in ERA-Interim than in the WRF forced ROMS simulation, however there were large spatial variations. This is also true for the coupled simulation period, Figure 4.15 shows the mean difference in SST between ERA-Interim and WRF-ROMS from April and December 2008. The temperatures in the North Caspian are lower in WRF-ROMS than ERA-Interim, as they are in the south of the Central Caspian, which in turn would lead to lower WRF air temperatures near the surface and a positive feedback reducing the SSTs in this area. This can explain the reduction in SSTs in North and south Central Caspian in the coupled WRF-ROMS simulation relative to those in WRF forced ROMS.

The addition of wave coupling is seen to decrease the bias of SSTs in the Central Caspian, while the RMSE is also reduced in this area. Taking an average over the whole Caspian, the mean bias is +0.57 °C and mean RMSE is 1.54 °C.

The decreased SSTs with wave coupling are harder to explain. Evaporation is increased by 6 mm over the WRF-ROMS simulation, which is not enough to account
Figure 4.14: Snapshots of SST (°C) from ROMS and WRF-ROMS simulations and AMSR-AVHRR. From: 1st May 2008 (a), 13th June 2008 (b), 21st Sep 2008 (c) and 29th Nov 2008 (d).
for the drop in SST. Rather it is seen that there is a decrease in both shortwave and downward longwave radiation at the water surface in the simulation with wave coupling, which is likely to account for the reduced surface temperatures. The reason for the decrease in downward radiation in this simulation is unclear.

Table 4.5 shows that the best agreement with the satellite SSTs comes from the coupling of WRF-ROMS-SWAN. WRF coupling reduces an overprediction of SSTs by 0.28 °C with a small corresponding decrease in RMSE. However wave coupling reduces SST by a further 0.15 °C, giving the smallest overall bias.

4.6 Precipitation and Evaporation

In this section the impact of the various levels of model coupling on the precipitation and evaporation fields and the overall water budget is considered. First, the model output fields are contrasted between the simulations, then these are compared with the precipitation observations from TRMM-GPCC. As the simulation period for the coupled runs is less than a year, direct comparison cannot be made against the reported climatological mean precipitation and evaporation values, however some conclusions can be drawn on the impact of coupling by considering the biases in the year long WRF simulations.
Figure 4.16: Total evaporation (mm) from coupled WRF simulations (top) and differences between the simulations (bottom) from 1st Apr - 1st Dec 2008.

### 4.6.1 Evaporation

Figure 4.16 shows a comparison of evaporation in the coupled simulations. The ocean-atmosphere coupled simulation has increased evaporation over most of the water surface. The west coast has increased evaporation particularly in the South Caspian, as do the shallow areas in the south east corner, North Caspian and KBG. Overall evaporation is increased from 829 mm to 906 mm (shown in Table 4.6), while that ignoring the KBG is 8.7% higher with the coupling. The inclusion of wave coupling has a much smaller effect on the evaporation fields, and total evaporation is increased by less than 1% over the WRF-ROMS run.

The most obvious explanation for the increase in evaporation in the coupled ocean-atmosphere run is that the sea-surface temperature fields received by the atmosphere are hotter than when WRF is run alone. The WRF run takes SST fields from ERA-Interim, while those in the coupled run are the SST output from ROMS within WRF-ROMS. It can be seen from Figure 4.15 that the SSTs are generally higher in WRF-ROMS than ERA-Interim. The differences are clearest along the west and south coasts and in the south east corner of the Caspian where the
coupled simulation is over 1 °C warmer. These areas correspond strongly with the areas of increased evaporation in the coupled model and this evaporation difference can therefore be attributed to increased sea surface temperatures.

In Figure 4.17 seasonal plots of total evaporation are shown for simulations of WRF and WRF-ROMS-SWAN. From this it can be seen that the areas of intense evaporation are increased through the coupling (for instance the North Caspian in summer), while those with little evaporation see even less (Central and South Caspian in spring and summer). This can at least partly be explained by the increased resolution of SST in the coupled simulation. The SST in WRF comes from ERA-Interim which is spatially smooth, this means, for instance, that the gradient of SST (and hence evaporation) in summer between North and Central Caspian will not be as large. It is also noted that the coupled simulation shows an area of decreased evaporation along the east coast of the Central Caspian in summer, which was expected from the literature, but not seen in WRF. This is because the SST fields in ROMS capture the cold area due to upwelling (Figure 4.14b), which is not seen in ERA-Interim.

Given the differences in evaporation between the coupled and uncoupled simulations can be primarily attributed to the difference in SST forcing, a simulation of WRF is run with SST fields from ROMS. The aim here is to determine the relative importance of coupling rather than using ROMS SST forcing as opposed to ERA-Interim. For this, the SST forcing in WRF was taken from the output of a ROMS run forced by WRF between April and December 2008. The total evaporation from this simulation is shown in Figure 4.18 along with that from WRF-ROMS. It can be seen here that evaporation is higher in the WRF run forced with ROMS SST than in the coupled simulation. Evaporation is increased throughout but most particularly in the North Caspian; total evaporation is 8.9% higher using ROMS SST than in
Figure 4.17: Seasonal evaporation (mm) from WRF (a) and WRF-ROMS-SWAN (b) simulations
Figure 4.18: Total evaporation (mm) from a coupled WRF-ROMS simulation and WRF using SST output from ROMS between 1st Apr - 1st Dec 2008.

the coupled WRF-ROMS run. This difference can be attributed to the difference in SST between the ROMS and WRF-ROMS simulations, as seen in Figure 4.12 where the temperatures are higher in ROMS, especially in the North Caspian.

This shows that the difference in evaporation between WRF and WRF-ROMS simulations is not solely down to SST forcing coming from different models or datasets. Clearly the atmospheric forcing of the ocean model used to provide SST is itself important. It can be assumed that if simulations of ROMS forced by WRF, and then WRF forced by the ROMS SST, then ROMS forced by this WRF output etc. were performed iteratively, then the SST and evaporation fields would be similar to those from coupled WRF-ROMS. This would be the truest test of the direct effect of coupling the models, however this is not practical and in any case the number of iterations required is not clear. In lieu of this it can be stated that without performing iterative simulations there are significant differences in the output fields of WRF when uncoupled and coupled with ROMS, as shown here for evaporation. Either coupling, or extensive iterative simulations, are required to capture these
feedbacks between the models.

Given that there are no direct measurements of evaporation to compare the output with, it is not possible to conclude whether the coupling improves model performance here. However in Section 3.1.2 it was seen that WRF predicted slightly more evaporation than expected, and as such it might be assumed that WRF over-predicts annual evaporation. In this case the coupling here would not improve model performance, as total evaporation is increased by each form of coupling.

While evaporation in coupled runs employing WRF is calculated in the atmospheric model component, that in ROMS-only simulations is calculated using the ROMS routines. The total evaporation from the WRF forced ROMS simulation run from April to December 2008 is shown in Figure 4.19. This can be compared with the evaporation calculated in the various WRF simulations in Figures 4.16 and 4.18. The evaporation from ROMS appears very similar to that in WRF-ROMS and WRF-ROMS-SWAN, which should not be surprising given the SST fields each come from ROMS simulations (albeit different ROMS simulations given the effect of coupling). Although the fields appear similar, the evaporation from ROMS of 839 mm (over the Caspian) is about 50 mm less than either of the coupled WRF simulations. Comparing the evaporation from ROMS with that from WRF when using ROMS SST forcing (Figure 4.18), it is obvious that the evaporation is much lower throughout in ROMS. This is despite identical SST fields in both simulations, and so the divergence must either be attributed to differences in the atmospheric fields or the method of computing evaporation.

### 4.6.2 Precipitation

The precipitation from simulations of WRF, WRF-ROMS and WRF-ROMS-SWAN is shown in Figure 4.20 along with differences between the outputs. The pattern of precipitation is for the most part unchanged by coupling the models, where the peak areas of rainfall remain over the Caucasus mountains and south west corner of the Caspian Sea. The intensities however are altered, and it is noticeable that the area of intense rainfall in the south west is increased by the coupling of WRF with ROMS, with the maximum increased from around 1100 to 1500 mm, while just to the north of this lies a region of reduced rainfall along most of the south coast of the Caspian. Elsewhere changes are relatively small. Overall the coupling of WRF with ROMS increases the total rainfall by 10%, and by 6.5% over the Caspian (see Table 4.6). When wave coupling is added, precipitation is increased over a large region in
the South Caspian. This leads to a further increase of 6.8% over the domain, and 16.9% over the Caspian.

The increase in precipitation in the WRF-ROMS run over that in WRF is likely due to the increase in evaporation meaning more water vapour is present in the atmosphere. This in turn means that there is more precipitable water, which might result in increased precipitation. The same could be true of the addition of wave coupling, however this only increases evaporation a little and so this probably wouldn’t account for the more dramatic increase in precipitation of 6.8% over the domain. The areas of increased and decreased precipitation also correspond fairly well with local changes in evaporation. For example, evaporation is greatly increased in the south west corner of the Caspian through ocean-atmosphere coupling and this is where rainfall is increased in this simulation, while in the central South Caspian, evaporation is reduced and along the south coast rainfall is decreased. The differences in evaporation caused by wave coupling are more subtle, but generally there is an increase in the South Caspian which leads to a rise in rainfall slightly further south.

The TRMM-GPCC rainfall observations are shown in Figure 4.21 and can be compared with the model results. As in Section 3.1.2 where annual precipitation was compared, the TRMM dataset has lower rainfall than the model in the area
Figure 4.20: Total rainfall (mm) from coupled WRF simulations (top) and the difference between simulations (bottom) from 1st Apr - 1st Dec 2008.
of peak precipitation in the south west of the Caspian. Rainfall over the Central Caspian into the North Caspian is still stronger in the observations than the models, although the general pattern of precipitation is well represented. The spatial differences in precipitation through coupling of the models are not such that differential comparisons can be made with TRMM-GPCC. The only change that is noted is that WRF-ROMS-SWAN has an extended region of increased rainfall in the south west of the Caspian, which is closer to the pattern seen in the observations.

Comparing spatially averaged precipitation totals it can be seen that agreement with the satellite data is improved through coupling (Table 4.6). WRF underpredicts the rainfall seen in the observations, however the total precipitation is increased when ocean and atmosphere are coupled. This still gives an underprediction, which is corrected by wave-ocean-atmosphere coupling. It can also be seen however that the total rainfall in WRF accurately represents that observed when run with ROMS SST forcing.

### 4.6.3 Water budget

In Section 3.1.2 the water balance from WRF was compared with values in the literature and, using the MYJ setup, the simulated $E - P$ lay at the high end of the quoted values. There are no values for comparison for the period April to December,
however the changes to the balance can be considered in light of the annual balance from WRF.

It is seen in Table 4.6 that the total $E - P$ is lowest in the WRF simulation. This might imply that if the coupled simulations were extended to one year, their $E - P$ would far exceed those in the literature, leading to a drop in Caspian sea level. However, most of the annual evaporation occurs during the simulation period of April to December, while only just over half of the precipitation occurs in these months, which means that the $E - P$ cannot be simply extrapolated to 12 months. The total evaporation and precipitation are individually extrapolated to 12 month totals by taking the percentages of precipitation and evaporation in the year-long WRF run occurring between April and December and applying these to the coupled simulations. These values are shown in Table 4.7, where it is seen that the $E - P$ is still increased in the coupled simulations, however by less than over the period of April - December. From this it is obvious that while WRF might agree better with TRMM-GPCC precipitation totals when using ROMS SST forcing, it gives a much larger total $E - P$ than expected and hence probably doesn’t accurately represent the overall water balance. The WRF-ROMS-SWAN simulation however does represent the observed rainfall over the April - December period, while also giving a reasonable $E - P$ value which is not much above that in the literature or calculated from WRF.

Clearly the above extrapolation must be viewed with some caution as it cannot necessarily be assumed that the same percentage of annual precipitation and evaporation will occur between April and December in each simulation. Using the same extrapolation on the TRMM-GPCC data from April to December yields an annual total precipitation of 369 mm compared to the 310 mm observed between December 2007 and December 2008. This shows that the ratio used to calculate yearly precipitation is not universally representative.

4.7 Conclusions

A number of coupled and uncoupled simulations have been considered in terms of their effects on surface winds, currents, waves, sea-surface temperatures and the water budget. No one simulation stands out as an improvement in all areas, rather different properties are improved by various levels of coupling.

Surface wind speeds are slightly increased by the coupling of ocean and atmosphere, due to increased surface temperature shears. A much more significant effect
of the coupling is the reduction of high wind speeds due to wave coupling. This shows that wave dependent roughness is important above around 7 m/s, acting as a feedback to reduce wind speeds. This is seen to improve model skill.

Wave heights are decreased with the inclusion of currents from ROMS in their calculation. This is as expected when waves and currents travel in the direction of wind forcing. This effect reduces wave height bias, but results in an underestimation of large wave events while not affecting the overprediction of smaller waves, so gives an inaccurate representation of the distribution of wave heights. Wave heights are also reduced by the additional coupling of wave and atmosphere models, and subsequent drop in wind speeds. When combined with the reduction due to currents, this tends to reduce wave heights by too much and increases the errors at two of the three stations considered, while giving the best results at the third. Removing the effect of currents on the wave field and only considering the wave-atmosphere coupling gives the best representation of the observed wave heights. This can perhaps be explained by “double-counting” of currents in coupled ROMS-SWAN simulations. The default tuning of SWAN was performed without the use of wind-driven current forcing, and as such includes this implicitly. Therefore the additional, explicit, inclusion of these currents likely means that they are considered twice, reducing the wave heights by too much.

The effect of coupling on the currents seen here is not particularly great. Ocean-atmosphere coupling slightly increases the currents, likely due to the increase in wind speeds. A further increase is seen when wave coupling to the atmosphere is included, despite the associated reduction of winds. This is caused by the increased roughness of the ocean surface leading to more momentum transfer from the atmosphere.
Sea-surface temperature prediction is improved by coupling ocean and atmosphere models, with the mean bias reduced from 1°C to 0.72°C. This change can be attributed to the feedback between SST and surface air temperature. In WRF simulations forced with ROMS SSTs this effect cannot be accounted for, and would not be seen in uncoupled simulations unless a number of iterative simulations of WRF and ROMS were performed. With coupling, the SST and surface atmospheric fields are consistent and rapidly reach an equilibrium state. SST model skill is improved further when wave coupling is added. The improvement through coupling is particularly noticeable in increasing the rate of cooling in the North Caspian in autumn, which is too slow in ROMS.

The changes in SST when ROMS is coupled to WRF result in increased evaporation, which is again increased slightly when wave coupling is added. The increase in evaporation when coupling ocean and atmosphere arises because WRF receives SST fields from ROMS rather than the lower fields of ERA-Interim. The increase in evaporation with model coupling leads to an increase in precipitation, improving agreement with the satellite observations. The best agreement with the expected water budget is found when WRF is run alone, however WRF-ROMS-SWAN gives a good representation while better predicting total rainfall.
Chapter 5

Inertial Currents

While there have been no previous studies of inertial waves on the Caspian Sea, the importance of near-inertial oscillations (NIOs) has been seen to vary greatly across the world’s oceans. Therefore a study of the spatial variation of NIOs within the Caspian is of interest. This work has been published in Farley Nicholls et al. [2012].

5.1 Inertial Currents on the Caspian Sea

Oscillating behaviour of the near-surface currents is observed at Shah Deniz, (Figure 5.1a), which implies wave behavior. Looking more closely at this, the period is of the order of three quarters of a day, from which it can be assumed that the oscillations might be caused by inertial waves, as the local Coriolis period is 18.7 hours. The oscillations are present through most of the record and there is no obvious decay from individual events.

As discussed in Section 1.2 inertial waves lead to circularly rotating currents which can be identified by oscillations in the $u$ and $v$ components, with $v$ leading $u$ by a quarter period. Figures 5.1b and c show that this behaviour can be identified in the measured near surface currents.

The rotary spectral analysis methods of Gonella [1972] can be applied to a vector time series of currents to decompose into clockwise and counterclockwise spectral components. This is of particular use here to produce a frequency spectrum and identify the period of current oscillation. Figure 5.2 shows the measured clockwise and anti-clockwise spectra of the near-surface currents between 1st December 2007 and 13th September 2008. From this there is a very obvious peak in the clockwise spectrum near the Coriolis frequency, while the corresponding anti-clockwise spec-
Figure 5.1: Timeseries of measured current magnitudes (a), west-east current component (b) and north-east current component (c) at Shah Deniz (all currents in m/s and time in days)

trum has no such peak. The inertial peak occurs at 0.052 /hr, compared to the local inertial frequency of 0.053 /hr. Another peak in the spectra occurs at 12 hours, however the cause of this semi-diurnal peak is unclear.

5.2 Validating ROMS

As inertial currents are seen in the measurements at Shah Deniz, a simulation of ROMS is performed to investigate whether this is captured in the model, and the spatial distribution of inertial waves. The simulation period is 1st December 2007 to 1st December 2008 with 3-hourly forcing from WRF. As the goal of the simulations is to look at oscillations with periods of less than a day, currents are output from the model at 10 minute intervals.

Figure 5.3 compares the clockwise rotary spectrum of the simulated currents with the measured data at 6 m depth at Shah Deniz. ROMS results show a near-inertial peak at 0.054 /hr compared to 0.052 /hr in the measurements. While the simulated spectrum appears to represent that measured, a more quantitative comparison is needed.

Three quantitative methods are used to compare the measured and simulated spectra. The first is to perform complex demodulation, as described in Perkins [1976], which gives an amplitude for the inertial current. Second is to calculate the
Figure 5.2: Measured clockwise and anti-clockwise rotary power spectra of currents at 6m at Shah Deniz. The inertial frequency is indicated with a dashed line.

Figure 5.3: Measured and ROMS clockwise rotary power spectra of currents at 6m at Shah Deniz. The inertial frequency is indicated with a dashed line.
Table 5.1: Comparison of measured and simulated near-surface inertial currents 1st Dec 2007 to 13th Sept 2008. Inertial amplitude; percentage of total variance (10 days - 30 minutes) in inertial band (f±10%); and percentage of time inertial amplitude exceeds 5 cm/s

<table>
<thead>
<tr>
<th></th>
<th>Depth</th>
<th>Measured</th>
<th>ROMS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amplitude (cm/s)</td>
<td>6m</td>
<td>5.18</td>
<td>5.27</td>
</tr>
<tr>
<td></td>
<td>15m</td>
<td>4.22</td>
<td>3.86</td>
</tr>
<tr>
<td>Variance (%)</td>
<td>6m</td>
<td>28</td>
<td>30</td>
</tr>
<tr>
<td></td>
<td>15m</td>
<td>26</td>
<td>30</td>
</tr>
<tr>
<td>Exceedance (%)</td>
<td>6m</td>
<td>46</td>
<td>47</td>
</tr>
<tr>
<td></td>
<td>15m</td>
<td>32</td>
<td>26</td>
</tr>
</tbody>
</table>

percentage of the total current variance in the inertial band, where this is defined here as the percentage of variance between 10 days and 30 minutes at frequencies f±10%. Finally the percentage of time the inertial current (as calculated by complex demodulation) exceeds 5 cm/s is used as a measure of how often events occur.

Table 5.1 compares the above quantities between the measurements and ROMS results at depths of 6 and 15 m as a representation of the near-surface behaviour. The mean inertial amplitude is around 5 cm/s and exceedance of 5 cm/s happens nearly 50% of the time at 6 m. The agreement between model and measurements is good, with a slight underprediction at 15 m, while the variance of around 30% is well represented. This shows that the model accurately represents the observed near-inertial motions and next the analysis can be extended across the Caspian.

5.3 Spatial Structure of Inertial Currents

From the model output, the surface clockwise rotary spectrum can be calculated at each grid point. For each location we can identify if a peak exists (at the 95% confidence level) in a band of f±10%.

Figure 5.4 shows the peak surface near-inertial period in the simulations (a) and the local Coriolis period for comparison (b). As expected, the period decreases with latitude, and follows closely the local inertial period. On average the model has a frequency 0.33% above the Coriolis frequency. The peak period is only plotted in Figure 5.4a when the spectral peak is significant at the 95% confidence level, so it is also noted that near-inertial oscillations are significant over much of the Caspian Sea. The areas where no significant peak is found correspond strongly to regions of shallow water with depths less than approximately 20 m.
Figure 5.4: (a) Simulated peak near-inertial period (hours) where significant at 95% level, with 20m isobath, and (b) corresponding coriolis period.
Complex demodulation is performed to find the amplitude at the local inertial period at 15 m depth for each model grid point. This is done in direct comparison with the global study of Chaigneau et al. [2008]. Figure 5.5 shows the annual mean amplitudes at 15 m, and selected depth contours. The amplitudes have large spatial variability and broadly follow the depth contours of the Caspian, with higher amplitudes at correspondingly larger local depths. It can also be seen that amplitude increases towards the center of the basin and so may depend on distance from the coast. From this it can be seen that, in the South Caspian in particular, amplitudes increase rapidly with offshore distance in the western part but more slowly in the eastern part. Mean inertial amplitudes of over 10 cm/s are seen over large parts of the interior of Caspian, with a maximum mean amplitude of 14 cm/s.

5.4 Depth Dependence

From Figure 5.5 the amplitudes seem to increase with water depth. To investigate further the dependence of inertial amplitude on water depth, each grid cell is binned according to the local water depth and the mean inertial amplitude amongst those

Figure 5.5: Simulated mean inertial amplitude (m/s) at 15m depth, with isobaths at 50, 100, 200, 400, 600 and 700m.
Figure 5.6: Seasonal mean inertial amplitude against water depth at 15m with 5m depth bins.

points is found. Figure 5.6 shows this mean inertial amplitude against depth for each season. The amplitudes show a rapid increase with depth up to nearly 100 m and then a slower increase with depth above this. In winter there is no obvious cutoff depth. The discontinuity at 800m in Figure 5.6 is because depths higher than this are only found in the South Caspian, whereas the points below this are from both the Central and South Caspian. This causes a discontinuity because the Central and South Caspian basins have slightly different relationships between depth and amplitude, and so up to 800m the trend is an average from both basins, while above this the trend is from fewer points and only those in the South Caspian.

Figure 5.6 also shows that the winter amplitudes are around half of those in summer, while in spring and autumn the amplitudes are slightly less than in summer. This seasonality is generally consistent with, but larger than, reported in previous studies (e.g. Park, 2005; Chaigneau et al., 2008).

5.4.1 Flat bottom experiment

Nothing in the literature would explain the dependence of inertial amplitudes on water depth. So to investigate this, a new simulation was performed to look at whether this model result is robust.

Within the model, depths were capped at a maximum of 300 m. The initialisation of the model was as described in Section 2.3, however the spin-up time was reduced to 2 years of WRF forcing. A simulation was then run as before, with the same atmospheric forcing.
Figure 5.7 shows the annual mean near-inertial amplitudes. The amplitudes are very similar to those in Figure 5.5, which implies that changing the depths has little effect. This is further seen in Figure 5.8 where the inertial amplitudes are binned in terms of the original water depth and which seems to show amplitudes increasing with the water depth. This is obviously a spurious effect as there are no depths above 300 m in this model simulation. This therefore shows that the apparent depth dependence of near-inertial amplitudes seen in the original simulation is not a physically robust result.

5.5 Dependence on Offshore Distance

It has been shown that the pattern of near-inertial oscillations cannot be explained by water depths, so now the dependence on distance from the coast is investigated.

Similar to before, inertial amplitudes are binned according to distance from the nearest coastline. Figure 5.9 shows that the amplitudes increase with distance out to 130 km from the coastline, while at further distances there are fewer data points and the trend is less clear. It is also clear from Figure 5.9 that the relationship between
Inertial amplitude and distance is unchanged in the experiment where depths are limited to 300 m. This suggests that this model relationship is not dependent on the bathymetry of the basin.

In fact, the trend of NIO increase with distance is stronger when distances are calculated from the 50 m isobath (Figure 5.10), with smaller errors in each bin. This result agrees with the measurements of Shearman [2005]. There, a linear relationship is observed between inertial kinetic energy and distance from the New England coastline, however the point from which the distances are calculated is in fact not the coastline but a latitude which corresponds to a depth of roughly 40-60 m.

The results from the simulation show an increase in inertial kinetic energy of 0.4 cm$^2$ s$^{-2}$ km$^{-1}$ from the coastline, while this increases to 0.8 cm$^2$ s$^{-2}$ km$^{-1}$ with distance from the 50 m isobath. This is in broad agreement with the measured value on the New England shelf from Shearman [2005] of 0.8 ± 0.2 cm$^2$ s$^{-2}$ km$^{-1}$.

As expected from Figure 5.5, amplitudes in the west of the South Caspian increase more quickly with distance from the shore than in the east (Figure 5.11a). However, when the distance is measured from the 50 m isobath, the rate of increase in both regions is similar as can be seen in Figure 5.11b. This provides more evidence that distance controls inertial amplitude, and that this relationship is strongest in terms of distance not from the coastline, but from a depth contour of around 50 m.
Figure 5.9: Mean inertial amplitude at 15m against offshore distance, with 1km offshore distance bins. Results from realistic (blue) and depth limited (red) simulations.

Figure 5.10: Relationship of inertial amplitude against distance from coastline and from 50m isobath
Figure 5.11: Increase in inertial amplitude with distance from the coastline (a), and the 50m isobath (b) in the South Caspian, with separate trends for the western (blue) and eastern (green) portions of the basin.

5.6 Other possible mechanisms?

NIOs are known to be a function of mixed layer depth (MLD) and high-frequency wind speed. Given that it is seen that near-inertial amplitudes increase with distance from a coastline, it is important to look at whether this is a function of either wind forcing or MLDs.

5.6.1 Wind speeds

It is clear from Figure 4.3 (in the previous chapter) that the wind speeds are generally not stronger in the regions of large NIOs. Wind speeds do however increase with offshore distance, as would be expected given the decreased drag over water, but this effect is only seen out to about 80 km (Figure 5.12). Given this, it would seem that the increase in NIO amplitude to 130 km from the coast cannot be explained by a corresponding increase in wind speed.

However looking at each basin individually, the wind speed does continue to increase with distance from the coast, right to the center of the basin, as shown in Figure 5.13b for the Central Caspian. The relationship between inertial amplitude and wind speed can then be looked at in Figure 5.13c where, as expected, an increase
in NIOs is found with increasing wind speed. The increase is small for winds between 3.5 and 6.5 m/s, and then rises sharply. Taking the maximum rate of increase of NIO amplitude with wind speed gives a value of 0.24 (this is dimensionless). Then combined with the observed rate of increase of wind speed with distance from the coast of 0.01 ms$^{-1}$km$^{-1}$, this would imply an increase of inertial amplitude of 0.0024 ms$^{-1}$km$^{-1}$. However the observed increase in amplitude with distance in the Central Caspian (seen in Figure 5.13a) is around 0.008 ms$^{-1}$km$^{-1}$. Given that the observed increase is markedly greater than that which could be caused directly by stronger winds, it can be concluded that the increase in NIOs with distance from the coastline is not caused by spatial variation in the wind field.

### 5.6.2 Mixed Layer Depth

Mixed layer depths can be calculated from the model output by finding the depth at which the temperature varies by more than 0.5 °C from that at the surface (Levitus, 1982). The MLD is calculated for each day at each model grid cell, so it can be compared with inertial amplitude. Amplitudes of NIOs are found to decrease with model MLD as expected.

MLD tends to increase slightly with offshore distance (Figure 5.14) which should lead to a decrease in inertial amplitude with offshore distance. The increase in MLD with distance is likely due to the increase in wind speeds, but also, water depths increase with distance, and so it is possible to have deeper mixed layers further from the coast. This implies that the increase in NIOs with distance cannot be explained by spatial variation of MLD.
Figure 5.13: (a) Mean inertial amplitude against distance from the coastline; (b) mean wind speed against distance from the coastline; (c) mean inertial amplitude against mean wind speed. All for the Central Caspian.

Figure 5.14: Dependence of mixed layer depth on distance from the coastline in the Central and South Caspian.
5.7 Mechanism for distance control of NIOs

Previous studies in the coastal ocean (e.g., Shearman, 2005; Jarosz et al., 2007) and in large lakes (e.g., Rao and Murthy, 2001; Zhu et al., 2001) have shown through observations and modelling studies that NIOs increase away from the shore.

Kundu et al. [1983], Millot and Crepon [1981] and Pettigrew [1981] all study this phenomenon using simplified models. Each involves a model domain with a flat bottom and a coastal wall at \( x=0 \), while there is no variation in the \( y \)-direction. Coastal inhibition means that the eastward component of the velocity must be zero at \( x=0 \).

Following the method in Kundu et al. [1983] who start with the two-dimensional equations of motion, taking the ocean to extend in the negative-\( x \) direction:

\[
\frac{\partial u}{\partial t} - fv = -\frac{\partial p}{\partial x} + \frac{\partial}{\partial z}\left( \nu \frac{\partial u}{\partial z} \right) \tag{5.1}
\]

\[
\frac{\partial v}{\partial t} + fu = \frac{\partial}{\partial z}\left( \nu \frac{\partial v}{\partial z} \right) \tag{5.2}
\]

\[
\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0 \tag{5.3}
\]

\[
\frac{\partial \rho}{\partial t} - \frac{N^2 w}{g} = \frac{\partial^2}{\partial z^2} (K\rho) \tag{5.4}
\]

\[
\frac{\partial p}{\partial z} = -\rho g \tag{5.5}
\]

where \( u \) and \( v \) are the eastward and northward velocities in the positive \( x \) and \( y \) directions, while \( w \) is the upward velocity in the \( z \) direction (\( z=0 \) at the surface). \( N \) is the stratification, \( K \) is the diffusion coefficient and \( \nu \) is the damping coefficient. Pressure, \( p \), and density, \( \rho \), are represented as changes from the background state, while \( g \) is the gravitational acceleration.

The variables \( u, v, p, w \) and \( \rho \) can then be expanded in terms of the vertical normal modes, e.g.:

\[
u = \sum_{n=0}^{\infty} u_n \phi_n \tag{5.6}
\]

where the eigenfunctions, \( \phi_n \), have eigenvalues \( c_n \) representing the wave speed of the mode, and \( c_n \) decreases with \( n \). They apply an alongshore wind stress, \( \tau_y H(t) \),
where $H(t)$ is the Heaviside function (taking a value of 0 for $t<0$ and 1 for $t>0$), to an initial state at rest. The wind stress is then expressed through the coupling coefficient to each wave mode, $\tau_n$, by normalising the modes. Assuming inviscid motion, the velocity can be expressed in the form (Kundu et al., 1983):

\[
 u_n = \frac{\tau_n}{f} (1 - \cos (ft)) - H(t + x/c_n) \frac{\tau_n}{f} [1 - \cos (ft + x_n)] \\
 - H(t + x/c_n) \tau_n x_n \{1 - \cos (ft)\} * \left\{ \frac{J_1(f^2 t^2 - x_n^2)^{1/2}}{(f^2 t^2 - x_n^2)^{1/2}} \right\}
\]  

(5.7)

where the offshore distance, can be expressed as a function of the Rossby radius, $x_n = x/r_n = x f/c_n$, for convenience. Here $J_1$ is the Bessel function and * represents convolution.

The first term in Equation 5.7 is independent of $x$ and consists of a pure inertial response and the Ekman drift, while the second term is a reflected wave at the inertial frequency, travelling at $c_n$ from the coast. The third term, involves a Bessel function which generates oscillations with frequencies greater than $f$, but decreasing towards the inertial frequency as the argument of the Bessel function increases (either with increasing time or moving closer to the shore); again this term involves a wave propagating from the coast at its modal wave speed.

For $t < -x/c_n$, the second and third terms vanish and the only response is the directly forced term. The flow is subsequently altered by the passage of waves originating from the coast: first the barotropic wave, and then the subsequent baroclinic wave modes.

In the coastal region, for small $x/c nt$ (i.e. long after the passage of the wave of a particular mode), the flow takes the form:

\[
 u_n = \frac{\tau_n}{f} \left[1 - e^{x_n} + x_n \left(\frac{2}{\pi ft}\right)^{1/2} \cos(ft + \frac{\pi}{4})\right]
\]

(5.8)

From this, the oscillations with each mode are seen to decay as $t^{-1/2}$. It can also be seen here that the oscillating part of the solution is proportional to $x_n$, and so increases in magnitude away from the coast.

An inertial event can then be considered in 3 phases. The first phase is the forced response to the wind forcing, which is confined to the mixed layer and lasts until the passage of the barotropic wave - generally a fraction of an inertial period. The no-flow condition at the coast generates barotropic and baroclinic waves, which
propagate offshore. The barotropic wave then combines with the initial response to create slab-like oscillations in the mixed layer, and oppositely directed inertial oscillations of smaller amplitude below the mixed layer. The final phase begins upon the passing of the baroclinic wave, travelling much more slowly than the barotropic wave, after which the inertial oscillations begin to decay away.

The barotropic signal at a given point is at the inertial frequency when the baroclinic wave arrives. It is noted by Pettigrew [1981] that, at this time, the baroclinic wave is at a frequency above the inertial frequency, which can result in a beating between the two signals. This can mean that the amplitude of the response originally increases upon the passage of the baroclinic wave, before decaying as the baroclinic frequency decreases to the local inertial frequency and combines destructively with the barotropic and forced responses. The result of this is that the maximum of the inertial oscillation might not occur immediately after the wind forcing, but at some point after the passage of a baroclinic wave from the coastline. Therefore the largest inertial oscillation, for a given wind event, would occur later further from the coastline.

Shearman [2005] applied these simple models to the New England shelf, and found that the model was able to represent the observed increase in amplitude of inertial oscillations away from the coastline. When comparing near-inertial kinetic energy against offshore distance, Shearman [2005] finds the linear trend has a zero-crossing further offshore than the coastline. The latitude at which the zero-crossing occurs corresponds to a depth of roughly 40-60 m, and from here an increase in near-inertial kinetic energy of $0.8 \pm 0.2 \text{ cm}^2 \text{s}^{-2} \text{ km}^{-1}$ is seen.

### 5.8 Wave propagation from coastline

As discussed in Section 5.7, a mechanism has been proposed by Kundu et al. [1983], Millot and Crepon [1981] and Pettigrew [1981] to explain the increase in NIOs from a coastline. One prediction of these idealised models is that for an inertial event, the peak amplitude observed at a given point occurs later further from the coastline.

To look to see if this effect is observed in the model, NIOs are analysed along a slice of the South Caspian around 39°N. The largest NIO event in each month is selected and at each model grid point the timing of the peak inertial amplitude associated with those 12 events is identified. Among the 12 events considered, the mean timing of the peak amplitude is calculated at each grid cell. It can often be difficult to identify the peak associated with a particular event and as such there
is a considerable error in calculating the timing of the peak inertial amplitude. To mitigate for this, the timing of the peak amplitude is averaged over a 10-by-10 grid cell box. In Figure 5.15 the mean peak timing from each box is plotted, and it can be seen that the peak occurs later moving towards the centre of the basin, from both the west and east coasts of the South Caspian. This is also shown in Figure 5.16 where the mean amplitude for the composite of events is plotted for the four boxes moving from the west coast of the South Caspian towards the centre of the basin. This result further supports the hypothesis for the propagation of barotropic and baroclinic waves from a coastline controlling inertial oscillations.

## 5.9 Conclusions

A high resolution Caspian Sea simulation was performed using the ROMS model forced by high spatio-temporal output from the WRF atmosphere model. The model’s near-inertial oscillations are compared with station measurements and found to accurately reproduce the observed spectrum, percentage of variance in the inertial band and near-inertial amplitude. The model shows significant near-inertial oscillations over most of the Caspian basin, with the exception of areas with depths of less than about 20 m; the periods of which correspond closely with the local Cori-
olis frequency. The seasonality shows greatest amplitudes in the summer months, when the mixed layer depths are shallowest, with slightly lower values in autumn and spring, and much reduced amplitudes in the well mixed winter months. The magnitude of the seasonal variability is greater than seen previously on the oceanic scale (e.g. Park, 2005; Chaigneau et al., 2008).

Previous studies (e.g. Shearman, 2005; Jarosz et al., 2007; Rao and Murthy, 2001; Zhu et al., 2001) have suggested that inertial oscillations increase with distance from the coast. The proposed mechanism is the interaction and propagation speed of baroclinic and barotropic waves originating from a coastline (e.g. Kundu et al., 1983; Shearman, 2005). This could explain the pattern of inertial amplitudes seen in our Caspian Sea model where maxima are located centrally in the basin. However, amplitudes of inertial currents also seem to follow depth contours; areas of significant near-inertial motions correspond strongly with the 20 m depth contour. An increase in inertial amplitude with both water depth and distance from the coastline is seen. By performing a simulation where the water depths are limited to 300 m, it is confirmed that depth is not the determining factor for amplitude of NIOs. The strongest relationship observed is a linear dependence of amplitude squared on distance from the 50 m depth contour which corresponds to the measurements of Shearman [2005]. The simulations support the hypothesis of Kundu et al. [1983]
for the mechanism of baroclinic and barotropic wave propagation from a coastline controlling NIOs and leading to larger amplitudes further offshore.
Chapter 6

Discussion and Conclusions

In Chapter 3, the WRF atmosphere, ROMS ocean and SWAN wave models were validated for simulations on the Caspian Sea. Two different planetary boundary layer schemes were employed in WRF, and found to produce similar results for surface winds, which slightly overestimate those measured. The wind speeds from WRF are higher than ERA-Interim throughout, and generally provide better agreement with the measurements. The seasonal trends are such that winds are underpredicted in winter, and the bias is greatest in spring and summer, while the errors tend to be highest near the coastlines, where the accuracy of satellite wind products, used for comparison, is lower.

Large differences are seen in evaporation between simulations of WRF using the different PBL schemes. The YSU scheme predicts much more evaporation, likely due to producing more vertical transport in the boundary layer, as seen previously by Hu et al. [2010]. Over the Caspian, this seems to produce too much evaporation when compared with the climatological mean. When the MYJ planetary boundary layer is used the annual evaporation and the seasonal, local, patterns agree well with the climatology.

Two different microphysics schemes are used within WRF to look at precipitation totals. Both produce totals, and spatial patterns, which are in broad agreement with the satellite observations. The Thompson microphysics scheme produces less precipitation than the WSM6 scheme, which is closer to the reported climatological mean. For this reason the combination of the MYJ planetary boundary layer, and Thompson microphysics schemes is best able to model the observed evaporation minus precipitation balance, which controls changes in the Caspian Sea level, falling between the reported values.
ROMS is shown to represent the observed current profile much better when using forcing from WRF simulations than ERA-Interim data. Using ERA-Interim, currents are about 50% too weak throughout the water column, while those using WRF are around 10-15% low at the Shah Deniz station. A number of different vertical mixing schemes have been tested, and it is found that the LMD scheme has the smallest errors at all the depths considered. Sea-surface temperatures are biased high in ROMS, the skill is worst around the coastlines, where the satellite data is less likely to be accurate. The evolution of SSTs is well captured by ROMS, including the area of cold water on the east coast in summer caused by upwelling. SSTs are lower when forcing with the YSU PBL scheme of WRF is used, rather than when either WRF with the MYJ scheme, or ERA-Interim, is used as forcing, which is attributed to increased evaporation caused by the drier atmospheric boundary layer with the YSU scheme.

Wave prediction with the SWAN model is compared with measurements at three stations and the model is found to overestimate wave heights. This could be explained by the wind fields from WRF being too strong, where the bias of winds is highest at the station of the largest wave height overprediction. The waves are seen to be well represented at the peaks, while it is the smaller wave events for which the model performs worst.

The models are then coupled through the use of the COAWST system, and the impact of this coupling is studied in the second results chapter. The coupling of ocean and atmosphere models is seen to slightly increase surface winds, probably due to increased SST gradients over ERA-Interim. The addition of wave coupling decreases wind speeds above 7 m/s due to increased wave dependent surface roughness. This brings simulation results closer into line with the observations and improves model skill.

Sea-surface temperature prediction is improved by ocean-atmosphere coupling, reducing bias from 1 °C to 0.72 °C. This is due to the feedback between SST and the surface layer in the atmosphere, an effect which cannot be seen when WRF is run with ROMS SSTs as boundary forcing. SST model skill is further improved by the inclusion of wave coupling, reducing SST bias to +0.57 °C, with RMSE also improved.

Evaporation in WRF is increased with ROMS coupling, and then more so with SWAN added. This results in more precipitation in the coupled model, which leads to better agreement with satellite measurements. When extrapolating to annual
totals the simulation of WRF-ROMS-SWAN gives a realistic balance between evaporation and precipitation.

Waves heights are found to be reduced by the inclusion of currents from ROMS, bringing the means closer to the observations, although resulting in an underestimation of large wave events, but not affecting the overestimation of small waves. Waves are also reduced by the additional coupling with the atmospheric model, and the resulting decrease in wind speeds. Overall, the best agreement is found when the wave-atmosphere coupling is considered, but the effect of currents on the waves is ignored. This implies that the coupling needs more work, probably to eliminate “double counting” of currents, which are implicitly included in the formulation of SWAN, and then explicitly added through coupling with ROMS.

Chapter 5 examines near-inertial waves on the Caspian Sea. The ROMS model is shown to be able to accurately represent the observed NIOs at the Shah Deniz station, where the inertial peak is clear in the clockwise rotary spectrum of currents. The model shows significant near-inertial oscillations over most of the Caspian Sea, with the exception of areas where water depths are less than 20 m. The amplitude of the NIOs seems to depend on water depth, but a simulation where the depths are cut-off at 300 m sees little change in inertial amplitudes, thus showing that water depth cannot control the oscillations. The amplitudes in summer are around twice those in winter, which is a larger seasonal variability than seen in previous studies in the global oceans.

Inertial amplitudes are then shown to increase with distance away from the coastline, a result which cannot be explained by variation of wind speeds or mixed layer depths. This result agrees with previous studies, and the simplified models of Kundu et al. [1983] and Shearman [2005], which also predict a delay in arrival of an event’s peak further from the coast - similarly observed in the ROMS simulation.

### 6.1 Future Work

Since much of the work in this thesis was completed, I have been made aware of the availability of temperature and salinity profile measurements in the Caspian Sea. This should allow for future comparison with model profiles, which would allow for greater confidence in the oceanic predictions, given that at present there has been little comparison with observations below the surface layer, and the comparison with currents here was not particularly strong.
It was seen in Chapter 4 that surface wind speeds are increased through coupling of ocean and atmosphere models. While it is suggested that this is caused by an increase in winds due to stronger sea-surface temperature gradients, this was not tested. By performing a simulation of WRF using SST forcing from ROMS, and another simulation where the same SST forcing is spatially smoothed, a comparison of surface winds can be made to determine the effect of SST gradients on winds. It is expected that surface winds will be weaker when the SST fields are smoothed.

It was also found that the coupling of ROMS and WRF leads to an increase in precipitation over the whole domain, and the Caspian specifically. The increase in precipitation could have been caused by increased evaporation from the Caspian Sea in the coupled simulation, however it is not clear whether the precipitation is driven by local evaporation. To test this, a simulation of WRF could be performed where the latent heat flux into the atmosphere from the ocean is suppressed. Comparison of precipitation could then be made between this simulation and a control with evaporation present. If the precipitation is driven by evaporation from the Caspian, then we would expect a drop in precipitation in the simulation with suppressed latent heat. From this, the impact of an increase (or decrease) in evaporation on the water budget of the Caspian could be quantified, by removing the resultant increase (or decrease) in precipitation. This would allow for better estimation of the impact of climate change on Caspian Sea level.

When comparing the peak frequency of near-inertial oscillations between model and observations, the model was found to have a frequency slightly higher than both the local inertial frequency and the measurements. Given that the effective frequency of oscillations is the local Coriolis frequency plus a vorticity term, it can be determined whether the frequency differences are caused by differences in the vorticity. This would be done by studying the vorticity field in the area of the measurements in both the simulations and observed data.

Further idealised experiments could be performed to understand the mechanism of distance control of inertial oscillations. An attempt could be made to identify the barotropic and baroclinic waves, propagating from the coast, associated with a particular inertial event. Then, if the simulation was performed in an idealised domain with a flat bottom and coastal wall, the propagation speed of the waves could be compared with that expected from the theory. If the waves are identified, and their propagation speeds correspond to that expected, then this would provide further proof of the mechanism determining the amplitude of near-inertial oscillations.
Acknowledgements

I would like to thank the National Environmental Research Council for funding this project. Many thanks to Professor Ralf Touni for his guidance, and also to Paul Budgell and various PhD students and Postdocs within the department for their assistance with the models used in this project. Thanks to BP’s Environmental Technology Program and Colin Grant for providing much of the data used in this thesis.

Finally, I would like to express my gratitude to Chris O’Reilly for frequently keeping me up to date with his opinions on:

• Sean Bean
• Pierce Brosnan
• Andrew Castle
• Lesley Cohen/Carlos Tevez
• Paul Daniels
• Didier Drogba
• Bruce Forsyth
• ITV1
• Chris Kamara
• Steve Kean
• Anthony Knott
• Damian Lewis
• Liam Neeson
• Chad Ochocinco
• Sarah Jessica Parker
• Paula Radcliffe
• Anna Ryder Richardson
• Stefanie Sanches
• Andy Schleck
• Rob Schneider
• Tim Tebow
• John Terry
• Hines Ward
• Wills & Kate

and to Cathy Ansell for helping me to keep track of the names on this list.
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