

# The Impacts of Ocean-Atmosphere Coupling Near Ocean Mesoscale Fronts

by

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### **Author's Declaration**

I hereby declare that I am the sole author of this thesis. This is a true copy of the thesis, including any required final revisions, as accepted by my examiners.

I understand that my thesis may be made electronically available to the public.

## Abstract

Understanding the interaction between the atmosphere and ocean is crucial to modeling or predicting the dynamics of both systems. Observational studies and numerical simulations of atmospheric responses to ocean mesoscale fronts have found strong correlation between the sea surface temperature (SST) and near surface wind speeds, which is interpreted as the SST driving the atmosphere at these scales. Furthermore, correlations studies have shown that SST fronts are capable of modifying surface turbulent heat and momentum fluxes which can affect winds, clouds and rainfall, as well as have larger scale effects on storm tracks. These larger scale effects have been observed to feed back onto the ocean and further evolve the SST fronts, thus making the need to understand the fully coupled atmosphere-ocean system more significant.

This thesis uses idealized large eddy simulations to investigate the evolution of a coupled atmosphere-ocean model initialized with a filament-like warm SST anomaly in the presence of weak  $3 \text{ m s}^{-1}$  winds blowing across the front, as well as no winds, over a period of 24 hours. Uncoupled systems are considered, holding the ocean constant in time for the atmosphere and vice versa, following an analysis of the simulations coupled together to compare. Large scale structures of the potential temperature and zonal velocity are considered alongside the mean and turbulent kinetic energy. To assist in the understanding of dominant factors of both systems a momentum budget and turbulent kinetic energy budget will be analyzed term by term. The coupling is found to have a small impact on the atmosphere over the short time scale, mainly affecting the turbulent kinetic energy strength and distribution, while having a significant impact on the ocean, causing large scale flow changes as well as altering the turbulent kinetic energy distribution.

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## **Dedication**

To my parents, Sandra and Joe, thank you for always supporting me no matter the choices I make.

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# Chapter 1

## Introduction

Atmosphere-ocean coupling has been a large and growing area of research in recent decades. From large scale observational studies to extensive numerical simulations, a better understanding of the interplay between the ocean and atmosphere has long been pursued to obtain more accurate models of our world (Small et al., 2008; Chelton and Xie, 2010; Seo et al., 2023). In this thesis we will investigate air-sea interactions and the significance of coupling through idealized simulations of an ocean mesoscale warm filament-like sea surface temperature (SST) distribution.

High resolution satellite observational studies as well as numerical experiments have revealed a positive correlation between near surface wind speeds and sea surface temperature at ocean mesoscales, which we define as scales less than 1000 km and greater than 10 km, and often a negative correlation at larger scales (Small et al., 2008; Seo et al., 2023). These phenomena are interpreted as evidence of the atmosphere driving the ocean when the correlation is negative, and vice versa when the correlation is positive (Small et al., 2008).

Studies have shown that in the ocean mesoscale regime SST fronts are capable of modifying surface turbulent heat and momentum fluxes which trigger local atmospheric responses in the boundary layer that affect winds, clouds, and rainfall (Frenger et al., 2013; Takahashi et al., 2020; Seo et al., 2023). These local changes then have larger scale effects on storm tracks as well as deep moisture convection (Czaja et al., 2019; Seo et al., 2023). The atmosphere then feeds back onto the ocean surface and SST furthering coupled atmospheric processes (Seo et al., 2023). On top of this it has been found that persistent wind stress responses to SST have the potential to directly affect surface ocean dynamics. In combination with SST, ocean mesoscale surface currents have also been found to affect

the surface heat and momentum fluxes furthering the dynamics of the coupled problem (Seo et al., 2023).

These mesoscale SST fronts of interest include filamentary eddy structures observed in coastal upwelling, mid latitude western boundary currents, and the equatorial tongue in the Atlantic and Pacific Oceans (Small et al., 2008; Seo et al., 2023). Each of these regions contains specific signatures of influence from the SST fronts associated with them. The primary influence for this thesis is the Gulf Stream, a western boundary current off the coast of the eastern United States and Canada. The Gulf Stream is a poleward warm current in the western Atlantic Ocean known for having persistent and sharp SST fronts, making it a popular region of study for understanding mesoscale air-sea interactions (Small et al., 2008). Additionally the Gulf Stream has been found to have filament-like warm structures throughout winter months which further influence this work (Seim and Edwards, 2019). Observational results in the area have shown winds blowing across and along the Gulf Stream to have strong responses that affect the wind speeds, boundary layer depth, and generation of clouds (Small et al., 2008). Some observations have also revealed jets forming within a few hundred meters of the ocean surface known as low level jets (Wayland and Raman, 1989), and were found to be strongest either with winds blowing along SST fronts or when warm air is advected over a cold surface (Vihma et al., 1998).

Alongside the many observational studies on SST fronts, many numerical simulations have been done to further our understanding of the smaller scale features, from boundary layer models to fully three dimensional regional atmospheric models forced by global models or reanalysis (Small et al., 2008). Despite this not many studies discuss idealized simulations of SST fronts in order to isolate and focus on the raw mechanics of the systems. Spall (2007) uses a 2D idealized coupled atmosphere-ocean simulation to better understand the momentum budget in the presence of various strength winds across various mesoscale temperature gradients over shorter time scales on the order of days. The coupling uses the Naval Research Laboratory Coupled Ocean–Atmosphere Mesoscale Prediction System model (Hodur, 1997) for the atmosphere and Regional Ocean Modeling System (Shchepetkin and McWilliams, 2005) where the atmospheric model receives the SST from the ocean, and the ocean model receives surface wind stress, latent and sensible heat flux, as well as net longwave and shortwave radiation from the atmosphere. Sullivan et al. (2020) also considers idealized winds blowing across uncoupled smaller (closer to submesoscale) fronts to analyze the turbulent fluxes and boundary layer at smaller scales. Along with this, Sullivan et al. (2021) also looks at winds blowing along SST fronts, and Sullivan et al. (2025) analyzes small scale and short time scale feedback of the atmosphere onto the ocean in a fully coupled system. The coupler in this case is custom built to allow the exchange of SST, currents, and momentum, temperature and humidity fluxes. These

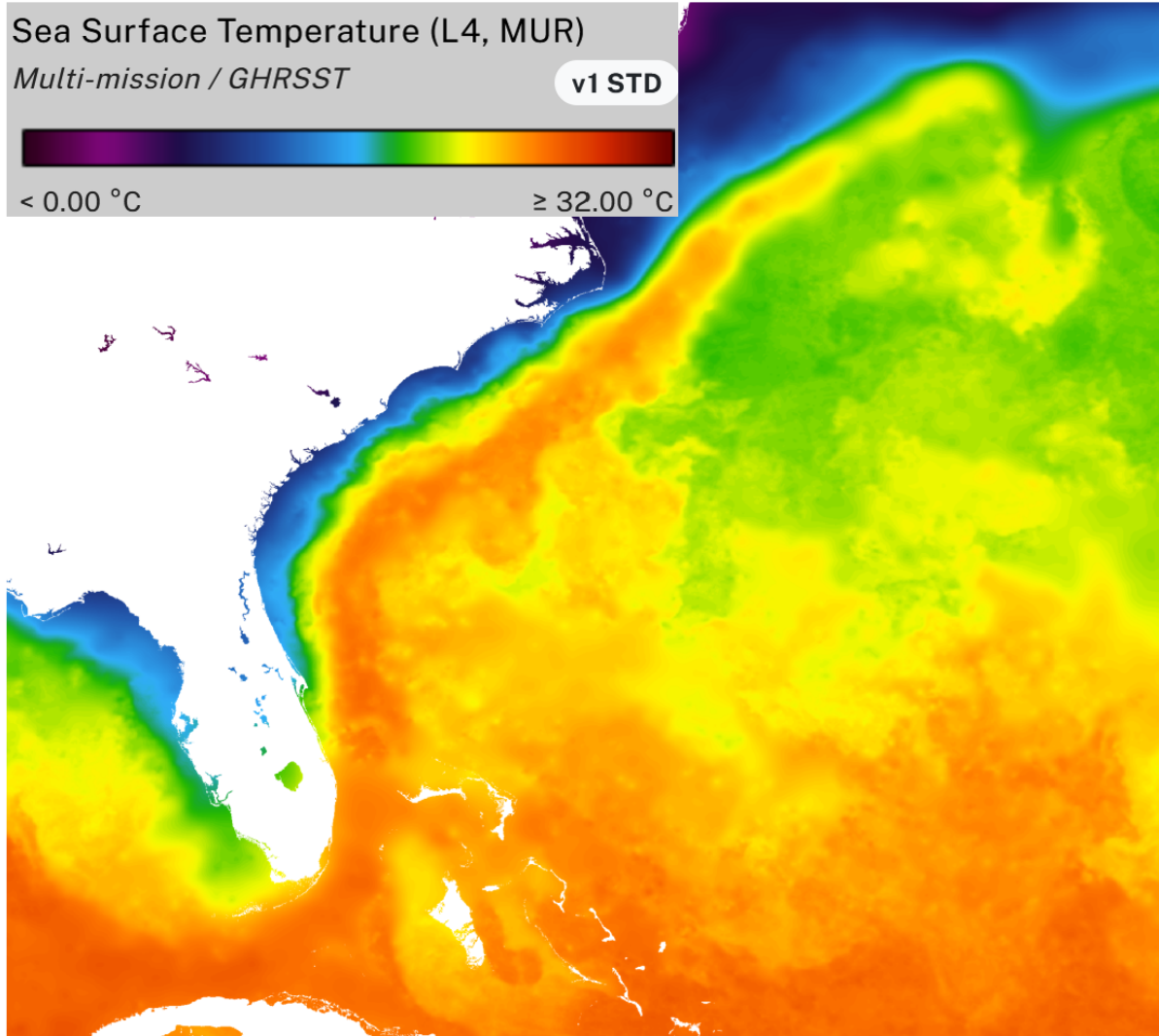


Figure 1.1: Sea surface temperature of the Gulf Stream on Jan 25, 2024. Image taken from NASA Worldview Snapshots (<https://wvs.earthdata.nasa.gov/>).

studies in particular make up a large influence for this work.

Despite the high interest in coupled air-sea models, very little literature investigates the specific effects of idealized dynamic coupling. This work aims to isolate these effects by investigating an atmospheric and oceanic system in the vicinity of a filament-like warm

mesoscale front. The systems will be analyzed first in atmospheric simulations where the ocean front is prescribed as constant in time, following oceanic simulations where the winds are prescribed as constant in time, then finally comparing these results to a fully coupled simulation. Large scale temperature and velocity structures will be analyzed alongside a momentum budget to better understand the dominate factors in each system. Alongside this, turbulent structures and major contributions to the turbulent energy will be looked at to get a better idea of the small scale phenomena in each simulation as well. Finally a comparison of each set of data will be compared with previous findings and literature.

# Chapter 2

## Background

### 2.1 The Fundamental Equations

We begin with the assumption that the reader is familiar with the governing equations of geophysical fluid dynamics, but reiterate them for use and reference later. The basis of this thesis starts with assumptions that allow us to use the  $f$ -plane and Boussinesq approximation (Vallis, 2017; Cushman-Roisin and Beckers, 2011), to be more explicit we assume the following:

- The flow is at mid-latitude (away from the equator).
- The flow is restricted to a limited meridional extent such that the Coriolis affect will remain constant throughout the domain.
- The density field can be decomposed into  $\rho(x, y, z, t) = \rho_0 + \tilde{\rho}(x, y, z, t)$  where  $\rho_0$  is a constant background density, and  $\tilde{\rho}(x, y, z, t)$  is a deviation such that  $|\tilde{\rho}| \ll \rho_0$ .

Under these assumptions, the momentum equation can be written as

$$\frac{\partial u_i}{\partial t} + u_j \frac{\partial u_i}{\partial x_j} + \epsilon_{ijk} f_j u_k = -\frac{1}{\rho_0} \frac{\partial p}{\partial x_i} + b \delta_{i3} + \frac{\partial \tau_{ij}}{\partial x_j}, \quad (2.1)$$

where  $u_i$  are the velocity components  $(u, v, w)$  aligned with Cartesian coordinates,  $f_i = (0, 0, f)$  is the Coriolis term with  $f$  being the Coriolis parameter,  $\epsilon_{ijk}$  is the Levi-Civita

symbol,  $\rho_0$  is a constant reference density,  $p$  is the pressure, and  $b$  is the buoyancy which is defined as  $-g\tilde{\rho}/\rho_0$ . The deviatoric part of the stress tensor  $\tau_{ij}$  is defined as

$$\tau_{ij} = 2\nu S_{ij}, \quad (2.2)$$

where  $S_{ij}$  is the strain rate tensor given by,

$$S_{ij} = \frac{1}{2} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right). \quad (2.3)$$

These assumptions also allow the continuity equation to be written as

$$\frac{\partial u_i}{\partial x_i} = 0. \quad (2.4)$$

Finally we use an equation of state to define the buoyancy

$$b = b(T, S), \quad (2.5)$$

where  $T$  is the temperature and  $S$  is the salinity (only in the ocean). In the atmosphere the potential temperature  $\theta$ , or the temperature of a parcel of air if it was adiabatically brought down to the surface, would be used as opposed to the temperature. The evolution for these quantities is given by ([Cushman-Roisin and Beckers, 2011](#))

$$\frac{\partial T}{\partial t} + u_i \frac{\partial T}{\partial x_i} = \kappa_T \frac{\partial^2 T}{\partial x_i \partial x_i}, \quad (2.6)$$

$$\frac{\partial S}{\partial t} + u_i \frac{\partial S}{\partial x_i} = \kappa_S \frac{\partial^2 S}{\partial x_i \partial x_i}, \quad (2.7)$$

where  $\kappa_T$  and  $\kappa_S$  are the diffusivity of temperature and salinity respectively.

## 2.2 Turbulence

### 2.2.1 Reynolds Averaging

To understand the effects of a coupled atmosphere-ocean system it will be important to understand both the mean and turbulent components of the flow; we start by finding the evolution of the mean portion of the flow. We define the decomposition of a field into a mean and fluctuation as,

$$X = \langle X \rangle + X', \quad (2.8)$$

where  $X$  is the full field,  $\langle X \rangle$  is an average taken of  $X$  (typically an ensemble, spatial, or temporal average), and  $X'$  is a fluctuation. This is known as a Reynolds Decomposition (Pope, 2001).

Note that under this definition we can average everything to obtain

$$\langle X' \rangle = 0,$$

as  $\langle \langle X \rangle \rangle = \langle X \rangle$ . With this in mind we can decompose our velocity fields, apply this to the continuity equation (2.4), and then average to obtain

$$\frac{\partial \langle u_i \rangle}{\partial x_i} = 0.$$

Given that the average and the derivative commute, the continuity equation (2.4) yields,

$$\frac{\partial u'_i}{\partial x_i} = 0.$$

We now wish to do the same for the momentum equation, however, due to the nonlinear advection we must take some care. First we write the advection term as

$$u_j \frac{\partial u_i}{\partial x_j} = \frac{\partial}{\partial x_j} (u_i u_j),$$

which holds following equation (2.4). We decompose this into

$$u_j \frac{\partial u_i}{\partial x_j} = \frac{\partial}{\partial x_j} (\langle u_i \rangle \langle u_j \rangle + \langle u_i \rangle u'_j + u'_i \langle u_j \rangle + u'_i u'_j),$$

and after averaging obtain

$$\left\langle u_j \frac{\partial u_i}{\partial x_j} \right\rangle = \frac{\partial}{\partial x_j} (\langle u_i \rangle \langle u_j \rangle + \langle u'_i u'_j \rangle).$$

Decomposing and averaging the momentum equation (2.1) while utilizing the above results allows us to have an evolution equation for the average of the flow

$$\frac{\partial \langle u_i \rangle}{\partial t} + \langle u_j \rangle \frac{\partial \langle u_i \rangle}{\partial x_j} + \epsilon_{ijk} f_j \langle u_k \rangle = -\frac{1}{\rho_0} \frac{\partial \langle p \rangle}{\partial x_i} + \langle b \rangle \delta_{i3} + \frac{\partial}{\partial x_j} (2\nu \langle S_{ij} \rangle - \langle u'_i u'_j \rangle). \quad (2.9)$$

This is known as the Reynolds averaged Navier Stokes. Note the term that was not present in the previous momentum equation, this term is commonly referred to as the Reynolds stress. Now that we have an evolution of the average, we can subtract it from the momentum equation to get an evolution equation for the fluctuation

$$\frac{\partial u'_i}{\partial t} + u'_j \frac{\partial u'_i}{\partial x_j} + \frac{\partial}{\partial x_j} (\langle u_i \rangle u'_j + u'_i \langle u_j \rangle) + \epsilon_{ijk} f_j u'_k = -\frac{1}{\rho_0} \frac{\partial p'}{\partial x_i} + b' \delta_{i3} + \frac{\partial}{\partial x_j} \left( 2\nu S'_{ij} + \langle u'_i u'_j \rangle \right). \quad (2.10)$$

### 2.2.2 Mean and Turbulent Kinetic Energy

We now look to define the kinetic energy associated with the mean and fluctuating components of the flow. We define the mean and turbulent kinetic energy (TKE) respectively as

$$\overline{K} = \frac{1}{2} \langle u_i \rangle \langle u_i \rangle, \quad (2.11)$$

$$k = \frac{1}{2} \langle u'_i u'_i \rangle. \quad (2.12)$$

We derive the evolution of the mean kinetic energy  $\overline{K}$  by multiplying the mean velocity equation (2.9) by  $\langle u_i \rangle$ , we go through this term by term to help with clarity and simplification. First the time derivative

$$\langle u_i \rangle \frac{\partial \langle u_i \rangle}{\partial t} = \frac{1}{2} \frac{\partial \langle u_i \rangle \langle u_i \rangle}{\partial t} = \frac{\partial \overline{K}}{\partial t}.$$

Similarly for the advection we have

$$\langle u_i \rangle \langle u_j \rangle \frac{\partial \langle u_i \rangle}{\partial x_j} = \langle u_j \rangle \frac{\partial \overline{K}}{\partial x_j}.$$

Note the Coriolis term is orthogonal to the velocity and does no work

$$\langle u_i \rangle \epsilon_{ijk} f_j \langle u_k \rangle = 0.$$

The pressure term will be written as

$$-\langle u_i \rangle \frac{1}{\rho_0} \frac{\partial \langle p \rangle}{\partial x_i} = -\frac{1}{\rho_0} \frac{\partial \langle p \rangle \langle u_i \rangle}{\partial x_i}.$$

The buoyancy term becomes

$$\langle u_i \rangle \langle b \rangle \delta_{i3} = \langle w \rangle \langle b \rangle.$$

The viscous stress term will be written as

$$\begin{aligned} 2\nu \langle u_i \rangle \frac{\partial \langle S_{ij} \rangle}{\partial x_j} &= 2\nu \frac{\partial \langle u_i \rangle \langle S_{ij} \rangle}{\partial x_j} - 2\nu \langle S_{ij} \rangle \frac{\partial \langle u_i \rangle}{\partial x_j} \\ &= 2\nu \frac{\partial \langle u_i \rangle \langle S_{ij} \rangle}{\partial x_j} - 2\nu \langle S_{ij} \rangle \langle S_{ij} \rangle. \end{aligned}$$

Lastly the Reynolds stress term will be written as

$$\langle u_i \rangle \frac{\partial \langle u'_i u'_j \rangle}{\partial x_j} = \frac{\partial}{\partial x_j} (\langle u_i \rangle \langle u'_i u'_j \rangle) - \langle u'_i u'_j \rangle \frac{\partial \langle u_i \rangle}{\partial x_j}.$$

Putting this all together gives

$$\frac{\partial \bar{K}}{\partial t} + \langle u_j \rangle \frac{\partial \bar{K}}{\partial x_j} + \frac{\partial \bar{\mathcal{T}}_i}{\partial x_i} = -\mathcal{P} + \langle w \rangle \langle b \rangle - \bar{\varepsilon}, \quad (2.13)$$

where the ‘transport’  $\bar{\mathcal{T}}_i$  is defined as

$$\bar{\mathcal{T}}_i = \langle u_j \rangle \langle u'_i u'_j \rangle + \frac{\langle p \rangle \langle u_i \rangle}{\rho_0} - 2\nu \langle u_j \rangle \langle S_{ij} \rangle, \quad (2.14)$$

the ‘production’  $\mathcal{P}$  is defined as

$$\mathcal{P} = -\langle u'_i u'_j \rangle \frac{\partial \langle u_i \rangle}{\partial x_j}, \quad (2.15)$$

and the ‘dissipation’  $\bar{\varepsilon}$  is defined as

$$\bar{\varepsilon} = 2\nu \langle S_{ij} \rangle \langle S_{ij} \rangle. \quad (2.16)$$

We now derive the TKE evolution equation by multiplying the fluctuation velocity equation (2.10) by  $u'_i$  and then averaging. Once again we go through the process term by term.

$$\left\langle u'_i \frac{\partial u'_i}{\partial t} \right\rangle = \left\langle \frac{1}{2} \frac{\partial u'_i u'_i}{\partial t} \right\rangle = \frac{\partial k}{\partial t}.$$

Now the advection, noting that

$$\frac{\partial}{\partial x_j} u'_i u'_i u'_j = 2u'_i u'_j \frac{\partial u'_i}{\partial x_j},$$

which gives

$$\left\langle u'_i u'_j \frac{\partial u'_i}{\partial x_j} \right\rangle = \frac{\partial}{\partial x_j} \frac{1}{2} \langle u'_i u'_j u'_i \rangle.$$

Now the advection of the mean velocity by the fluctuating flow and the advection of the fluctuation by the mean flow

$$\begin{aligned} \left\langle u'_i \frac{\partial}{\partial x_j} (\langle u_i \rangle u'_j + u'_i \langle u_j \rangle) \right\rangle &= \langle u'_i u'_j \rangle \frac{\partial \langle u_i \rangle}{\partial x_j} + \frac{1}{2} \langle u_j \rangle \frac{\partial \langle u'_i u'_i \rangle}{\partial x_j} \\ &= \langle u'_i u'_j \rangle \frac{\partial \langle u_i \rangle}{\partial x_j} + \langle u_j \rangle \frac{\partial k}{\partial x_j}. \end{aligned}$$

Note that the Coriolis term is orthogonal to the velocity and does no work

$$u'_i \epsilon_{ijk} f_j u'_k = 0.$$

The pressure term becomes

$$\left\langle -\frac{1}{\rho_0} u'_i \frac{\partial p'}{\partial x_i} \right\rangle = -\frac{1}{\rho_0} \left( \frac{\partial \langle u'_i p' \rangle}{\partial x_i} - p' \frac{\partial u'_i}{\partial x_i} \right) = -\frac{1}{\rho_0} \frac{\partial \langle u'_i p' \rangle}{\partial x_i}.$$

The buoyancy term becomes

$$\langle u'_i b' \delta_{i3} \rangle = \langle w' b' \rangle.$$

And finally the kinematic and Reynolds stress terms become

$$\begin{aligned} \left\langle u'_i \frac{\partial}{\partial x_j} (2\nu S'_{ij} + \langle u'_i u'_j \rangle) \right\rangle &= 2\nu \frac{\partial \langle u'_i S'_{ij} \rangle}{\partial x_j} - 2\nu \left\langle S'_{ij} \frac{\partial u'_i}{\partial x_j} \right\rangle, \\ &= 2\nu \frac{\partial \langle u'_i S'_{ij} \rangle}{\partial x_j} - 2\nu \langle S'_{ij} S'_{ij} \rangle. \end{aligned}$$

Putting all of this together we have

$$\frac{\partial k}{\partial t} + \langle u_j \rangle \frac{\partial k}{\partial x_j} + \frac{\partial \mathcal{T}_i}{\partial x_i} = \mathcal{P} + \langle w' b' \rangle - \varepsilon, \quad (2.17)$$

where we define the conservative rearrangement of turbulence as the ‘transport’

$$\mathcal{T}_i = \frac{1}{2} \langle u'_j u'_j u'_i \rangle + \frac{1}{\rho_0} \langle u'_i p' \rangle - 2\nu \langle u'_j S'_{ij} \rangle, \quad (2.18)$$

and the loss of TKE as the ‘dissipation’

$$\varepsilon = 2\nu \langle S'_{ij} S'_{ij} \rangle. \quad (2.19)$$

Note that the ‘production’ term  $\mathcal{P}$  defined in equation (2.15) shows up with opposite signs in both the turbulent (2.17) and mean (2.13) kinetic energy equations. This term represents a transfer of energy between the mean and turbulent flow.

## 2.3 Large Eddy Simulations (LES)

Kolmogorov theory describes to us the scales at which eddies dissipate and that the energy cascade moves from large scale to small scale (Pope, 2001). In order to fully resolve these scales and the turbulent dissipation associated with them the computational cost increases as the cube of the Reynolds number, making resolving high Reynolds number large scale flows infeasible (Pope, 2001). In order to allow for modeling of these flows, we must parameterize the small scale turbulent flows and dissipation, and only explicitly model the large scale motions, this is the core idea of LES and the models used within this thesis.

### 2.3.1 The Weather Research and Forecast Model

The Weather Research and Forecast model (WRF) (Skamarock et al., 2019) is a state of the art atmospheric simulation software used in both research environments as well as operational forecasting applications. The model supports simulations of atmospheric motions on scales from tens of meters to thousands of kilometers with various parameterization and closure options within these regimes. The model solves the compressible equations but acoustic motions are damped numerically. Given that our analysis will be restricted to the bottom 2 km of the atmosphere we will use the Boussinesq approximation in our discussion of the momentum and TKE budgets.

## Governing Equations

Before getting to the governing equations it is important to understand that WRF uses a terrain-following hydrostatic-pressure vertical coordinate  $\eta$  defined by

$$\eta = \frac{p_d - p_t}{p_s - p_t} = \frac{p_d - p_t}{\mu}, \quad (2.20)$$

where  $p_d$  is the hydrostatic component of the pressure of the dry air,  $p_s$  and  $p_t$  are the values of  $p_d$  at the bottom and top of the domain respectively, and  $\mu = (p_s - p_t)$  will be a short hand to make the equations simpler. Note that  $\mu$  is also the metric for this coordinate transformation

$$\frac{\partial p_d}{\partial \eta} = \mu.$$

Under this coordinate transformation we define the coordinate vector and a new transformed velocity respectively as

$$\begin{aligned} \xi_i &= (x, y, \eta), \\ \zeta_i &= \left( \mu u, \mu v, \mu \frac{d\eta}{dt} \right). \end{aligned}$$

The governing equations written out under this coordinate transformation are,

$$\frac{\partial \mu u}{\partial t} + \frac{\partial \zeta_i u}{\partial \xi_i} - f \mu v = \frac{\partial}{\partial x} \left( p \frac{\partial \phi}{\partial \eta} \right) - \frac{\partial}{\partial \eta} \left( p \frac{\partial \phi}{\partial x} \right) + \mathcal{D}_u, \quad (2.21)$$

$$\frac{\partial \mu v}{\partial t} + \frac{\partial \zeta_i v}{\partial \xi_i} + f \mu u = \frac{\partial}{\partial y} \left( p \frac{\partial \phi}{\partial \eta} \right) - \frac{\partial}{\partial \eta} \left( p \frac{\partial \phi}{\partial y} \right) + \mathcal{D}_v, \quad (2.22)$$

$$\frac{\partial \mu w}{\partial t} + \frac{\partial \zeta_i w}{\partial \xi_i} = g \left( \frac{\partial \phi}{\partial \eta} + \mu \right) + \mathcal{D}_w, \quad (2.23)$$

$$\frac{\partial \mu \theta}{\partial t} + \frac{\partial \zeta_i \theta}{\partial \xi_i} = \mathcal{D}_\theta, \quad (2.24)$$

$$\frac{\partial \mu}{\partial t} + \frac{\partial \zeta_i}{\partial \xi_i} = 0, \quad (2.25)$$

$$\frac{\partial \phi}{\partial t} + \frac{1}{\mu} \left( \zeta_i \frac{\partial \phi}{\partial \xi_i} - g \mu w \right) = 0, \quad (2.26)$$

where  $\phi = gz$  is the geopotential, and  $\mathcal{D}_a$  is the diffusive term for  $a$ . The diffusive terms are calculated as

$$\begin{aligned}\mathcal{D}_u &= \frac{\partial \mu \nu_H D_{11}}{\partial x} + \frac{\partial \mu \nu_H D_{12}}{\partial y} - z_x \frac{\partial \mu \nu_H D_{11}}{\partial z} - z_y \frac{\partial \mu \nu_H D_{12}}{\partial z} + \frac{\partial \mu \nu_V D_{13}}{\partial z}, \\ \mathcal{D}_v &= \frac{\partial \mu \nu_H D_{12}}{\partial x} + \frac{\partial \mu \nu_H D_{22}}{\partial y} - z_x \frac{\partial \mu \nu_H D_{12}}{\partial z} - z_y \frac{\partial \mu \nu_H D_{22}}{\partial z} + \frac{\partial \mu \nu_V D_{23}}{\partial z}, \\ \mathcal{D}_w &= \frac{\partial \mu \nu_V D_{13}}{\partial x} + \frac{\partial \mu \nu_V D_{23}}{\partial y} - z_x \frac{\partial \mu \nu_V D_{13}}{\partial z} - z_y \frac{\partial \mu \nu_V D_{23}}{\partial z} + \frac{\partial \mu \nu_V D_{33}}{\partial z}, \\ \mathcal{D}_\theta &= \left[ \left( \frac{\partial}{\partial x} - \frac{\partial}{\partial z} z_x \right) \left( \mu \nu_H \left( \frac{\partial}{\partial x} - z_x \frac{\partial}{\partial z} \right) \right) \right. \\ &\quad \left. + \left( \frac{\partial}{\partial y} - \frac{\partial}{\partial z} z_y \right) \left( \mu \nu_H \left( \frac{\partial}{\partial y} - z_y \frac{\partial}{\partial z} \right) \right) + \mu \nu_V \frac{\partial}{\partial z} \right] \theta,\end{aligned}$$

where  $\nu_H$  and  $\nu_V$  are the horizontal and vertical eddy viscosities respectively, the deformation tensor  $D_{ij}$  is defined as follows

$$\begin{aligned}D_{11} &= 2 \left( \frac{\partial u}{\partial x} - z_x \frac{\partial u}{\partial z} \right), \\ D_{22} &= 2 \left( \frac{\partial v}{\partial y} - z_y \frac{\partial v}{\partial z} \right), \\ D_{33} &= 2 \frac{\partial w}{\partial z}, \\ D_{12} &= \frac{\partial u}{\partial y} - z_y \frac{\partial u}{\partial z} + \frac{\partial v}{\partial x} - z_x \frac{\partial v}{\partial z}, \\ D_{13} &= \frac{\partial w}{\partial x} - z_x \frac{\partial w}{\partial z} + \frac{\partial u}{\partial z}, \\ D_{23} &= \frac{\partial w}{\partial y} - z_y \frac{\partial w}{\partial z} + \frac{\partial v}{\partial z},\end{aligned}$$

where finally

$$\begin{aligned}z_x &= \frac{1}{g} \frac{\partial \phi}{\partial x}, \\ z_y &= \frac{1}{g} \frac{\partial \phi}{\partial y}.\end{aligned}$$

Alongside this we also have the diagnostic relation for the density

$$W \frac{\partial \phi}{\partial \eta} = -\frac{\mu}{\rho}, \quad (2.27)$$

and the equation of state for the pressure

$$p = p_0 \left( \frac{R_d \theta \rho}{p_0} \right)^\gamma, \quad (2.28)$$

where  $\gamma = 1.4$  is the ratio of the heat capacities of dry air,  $R_d$  is the gas constant for dry air, and  $p_0$  is a reference pressure. These are the governing equations for dry air with the inclusion of rotation, WRF is also capable of solving the equations in the presence of moisture, however this is not utilized for this thesis.

## Numerics and Closure

WRF utilizes a time split integration scheme that allows for slow frequencies to be integrated using a third-order Runge-Kutta (RK3) time stepping scheme, while high frequency acoustics are integrated over shorter time steps within the RK3 time step to maintain model stability. WRF has multiple options for spatial discretizations, including the ability to use different ordered schemes for the horizontal and vertical. In this thesis fifth-order schemes are used in the horizontal and third-order for the vertical. WRF also includes a variety of turbulence closure models. For this thesis the 1.5 Order TKE closure will be utilized. On top of this WRF provides surface parameterization options to calculate estimates of near surface quantities that are not resolved. For more information on WRF and any of the schemes listed above more information can be found in [Skamarock et al. \(2019\)](#).

### 2.3.2 Oceananigans.jl

Oceananigans ([Ramadhan et al., 2020](#)) is a free and open source software package for the Julia programming language that allows for LES of nonhydrostatic and hydrostatic solutions to the Boussinesq equations. Given this thesis utilizes the nonhydrostatic model we give an overview of the equations solved and methods used.

## Governing Equations

The nonhydrostatic model of Oceananigans solves the momentum equation (2.1) with additional features to include a background flow, a Stokes drift, and additional source terms. Given that we will not be using these additional features the momentum equation takes the

same form as equation(2.1). It is worth noting that this model supports LES, which means  $\tau_{ij}$  need not be proportional to the molecular viscosity and is typically parameterized as

$$\frac{\partial \tau_{ij}}{\partial x_j} = 2\nu_E S_{ij},$$

where  $\nu_E$  is a viscosity calculated by a closure or specified by the user, typically referred to as an ‘eddy viscosity’.

The tracer evolution equation that is also used to evolve the buoyancy/temperature is given by

$$\frac{\partial c}{\partial t} = -u_i \frac{\partial c}{\partial x_i} + \frac{\partial q_i^c}{\partial x_i},$$

where  $c$  is the concentration of the tracer and  $q_i^c$  is the diffusive flux of the tracer. The diffusive flux is set by the closure scheme used and takes the typical form of

$$q_i^c = \kappa_E \frac{\partial c}{\partial x_i},$$

where  $\kappa_E$  is a diffusivity either prescribed by the user or calculated by a closure, typically referred to as an ‘eddy diffusivity’.

## Numerics and Implicit Damping

Oceananigans provides various standard high order temporal integration and spatial differencing schemes. For this thesis we will be using RK3 for the temporal integration and 5th order up wind spatial differencing. Oceananigans provides standard turbulence closure schemes including constant isotropic and anisotropic dissipation, Smagorinsky-Lilly (Smagorinsky, 1963; Lilly, 1962), anisotropic minimum dissipation (Vreugdenhil and Taylor, 2018), and more. However, the non-constant schemes do not provide an option to separate out the horizontal and vertical eddy viscosity. On the surface of the ocean we have an aspect ratio of  $\Delta x/\Delta z \approx 1200$ , where  $\Delta a$  represents the grid spacing in the  $a$  direction, which means not separating out the eddy viscosities would result in a gross overestimate or underestimate of the appropriate eddy viscosity. In response to this, using no closure scheme and allowing numerical damping to dissipate energy was chosen as the preferred option to ensure that viscous effects remained as small as possible.

To give an illustration of the numerical damping that takes place when using odd ordered spatial derivatives we use a simple first order forward difference scheme and Taylor

expand about a specific point  $x_0$

$$\frac{\partial u}{\partial x}(x_0) \approx \frac{u(x_0 + \Delta x) - u(x_0)}{\Delta x} = \frac{\partial u}{\partial x}(x_0) + \frac{\Delta x}{2} \frac{\partial^2 u}{\partial x^2}(x_0) + \mathcal{O}(\Delta x^2).$$

The leading order term here acts as diffusion proportional to  $\Delta x$ , and this principle extends to any odd ordered difference scheme with an even ordered leading error term. Although we will be using a 5th order scheme as opposed to the one mentioned above, the same principle applies but with 6th order hyperdiffusion.

## Flux Boundary Conditions

Within Oceananigans we will be imposing what they refer to as ‘flux boundary conditions’ at the surface of the ocean. These conditions allow us to specify an exact flux  $\mathcal{Q}$  associated with the surface conditions in the atmosphere and ensure continuous stress and heat flux at the air-sea interface. To illustrate how these boundary conditions work, we consider the following tracer evolution equation

$$\frac{\partial c}{\partial t} = -u_i \frac{\partial c}{\partial x_i} + \frac{\partial q_i^c}{\partial x_i}.$$

Averaging this over a finite volume  $\Omega$  and utilizing Gauss’ divergence theorem we have

$$\frac{\partial c}{\partial t} \Big|_{\Omega} = \frac{1}{V_{\Omega}} \oint_{\partial\Omega} (cu_i + q_i^c) n_i dS,$$

where  $V_{\Omega}$  is the volume of the element,  $\partial\Omega$  is the boundary of  $\Omega$ , and  $n_i$  is the normal vector to the surface. Here we have an advective and diffusive flux across each face of the element, however we want to impose our boundary conditions along faces at the boundary  $\partial\Omega_b$ . To do this we set these fluxes to be zero and then we can add the prescribed flux in after. To achieve this we note that across a fixed boundary  $\partial\Omega_b$  we have no normal flow

$$n_i u_i \Big|_{\partial\Omega_b} = 0,$$

with this we use the fact that,

$$q_i^c \propto \frac{\partial c}{\partial x_i}$$

and set

$$n_i \frac{\partial c}{\partial x_i} \Big|_{\partial\Omega_b} = 0.$$

Numerically this is done by putting an even extension of  $c$  in the halo points. This allows for

$$n_i(cu_i + q_i^c) \Big|_{\partial\Omega_b} = 0.$$

This imposes that the fluxes calculated across external boundaries are 0, and allows us to now impose our own specific flux. We now set the tracer evolution of the finite volume of a boundary adjacent cell to be

$$\frac{\partial c}{\partial t} \Big|_{\Omega} = \frac{1}{V_{\Omega}} \left( \oint_{\partial\Omega \setminus \partial\Omega_b} (cu_i + q_i^c) n_i dS + A_{\partial\Omega_b} \mathcal{Q} \Big|_{\partial\Omega_b} \text{sign}(\mathbf{n}) \right),$$

where  $A_{\partial\Omega_b}$  is the area of the external boundary,  $\mathcal{Q}$  is the prescribed flux, and  $\text{sign}(\mathbf{n})$  ensures consistency of positive flux being inward or outward.

This formulation is significant to understand as we will not be using a turbulence closure, and therefore  $\nu_E = 0$  and  $\kappa_E = 0$ , which means that the flux conditions imposed in this thesis will act as

$$\frac{\partial c}{\partial t} \Big|_{\Omega} = \frac{1}{V_{\Omega}} \left( \oint_{\partial\Omega \setminus \partial\Omega_b} cu_i n_i dS + A_{\partial\Omega_b} \mathcal{Q} \Big|_{\partial\Omega_b} \text{sign}(\mathbf{n}) \right).$$

For the momentum equation a similar process is done on the dissipation term  $\tau_{ij}$  as opposed to the diffusive flux  $q_i^c$ .

# Chapter 3

## Methodology

A series of 24 hour idealized coupled and uncoupled large eddy simulations of the ocean and atmosphere have been carried out to better understand the interaction between the two and to look into the effects and significance of coupling. Throughout this thesis quantities denoted with a super scripted (*a*) are associated with the atmosphere, while quantities denoted with a super scripted (*o*) are associated with the ocean. Quantities with neither superscript relevant to both domains. For simplicity, the horizontal domain is the same across both models with  $L_x = 691.2$  km in the zonal direction and  $L_y = 76.8$  km in the meridional direction with resolutions  $\Delta x = \Delta y = 600$  m and periodic lateral boundary conditions. The formulation that follows assumes that the interface occurs at  $z = 0$  m; the ocean where  $z < 0$  m and the atmosphere where  $z > 0$  m. The specifics of each models configuration are as follows. A depiction of the idealization is seen in Figure 3.1.

### 3.1 Atmospheric Model

For the atmosphere, WRF (Skamarock et al., 2019) is used to model the evolution of a dry atmosphere. The depth of the atmosphere is  $L_z^{(a)} = 20$  km, vertical grid spacing remains approximately constant from  $z = 0$  m to  $z \approx 2048$  m with  $\Delta z \approx 16$  m, and then increases linearly for another 32 grid levels until reaching  $z = L_z^{(a)} = 20$  km with a top level grid spacing of  $\Delta z \approx 2500$  m. The time step is set to  $\Delta t^{(a)} = 5$  s to ensure the CFL condition is not exceeded. A sponge layer with implicit Rayleigh damping as seen in Klemp et al. (2008), is used in the top 15 km of the domain to absorb any reflected gravity waves. A surface layer parametrization scheme given by Jiménez et al. (2012) is used to

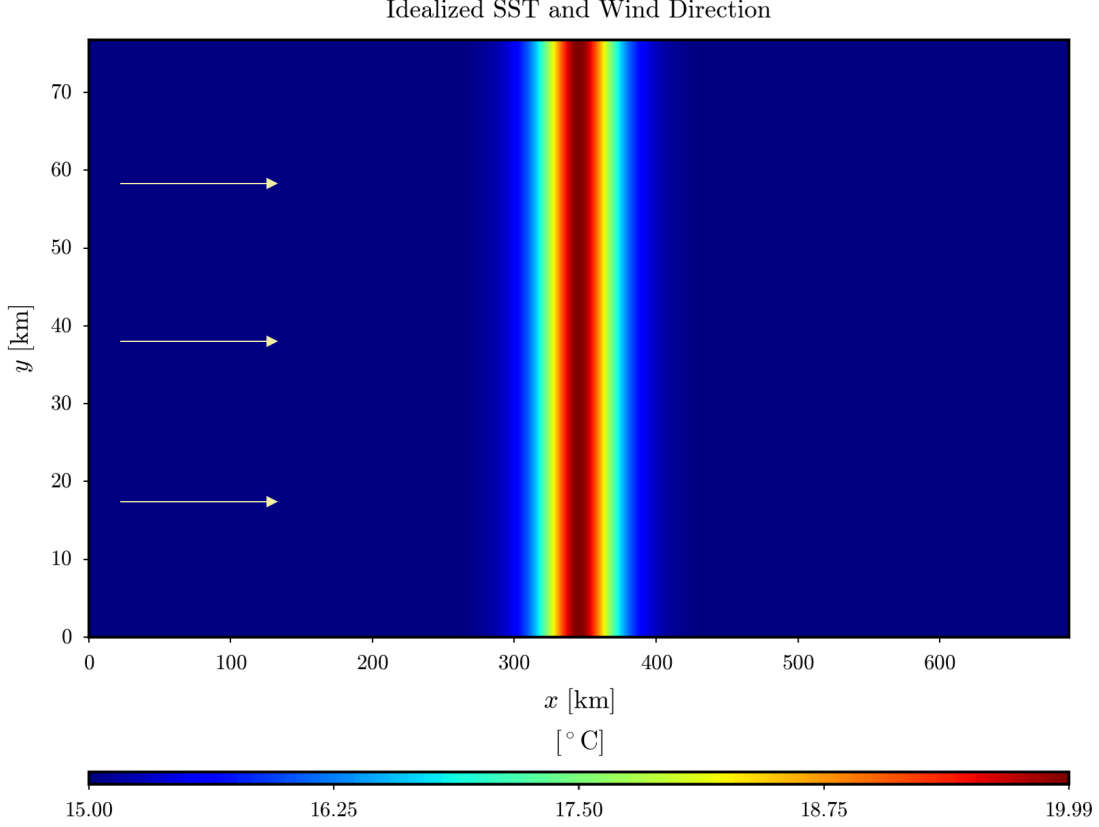


Figure 3.1: Sea surface temperature distribution of the idealized filament looked at in this thesis, the arrows depict the direction the wind will be blowing in.

calculate near surface values and statistics including the friction velocity  $u_*^{(a)}$ , the 10 m wind speeds  $u_{10}^{(a)}, v_{10}^{(a)}$ , the surface heat flux  $Q_h^{(a)}$ , and the 2 m potential temperature  $\theta_2^{(a)}$ . The surface parameterization scheme assumes that the surface of the ocean is stationary which introduces a limitation to our coupled model in that surface ocean currents will not be coupled. Lastly, the 1.5 order TKE closure scheme is used to calculate a separate horizontal and vertical eddy viscosity (Skamarock et al., 2019).

Following the work of Spall (2007), the atmosphere is initialized with a stratified po-

tential temperature field given by

$$\theta^{(a)}\Big|_{t=0} = \theta_0^{(a)} \left( 1 - \frac{\Gamma^{(a)}}{\theta_0^{(a)}} z \right)^{(\Gamma_d^{(a)} - \Gamma^{(a)})}, \quad (3.1)$$

where  $\theta^{(a)}(x, y, z, 0)$  is the initial potential temperature of the atmosphere and  $\theta_0^{(a)} = 14$  °C is the initial temperature of the air at  $z = 0^+$ , just above the ocean surface. The constants  $\Gamma_d^{(a)} = 9.8 \times 10^{-3}$  °C m<sup>-1</sup> is the adiabatic lapse rate of dry air, and  $\Gamma^{(a)} = 6.7 \times 10^{-3}$  °C m<sup>-1</sup> is a prescribed lapse rate such that the potential temperature is initially stratified. Small perturbations of about 1% the magnitude of potential temperature are added to the bottom few grid levels to ensure the development of three dimensional turbulence.

Two scenarios are considered: cases with no initial background winds and others with weak uniform zonal winds,  $u^{(a)}(x, y, z, 0) = 3$  m s<sup>-1</sup> throughout the depth of the domain. Finally, convection is forced by heat flux generated by a warm SST with a filament like distribution in the zonal direction. In the non-coupled case the SST is constant in time and described by the initial state of the ocean temperature  $T^{(o)}(x, y, 0, 0)$ , as described in equation (3.2).

## 3.2 Oceanic Model

For the ocean we use the Oceananigans model as described in [Ramadhan et al. \(2020\)](#) to model the evolution of a filament-like temperature distribution. The depth of the ocean is set to  $L_z^{(o)} = 1024$  m with a half Chebychev grid ([Boyd, 2001](#)) in the vertical where  $\Delta z \approx 0.5$  m at the surface and  $\Delta z \approx 50$  m at the bottom, the time step is set to  $\Delta t^{(o)} = 30$  s. No turbulence closure is employed and numerical damping is responsible for the diffusion. As discussed in Section 2.3.2 the aspect ratio of the surface of the ocean is  $\Delta x / \Delta z \approx 1200$  and a separation of the eddy viscosities would be best to not over or underestimate the viscous effects; however Oceananigans currently does not provide a closure that does this beyond setting a constant vertical and horizontal eddy viscosity. In order to ensure the viscous effects remain as small as possible in both the horizontal and vertical, no closure scheme was chosen as the system is stable enough with just implicit damping.

We will investigate the coupling interaction of a filament like temperature distribution with background linear stratification as described by

$$T^{(o)}\Big|_{t=0} = T_0^{(o)} + \frac{(N^{(o)})^2}{g\alpha^{(o)}} z + \Delta T^{(o)} \text{sech}^2 \left( \frac{c_j}{\ell_x} (x - \ell_c) \right) \text{sech}^2 \left( \frac{c_j}{\ell_z} z \right), \quad (3.2)$$

where  $T^{(o)}(x, y, z, 0)$  is the initial temperature profile of the ocean,  $T_0^{(o)} = 15$  °C is the initial SST far from the filament and is chosen such that it is 1 degree warmer than the atmosphere above to ensure weak convection occurs throughout the entire domain.  $(N^{(o)})^2 = 1.4 \times 10^{-5} \text{ s}^{-2}$  is the constant background buoyancy frequency,  $g = 9.8 \text{ m s}^{-2}$  is the acceleration due to gravity,  $\alpha^{(o)} = 2 \times 10^{-4} \text{ }^\circ\text{C}^{-1}$  is the thermal expansion coefficient of the ocean,  $\Delta T^{(o)} = 5$  °C is the maximum temperature perturbation of the filament,  $c_j = 6$  is a scaling constant chosen such that  $\ell_x = 153.6 \text{ km}$  and  $\ell_z = 1024 \text{ m}$  are the dimensions of the region that is above 10% of  $\Delta T^{(o)}$  i.e. the length scale of most of the filament, and  $\ell_c = L_x/2$  is the point at which the filament is centered.

For the filament to be initially stable we must ensure that it is in geostrophic and hydrostatic balance

$$\begin{aligned} f v^{(o)} &= \frac{\partial \Phi^{(o)}}{\partial x}, \\ f u^{(o)} &= -\frac{\partial \Phi^{(o)}}{\partial y}, \\ b^{(o)} &= \frac{\partial \Phi^{(o)}}{\partial z}, \end{aligned}$$

where  $f = 10^{-4} \text{ s}^{-1}$  is the Coriolis parameter,  $b^{(o)}$  is the buoyancy, and  $\Phi^{(o)}$  is the normalized pressure. Using a linear equation of state, while assuming salinity to be uniform,

$$b^{(o)} \Big|_{t=0} = g \alpha^{(o)} \left( T^{(o)} \Big|_{t=0} - T_0^{(o)} \right) + (N^{(o)})^2 z,$$

where  $b_0^{(o)}$  is the linear constant background stratification of the ocean, we arrive at the following initial geostrophic velocity profile

$$v^{(o)} \Big|_{t=0} = -\frac{2g\alpha^{(o)}\Delta T^{(o)}\ell_z}{f\ell_x} \text{sech}^2 \left( \frac{c_j}{\ell_x} (x - \ell_c) \right) \tanh \left( \frac{c_j}{\ell_x} (x - \ell_c) \right) \left( \tanh \left( \frac{c_j}{\ell_z} z \right) + 1 \right).$$

The jet parameters are also checked to ensure that there are no inertial and gravitational instabilities present in the flow. Finally, small perturbations of about 1% the magnitude of  $v^{(o)}(t = 0)$  are added to the horizontal velocities to more easily enable the onset of three-dimensional turbulence.

Similarly to the atmosphere, cases are considered with no initial atmospheric winds as well as weak zonal winds,  $u^{(a)}(x, y, z, 0) = 3 \text{ m s}^{-1}$ . For the uncoupled ocean cases flux boundaries must be supplied at the top of the domain. For the wind stress a constant in

time, uniform in space quadratic drag (Kara et al., 2007) is supplied and the temperature flux is chosen such that the background stratification is smooth at the boundary,

$$\mathcal{Q}_u^{(o)} = -\frac{\rho_0^{(a)}}{\rho_0^{(o)}} C_D U_{10}^{(a)} |U_{10}^{(a)}|, \quad (3.3)$$

$$\mathcal{Q}_v^{(o)} = -\frac{\rho_0^{(a)}}{\rho_0^{(o)}} C_D V_{10}^{(a)} |V_{10}^{(a)}|, \quad (3.4)$$

$$\mathcal{Q}_T^{(o)} = \frac{\left(N^{(o)}\right)^2}{g\alpha^{(o)}}, \quad (3.5)$$

where  $\mathcal{Q}_u^{(o)}$ ,  $\mathcal{Q}_v^{(o)}$ ,  $\mathcal{Q}_T^{(o)}$  are the prescribed  $u$ ,  $v$ , and  $T$  fluxes respectively as described in Section 2.3.2,  $\rho_0^{(a)} = 1.225 \text{ kg m}^{-3}$  is the average density of the atmosphere at the surface,  $\rho_0^{(o)} = 1026 \text{ kg m}^{-3}$  is the average density of the ocean at the surface,  $C_D = 1.6 \times 10^{-3}$  (Kara et al., 2007) is the dimensionless drag coefficient, and  $U_{10}^{(a)}, V_{10}^{(a)}$  are constant in time uniform in space approximations to the 10 meter wind speeds of the atmosphere. In the case of no winds  $U_{10}^{(a)} = V_{10}^{(a)} = 0 \text{ m s}^{-1}$  while in the case of weak zonal winds  $U_{10}^{(a)} = 2.7 \text{ m s}^{-1}$  which was obtained by taking the first time step of the uncoupled WRF simulation and using the average value of  $u_{10}^{(a)}$  from the surface layer parametrization scheme. Note that we use the velocity of the winds alone, as opposed to taking the difference between the surface currents and winds. This is to remain consistent with how the surface stress will be calculated in the coupled case, and given the very small magnitude of the zonal surface currents of the uncoupled simulations, this does not make much of a difference.

### 3.3 Coupling

To couple the above mentioned models together a custom script is written, the outline of which is as follows. First WRF is initialized and the initial conditions for Oceananigans are set. Then, WRF reads the SST from Oceananigans and begins running for the duration of  $\Delta t^{(o)}$ . Once completed, Oceananigans reads in the last output of  $u_*^{(a)}$ ,  $u_{10}^{(a)}$ ,  $v_{10}^{(a)}$ , and  $Q_h^{(a)}$

| Data Set | Coupling  | Domain     | $U^{(a)}$ [m s <sup>-1</sup> ] |
|----------|-----------|------------|--------------------------------|
| UA0      | Uncoupled | Atmosphere | 0                              |
| CA0      | Coupled   | Atmosphere | 0                              |
| UA3      | Uncoupled | Atmosphere | 3                              |
| CA3      | Coupled   | Atmosphere | 3                              |
| UO0      | Uncoupled | Ocean      | 0                              |
| CO0      | Coupled   | Ocean      | 0                              |
| UO3      | Uncoupled | Ocean      | 3                              |
| CO3      | Coupled   | Ocean      | 3                              |

Table 3.1: Experiments and Datasets, the first letter indicates if the set is coupled or uncoupled, second is for the ocean or atmosphere, and third is the initial zonal wind speed prescribed.

from WRF, the wind stress and heat flux are then calculated as

$$\mathcal{Q}_u^{(o)} = -\frac{\rho_0^{(a)}}{\rho_0^{(o)}} \left(u_*^{(a)}\right)^2 \frac{u_{10}^{(a)}}{\sqrt{\left(u_{10}^{(a)}\right)^2 + \left(v_{10}^{(a)}\right)^2}}, \quad (3.6)$$

$$\mathcal{Q}_v^{(o)} = -\frac{\rho_0^{(a)}}{\rho_0^{(o)}} \left(u_*^{(a)}\right)^2 \frac{v_{10}^{(a)}}{\sqrt{\left(u_{10}^{(a)}\right)^2 + \left(v_{10}^{(a)}\right)^2}}, \quad (3.7)$$

$$\mathcal{Q}_T^{(o)} = \frac{Q_h^{(a)}}{\rho_0^{(o)} c_p^{(o)}}, \quad (3.8)$$

where  $c_p^{(o)} = 3991 \text{ J } ^\circ\text{C}^{-1} \text{ kg}^{-1}$  is the specific heat capacity of water. This ensures the fluxes are continuous across the boundary. Subsequently, Oceananigans takes a single time step, and the process is repeated.

An important note to make is that the WRF surface parameterization always assumes that the ocean surface is stationary, so the atmosphere can not feel or respond to the velocity on the ocean surface, and the stress on the ocean will be calculated as such. This approximation is supported based on the surface velocities in the zonal direction differing by an order of magnitude.

## 3.4 Analysis

### 3.4.1 Averaging

The idealized simulation set up is such that there is no variation in the initial conditions in the meridional  $y$  direction apart from random noise added to the potential temperature and velocity fields to ensure 3D turbulence. Our structures remain fairly homogeneous in the  $y$  direction throughout the domain so we will focus on looking at an  $x$ - $z$  cross section. Given that the meridional direction is homogeneous we look to average in this direction for our statistics. To be more explicit, for the results of this thesis we define the Reynolds decomposition as seen in Section 2.2.1 as follows,

$$X' = X - \langle X \rangle_y. \quad (3.9)$$

where  $\langle X \rangle_{ab..}$  is an average of  $X$  in  $a, b, \dots$ . This then allows us to define the covariance as

$$\langle X'Y' \rangle_y = \frac{1}{N-1} \sum_{i=0}^N \left( X \Big|_{y=y_i} - \langle X \rangle_y \right) \left( Y \Big|_{y=y_i} - \langle Y \rangle_y \right), \quad (3.10)$$

and lastly to define the mean and turbulent kinetic energy respectively as

$$\overline{K} = \frac{1}{2} \langle u_i \rangle_y \langle u_i \rangle_y, \quad (3.11)$$

and

$$k = \frac{1}{2} \langle u'_i u'_i \rangle_y. \quad (3.12)$$

### 3.4.2 Budgets

To understand which factors dominate the flow toward the end of the simulation after the atmosphere has reached a more statistically steady equilibrium, we look to a momentum and turbulent kinetic energy budget. Within the final hour of simulated time the outputs are generated every 30 seconds of simulated time to allow for better resolved time averaging and derivatives. In this time frame components of both the momentum equation for the zonal velocity under the Boussinesq approximation as described in equation (2.1), as well as the evolution of the turbulent kinetic energy equation as described in equation (2.17) will be averaged in time and the meridional direction.

## Momentum Budget

To better understand the contributions to the zonal velocity of the uncoupled and coupled cases we consider the zonal component of the momentum equation (2.1) which we break down in the following way,

$$\underbrace{\frac{\partial u}{\partial t}}_{\text{Change}} = - \underbrace{u_i \frac{\partial u}{\partial x_i}}_{\text{Advection}} + \underbrace{fv}_{\text{Coriolis}} - \underbrace{\frac{1}{\rho_0} \frac{\partial p}{\partial x}}_{\text{Pressure}} + \underbrace{\frac{\partial \tau_{1i}}{\partial x_i}}_{\text{Dissipation}}. \quad (3.13)$$

Given that we are using LES we must define  $\tau_{ij}$  not based on molecular viscosity but on how the closure in the simulations calculates them with their respective eddy viscosities. In the atmosphere, WRF uses a 1.5 Order TKE closure (Skamarock et al., 2019)

$$\begin{aligned} \tau_{1i}^{(a)} &= -2\nu_H^{(a)} S_{1i}^{(a)}, \quad i = 1, 2, \\ \tau_{13}^{(a)} &= -2\nu_V^{(a)} S_{13}^{(a)}, \end{aligned}$$

where  $\nu_H^{(a)}$  and  $\nu_V^{(a)}$  are the horizontal and vertical eddy viscosities as calculated by the closure.

For the ocean we do not use an explicit turbulence model but instead let numerical dissipation from the advection scheme stabilize the simulations. In this case we simply use the residual of the known momentum equation to calculate the dissipation

$$\frac{\partial \tau_{1i}^{(o)}}{\partial x_i} = \frac{\partial u^{(o)}}{\partial t} + u_i^{(o)} \frac{\partial u^{(o)}}{\partial x_i} - fv^{(o)} + \frac{1}{\rho_0^{(o)}} \frac{\partial p^{(o)}}{\partial x}.$$

## TKE Budget

We also look to analyze the contributions to the evolution of the turbulent kinetic energy seen in equation (2.17). Reiterated with our specific averaging notation the equation is

$$\underbrace{\frac{\partial k}{\partial t}}_{\text{Change}} = - \underbrace{\langle u_i \rangle_y \frac{\partial k}{\partial x_i}}_{\text{Advection}} - \underbrace{\frac{\partial \mathcal{T}_i}{\partial x_i}}_{\text{Transport}} + \underbrace{\mathcal{P}}_{\text{Production}} + \underbrace{\langle w'b' \rangle_y}_{\text{Buoyant Production}} - \underbrace{\varepsilon}_{\text{Dissipation}}, \quad (3.14)$$

where  $\mathcal{T}_i$  is the transport,  $\mathcal{P}$ , is the production, and  $\varepsilon$  us the dissipation. Similar to the momentum budget, we must define the transport  $\mathcal{T}_i$ , and the dissipation  $\varepsilon$  differently for the atmosphere and ocean.

For the atmosphere the transport  $\mathcal{T}_i$  is defined by

$$\mathcal{T}_i^{(a)} = \frac{1}{2} \left\langle u_i^{(a)'} u_j^{(a)'} u_j^{(a)'} \right\rangle_y + \frac{1}{\rho_0^{(a)}} \left\langle u_i^{(a)'} p^{(a)'} \right\rangle_y - \Xi_i^{(a)},$$

where  $\Xi_i$  is the viscous transport. Due to the non-isotropic eddy viscosities calculated in the atmosphere it is defined as

$$\Xi_i^{(a)} = 2 \left( \nu_H^{(a)} (1 - \delta_{i3}) + \nu_V^{(a)} \delta_{i3} \right) \left( \left\langle u^{(a)'} S_{i1}^{(a)'} \right\rangle_y + \left\langle v^{(a)'} S_{i2}^{(a)'} \right\rangle_y \right) + 2\nu_V^{(a)} \left\langle w^{(a)'} S_{i3}^{(a)'} \right\rangle_y.$$

The transport in the ocean is defined as

$$\mathcal{T}_i^{(o)} = \frac{1}{2} \left\langle u_i^{(o)'} u_j^{(o)'} u_j^{(o)'} \right\rangle_y + \frac{1}{\rho_0^{(o)}} \left\langle u_i^{(o)'} p^{(o)'} \right\rangle_y,$$

note that there is no turbulent transport term in the ocean; similar to the momentum budget we will be calculating a residual and grouping this term to the dissipation. The production is defined the same for the atmosphere and the ocean, just as it is defined in the derivation of equation (2.17). Lastly the dissipation is defined for the atmosphere as

$$\begin{aligned} \varepsilon^{(a)} = & 2\nu_H^{(a)} \left( \left\langle S_{11}^{(a)'} S_{11}^{(a)'} \right\rangle_y + 2 \left\langle S_{12}^{(a)'} S_{12}^{(a)'} \right\rangle_y + \left\langle S_{22}^{(a)'} S_{22}^{(a)'} \right\rangle_y \right) \\ & + 2\nu_V^{(a)} \left( 2 \left\langle S_{13}^{(a)'} S_{13}^{(a)'} \right\rangle_y + 2 \left\langle S_{23}^{(a)'} S_{23}^{(a)'} \right\rangle_y + \left\langle S_{33}^{(a)'} S_{33}^{(a)'} \right\rangle_y \right), \end{aligned}$$

while the dissipation for the ocean is defined as the residual, i.e.

$$\varepsilon^{(o)} = -\frac{\partial k^{(o)}}{\partial t} - \left\langle u_i^{(o)} \right\rangle_y \frac{\partial k^{(o)}}{\partial x_i} - \frac{\partial \mathcal{T}_i^{(o)}}{\partial x_i} + \mathcal{P}^{(o)} + \left\langle w^{(o)'} b^{(o)'} \right\rangle_y.$$

# Chapter 4

## Results

To better understand the simulated results we first consider the system with no initial winds and then weak winds. First looking at the system with no winds allows us to reduce the problem to a simpler state and isolate out certain phenomena. We will first analyze the potential temperature acting as the primary driver of the atmosphere as discussed in the literature, followed by the effects of this temperature evolution on the zonal velocity. Beyond this we do term by term analysis of the momentum budget to better understand the mechanics of the overall flow and then look to the mean and turbulent kinetic energy to better understand how energy is distributed, transported, generated, and dissipated in both systems. Once we have a basis for comparison in the no-winds case, we repeat the above procedure for the case of weak winds in order to assess the effect of winds on the coupling.

### 4.1 No Initial Winds

#### 4.1.1 Potential Temperature

We first investigate the time evolution of the potential temperature for the coupled and uncoupled cases since convection is the primary driver in this configuration. Figure 4.1 illustrate how the potential temperature of the atmosphere and ocean evolve due to their interactions. Each panel of the figure plots a snapshot of the potential temperature  $\theta$  averaged in the homogeneous  $y$  direction. Given that the oceanic response to the atmosphere occurs at the top few meters, the vertical scale is stretched out for the ocean to provide a

better picture. The left column displays the time evolution of the uncoupled data, while the right column displays the coupled data. Each row is a snapshot at the time of the simulation specified in the title.

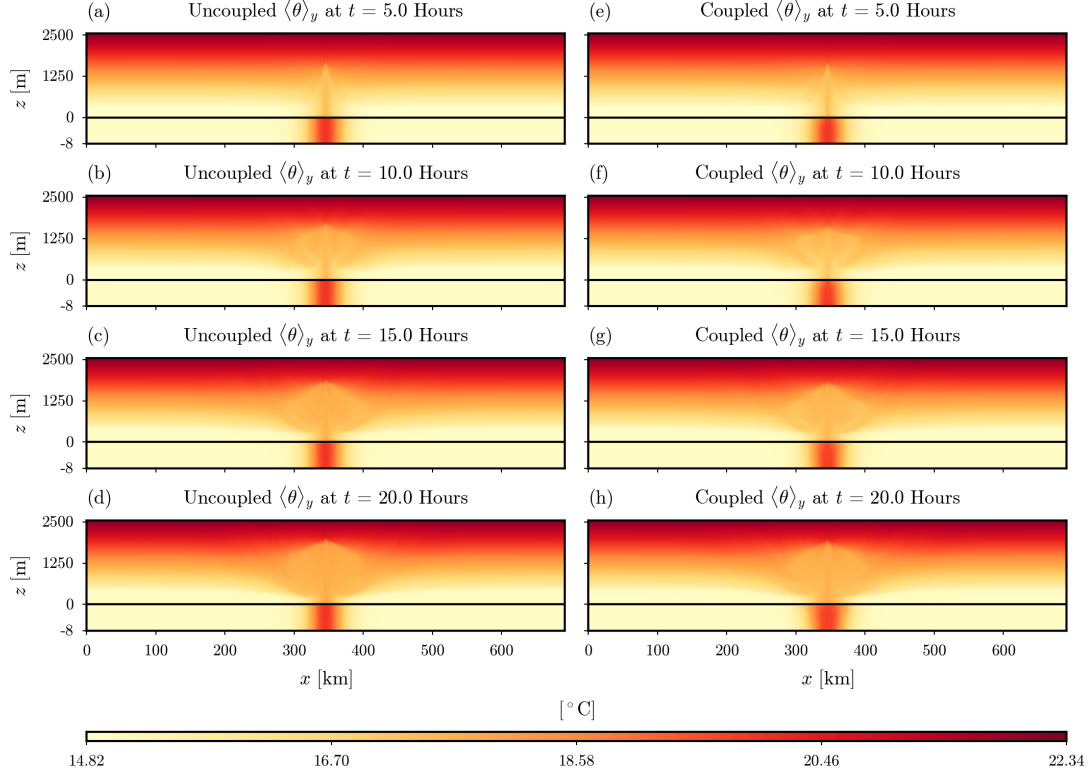


Figure 4.1: Evolution of the potential temperature  $\theta$  averaged in the homogeneous direction in both the coupled and uncoupled cases with no initial winds. The black line indicates where the datasets are separated between the ocean data (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere.

We start with the uncoupled simulations UA0 and UO0 which are seen in the left column of Figure 4.1 (a)-(d). In the atmosphere, we see a plume of warm air rising from the center of the filament until it reaches about  $z = 1$  km where the potential temperature of the stratified background is similar to that of the plume. From here the plume expands creating a bubble of warm air going through various spatial patterns until remaining fairly

constant with the pattern as seen in (d). Within the ocean, UO0, we see no obvious response or evolution. There is no wind stress, a very small surface heat flux, and as mentioned the jet was initialized to be inertially and gravitationally stable. On these short time scales we would expect to see little to no evolution since baroclinic instability grows more slowly.

The coupled simulation results seen in Figure 4.1 (e)-(h) show the coupled potential temperatures. Within the atmosphere, we see very similar patterns to that of the uncoupled case; the only notable difference being the larger variations within the warm air bubble as it evolves in time. As time goes on we see cooler regions on the interior of the bubble toward the center of the filament. Within the ocean, we begin to see evolution in time over the course of the simulation. The temperature of the ocean filament visibly cools, unlike in the uncoupled case, as well as evolves in structure, first narrowing near the surface in (f) and then expanding in (h).

### 4.1.2 Velocity and Momentum Budget

We now present the evolution of the zonal velocities to better understand the response to the temperature evolution. Figure 4.2 illustrate how the zonal velocities of the atmosphere and ocean evolve in response to each other. Similarly to Figure 4.1, each panel of both figures displays a snapshot of zonal velocity averaged in the homogeneous  $y$  direction  $\langle u \rangle_y$ . Once again the vertical scale is stretched out for the ocean to provide a better picture. Unlike the depictions of the potential temperature in Figure 4.1, to better understand the differences and similarities within the ocean and atmosphere each have their own scale and color bar. Lastly, there is a white dashed line along the  $z = 0$  m line which represents the SST, however the  $z$  scale is unrelated to the height of the dashed line, it is purely a visual aid for the reader to better see the qualitative shape and location of the SST anomaly at that time. The left column in Figure 4.2 (a)-(d) displays the time evolution of the uncoupled data, while the right column (e)-(h) display the coupled data. Each row is a snapshot at the time of the simulation specified in the title.

We start with looking at the uncoupled data which are seen in the left column of Figure 4.2 (a)-(d). Within the atmosphere, even in early response as seen in Figure 4.2 (a), we see an acceleration of the near surface wind speeds within the vicinity of the warmer SST as expected from the results of other literature (Spall, 2007; Small et al., 2008; Chelton and Xie, 2010; Seo et al., 2023). We see surface convergence toward the center of the filament with an inverse response around  $z = 1$  km forming two convective cells around the filament. This is an expected response as the warmer ocean temperature causes larger heat flux and

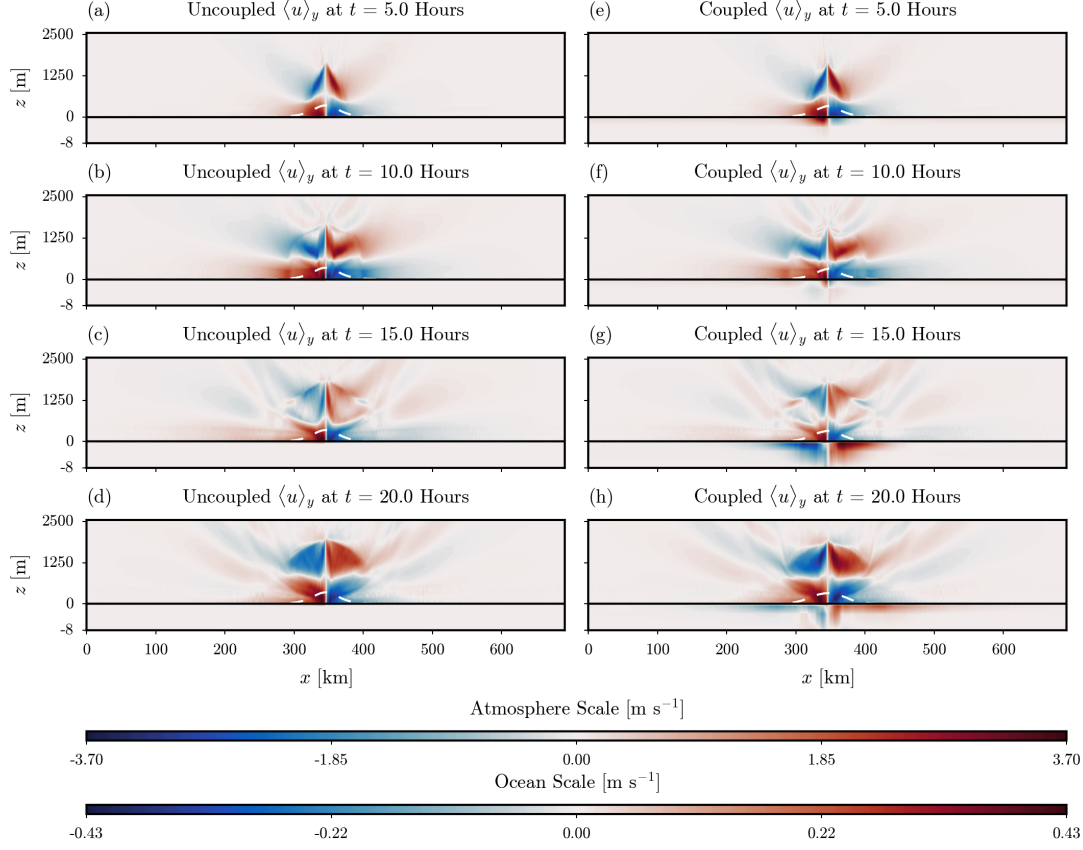


Figure 4.2: Evolution of the zonal velocity averaged in the homogeneous direction in both the coupled and uncoupled cases with no initial wind. The black line indicates where the simulations are separated between the ocean (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere. The white dashed line represents the qualitative shape and location of the SST anomaly, however is unrelated to the scale of the  $z$  axis and should be seen as a visual aid only.

a plume of warm air to rise until the potential temperature equalizes and then spreads out. This response is consistent with the potential temperature evolution in Figure 4.1 with the plume rising and spreading out into a bubble. Once again we see the cell go through an evolving spatial pattern while spreading out over time, but after achieving the state seen

in (d) the pattern remains more steady. As for the ocean, we see no obvious response or evolving zonal velocities. As with the potential temperature, there is no wind stress and a very small surface heat flux.

We now look at the coupled simulations in Figure 4.2 (e)-(h) to better understand the effects of the dynamic coupling on the zonal velocity. We see very similar response in the CA0 simulation as we do for the uncoupled UA0 simulation and there does not seem to be any obvious qualitative difference throughout the simulated time. In contrast for the ocean data CO0 we see some very notable differences from its uncoupled counterpart UO0. Early on in Figure 4.2 (e) we see an immediate response to the wind stress induced by the atmosphere with surface velocities qualitatively matching the structure of the near surface atmospheric winds. The surface convergence seen in the atmosphere due to the plume of warm air rising is reflected across the surface and the same velocity patterns are seen in the ocean at the surface, suggesting that the wind stress in this case is enough to drive the ocean surface. At later times the evolution of the ocean deviates away from this, in Figure 4.2 (f) we see the near surface velocities to be weaker than panel (e) despite having more time to be accelerated by the atmospheric winds above and a weak horizontal divergence can be seen just below the surface. In (g) we see that the flow has grown in magnitude and spread out, but reversed in direction from panel (e) becoming negatively correlated to the near surface atmospheric winds. Near the end of the simulation in panel (h) we see a structure similar to panel (g) but with weaker velocities, and a weak flow acting against the surface flow at  $z \approx -8$  m and  $x \approx 310$  km. Although not illustrated well in the snapshots, this flow pattern is indicative of oscillations occurring, and an animation reveals this to be the case. Given the only forcing here is from the wind stress which is relatively steady in time, the oscillations are likely due the fact that the wind stress from the atmosphere is strong enough to cause adjustments within the ocean jet. This is consistent with the potential temperature evolution which seems to expand and contract with the structure of the velocities.

To better understand the significant contributions that affect the dynamics toward the end of the simulation when the atmosphere has settled into a more stable state, we look to the momentum budget of the zonal velocity as described in equation (3.13). Figure 4.3 displays the time averaged normalized terms within the 24th hour of simulated time ( $23 \text{ h} < t < 24 \text{ h}$ ). Given the scale difference of the atmosphere and ocean each data set is normalized with respect to the maximum magnitude term in the budget for both the ocean and atmosphere. This allows us to not only see relative strength of each term within its own simulation but also a comparison of the strength in the uncoupled and coupled simulations. The sign convention is such that the positive regions are acting to accelerate toward the right. Once again a white dashed line is present along the  $z = 0$  axis to indicate

the location of the SST anomaly for ease of the reader. For all simulations and budgets, the sum of the budget terms were compared to that of the time derivative, and found to be very similar, suggesting the 30 s sample rate is sufficient to capture an accurate time derivative.

Looking at the uncoupled data UA0 and UO0 in the atmosphere we see the dominant balance of the zonal momentum is between advection, Coriolis, and pressure. The time derivative is small and the dissipation term is almost entirely negligible. Specifically, the exterior region of the convective cells are almost entirely balanced by the Coriolis term and the pressure gradient with advection making a smaller contribution, suggesting by this point in time these regions are in approximate geostrophic balance. Within the interior and between the cells, specifically above the region where the filament is the strongest, we see contributions toward a thin vertical structure from both advection and pressure. This aligns well with the position of where the SST would be injecting consistent energy into the atmosphere by heating the air directly above its warmest region. The heating generates buoyancy which causes vertical convergence that contributes to both advection and the pressure gradient. Within the ocean the results are as expected based off the lack of evolution within the potential temperature and the zonal velocity. Within the budget terms we see the dynamics fully dominated by the Coriolis and pressure terms, indicative of the flow remaining in geostrophic balance as it was initialized, even at the end of the simulated time.

Compared to the uncoupled cases, the coupled atmosphere shows very similar results which is once again consistent with the evolution of the potential temperature and velocity as seen previously. In the ocean, however, where we previously saw large differences within the coupled and uncoupled cases, we see that the jet breaks away from geostrophic balance at the surface near the filament, and the new dominant term in this region is the dissipation as the residual.

### 4.1.3 Mean and Turbulent Kinetic Energy

With a better understanding of the dynamics we now turn to the kinetic energy of the system. We consider the decomposed mean and turbulent kinetic energy of the system as described in equations (3.11) and (3.12). The time evolution of the mean kinetic energy of the no-winds case is seen in Figure 4.4, with a layout very similar to that of the zonal velocities as described previously. Looking at the uncoupled data in Figure 4.4 (a)-(d), for the atmosphere we see within the mean kinetic energy an evolution that aligns well with the potential temperature and zonal velocity in terms of structure. A narrow structure coming from the center of the filament spreads out over time until reaching a more

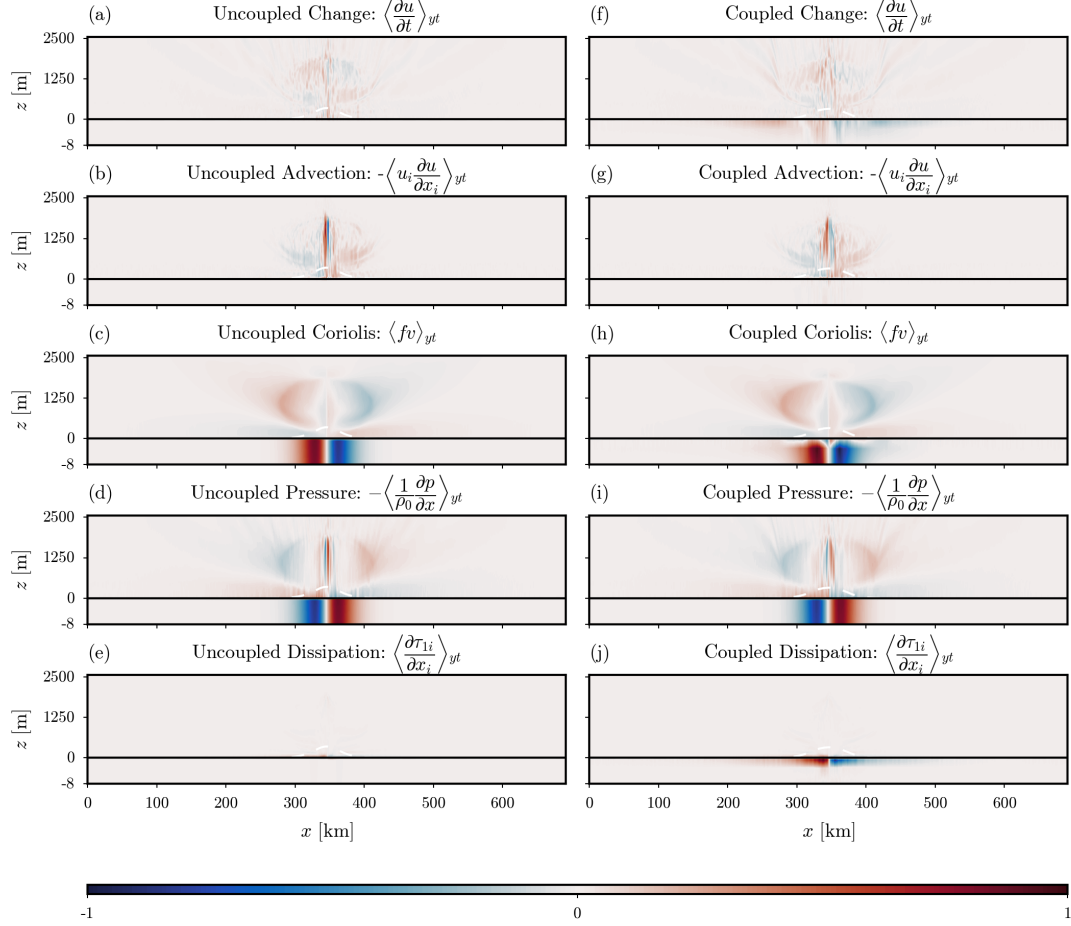


Figure 4.3: Normalized budget of the zonal momentum equation over time  $23 \text{ h} < t < 24 \text{ h}$  with no initial winds. The atmosphere and ocean values are normalized with respect to the maximum magnitude budget term in each set of uncoupled and coupled simulations such that only a single color bar is needed to understand the relative importance of each term. The black line indicates where the simulations are separated between the ocean (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere. The white dashed line represents the qualitative shape and location of the SST anomaly, however is unrelated to the scale of the  $z$ -axis and should be seen as a visual aid only.

stable configuration. However, the mean kinetic energy ends up being more concentrated toward the outside and bottom of the cells, which aligns especially well with the regions in which the atmosphere became geostrophically balanced within the momentum budget (Figure 4.3). This is unsurprising given that the zonal velocities, although dominant, are only a contribution to the mean kinetic energy alongside that of the meridional velocities generated by the Coriolis term of the momentum budget. For the uncoupled ocean, we see a constant in time structure that is expected from the structure of the initial ocean jet and the lack of any evolution seen in all previous results.

Moving to the coupled cases, Figure 4.4 (e)-(h), we see similar differences as with the velocity and potential temperature, where the coupling does not qualitatively affect the atmosphere much but largely affects the ocean. Within the ocean the mean kinetic energy gets rearranged in time, and even becomes quite strong near the end of the simulation in panel (h) further supporting the evidence that the wind stress induces an adjustment in the jet that allows for the generation of more kinetic energy. It is also of note to see that although the kinetic energy is stronger below the surface in the final time of the coupled case (panel (h)), it is significantly weaker at the surface, which aligns well with the dissipation dominating the momentum budget as seen previously.

We can now look at the resolved turbulent kinetic energy  $k$  to better understand the smaller scale dynamics of the system. Figure 4.5 displays the time evolution of the TKE with the same layout as that of the mean kinetic energy. Following the uncoupled data seen in Figure 4.5 (a)-(d) we see features of the system not hinted at previously. In the atmosphere the TKE early on is in a thick weak plume above the filament, but as the dynamics evolve, it becomes more concentrated toward the center, becoming a very fine plume centered on the filament. Near the end of the simulation in panel (d) the plume also begins to weaken, suggesting the mixing reaching a more stable configuration. This is similar to the results seen in the momentum budget (Figure 4.3) where the advection and pressure terms had a structure similar to the thin vertical plume seen here. Within the ocean there is no trace of TKE, as the ocean does not evolve from its initial stable state.

Now, comparing this to the coupled no-winds case seen in Figure 4.5 (e)-(f), within the atmosphere we actually see some qualitative differences for the first time. Throughout the simulation, the TKE of the coupled atmosphere is noticeably weaker. In panels (g) and (h) specifically, we see the the plume of TKE have a weaker core that does not stretch as high as in the uncoupled case. The TKE in the ocean also deviates from the patterns seen in previous results. Despite the adjustment that is induced in the ocean jet we actually see the TKE response in the ocean to be directly aligned with the TKE of the atmosphere, and much more concentrated than where the dissipation in the momentum budget is seen. Similar to the atmosphere the TKE in the ocean also weakens near the end of the simulation

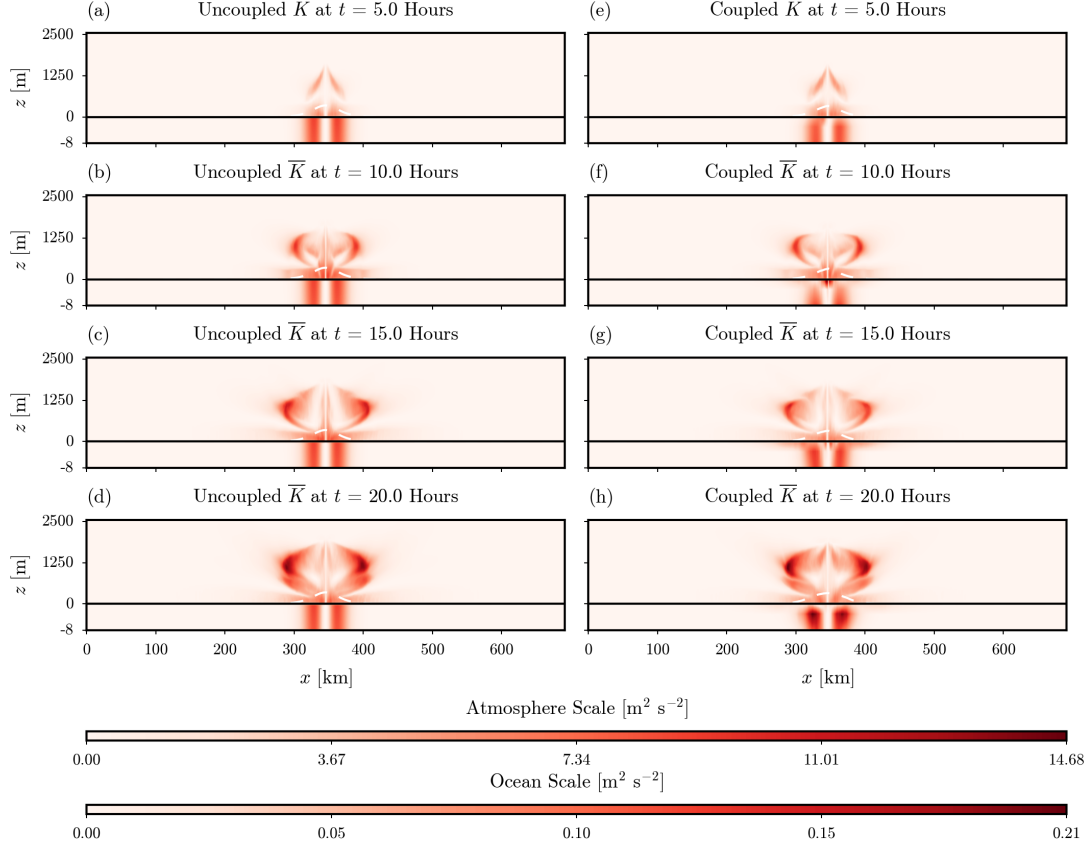


Figure 4.4: Evolution of the mean kinetic energy as described in equation (3.11) in both the coupled and uncoupled cases with no initial wind. The black line indicates where the simulations are separated between the ocean (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere. The white dashed line represents the qualitative shape and location of the SST anomaly, however is unrelated to the scale of the  $z$ -axis and should be seen as a visual aid only.

time.

To better understand the evolution of the TKE we consider the contributions to the budget described in equation (3.14) similar to the momentum budget. Figure 4.6 shows the significant contributions of each term with a layout similar to that of the momentum

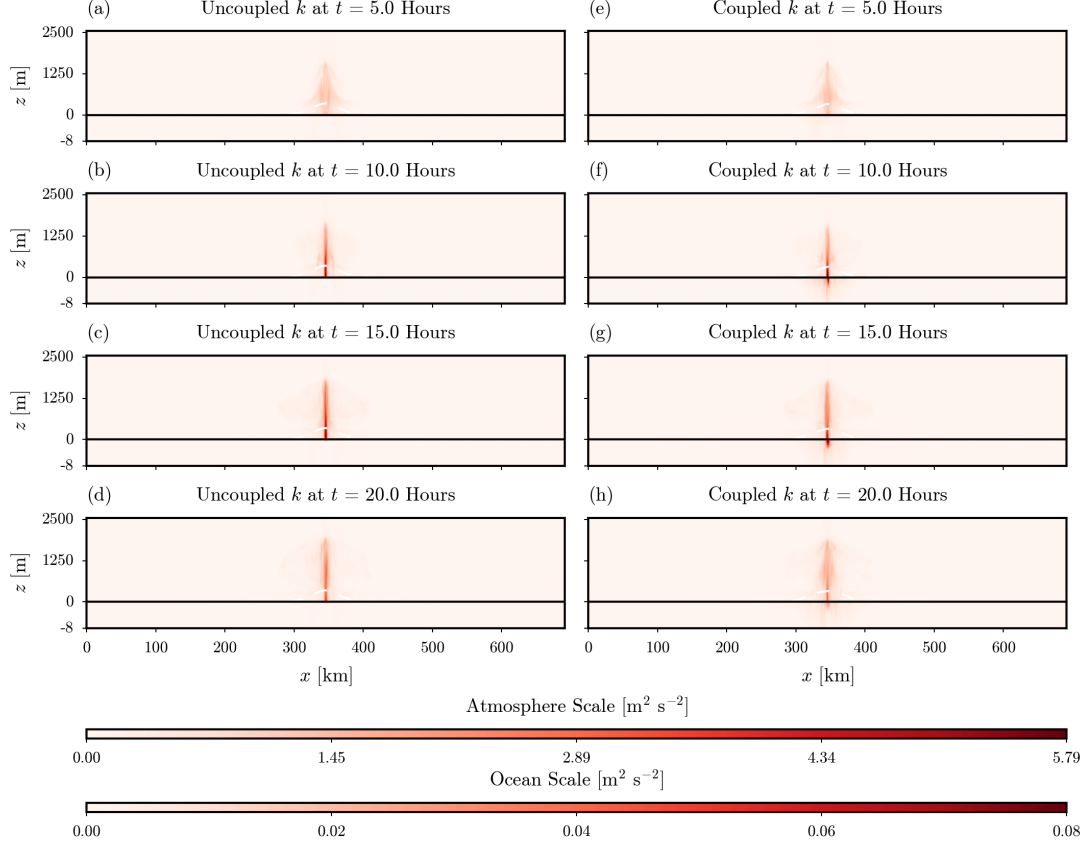


Figure 4.5: Evolution of the turbulent kinetic energy  $k$  as described in equation (3.12) in both the coupled and uncoupled cases with no initial wind. The black line indicates where the simulations are separated between the ocean (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere. The white dashed line represents the qualitative shape and location of the SST anomaly, however is unrelated to the scale of the  $z$  axis and should be seen as a visual aid only.

budget (Figure 4.3) as described previously. Within the uncoupled data given in panels (a)-(f) we focus on the atmosphere as the ocean does not evolve. We see a weak time derivative term around the region of the narrow plume seen in the TKE Figure 4.5 which suggests that the TKE is not evolving strongly at this point in the simulation. The advection

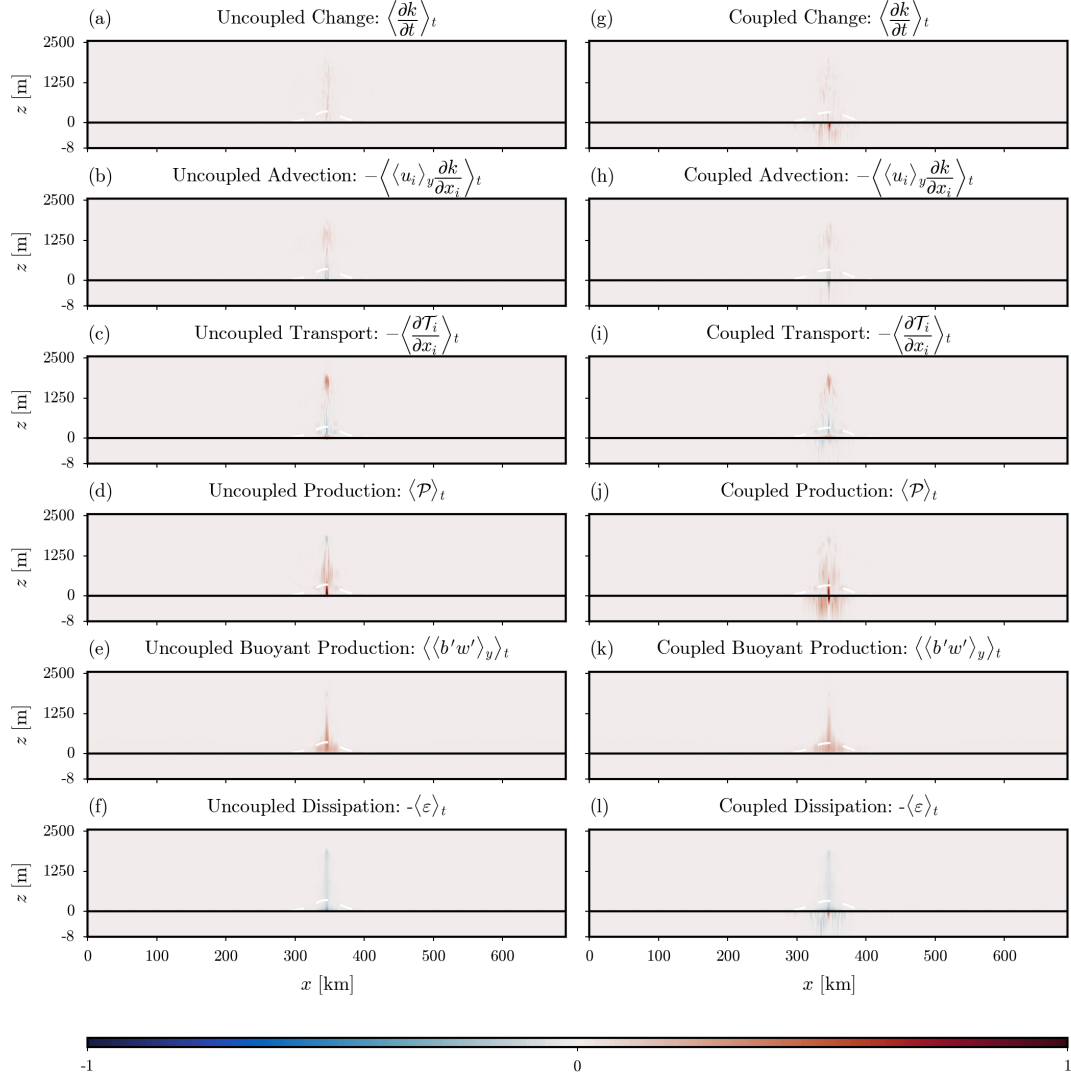


Figure 4.6: Normalized budget of the TKE equation (3.14) over time  $23 \text{ h} < t < 24 \text{ h}$ . The details of the average and layout are the same as the momentum budget in Figure 4.3.

term is also weak along the length of the plume, playing a larger role at the top. The transport production and buoyant production terms are the three largest contributors, with the transport dominating at the top of the plume where the advection is also strongest. The production contributes to a wide region around the plume and also dominates the

relationship at a small region where the core of the TKE plume is. We also see the production become negative and a transfer of TKE to the mean flow at the very top of the plume. The buoyant production covers a wider region near the surface centered on the filament. Above this region the buoyant production decays away from the edges of the filament until making a thin plume similar to that of the TKE. Lastly the dissipation is a notable contributor evenly spread out around the region of the plume. It is worth noting that although the heat flux of the filament is the primer driver of the atmosphere, the buoyant production does not dominate directly over the center of the filament where the TKE is greatest, instead the production, or the transfer of energy from the mean flow is where the TKE plume gets most of its energy.

For the coupled data in Figure 4.6 (f)-(l), similar to the TKE in Figure 4.5 we see qualitatively different results in the atmosphere and dramatically different results in the ocean. The atmospheric results show a weaker time derivative than the uncoupled case which suggest that the TKE in the coupled case actually achieves a more steady state. The advection and transport are similar to the uncoupled case but weaker overall. The production encompasses a wider region around the plume than the uncoupled simulation while the buoyant production and the dissipation are qualitatively similar. For the ocean we see a large region where the time derivative is quite strong, which aligns well with the fact that the ocean was not moving toward a steady state as seen in the evolution. The advection has a minor contribution in a thin band centered on the filament, and the transport is negligible. The production is the bulk of the contribution to the TKE in the ocean and occurs strongest along the surface, with a strong region centered just below the core of the TKE plume above. The buoyant production term is negligible which suggests the TKE dynamics are not dictated by the convection in the ocean but instead by the momentum flux from the atmosphere, and the internal adjustment of the jet. Lastly the dissipation coincides well with the region of the production but counteracting its effect.

#### 4.1.4 Discussion

The evolution of the atmosphere and ocean in the vicinity of a filament-like SST anomaly in the absence of winds allows us to understand the basic mechanics of the air-sea interaction as well as the immediate effects of dynamic coupling. An immediate surface convergence in the atmosphere centered on the ocean front was observed and followed by a plume of warm air rising until achieving equilibrium with the atmospheric stratification. From this plume two convective cells form around the center of the filament and the flow was able to quickly achieve a somewhat steady state. The momentum budget reveals the exterior edges of both cells to be in approximately geostrophic balance, with strong pressure gradients

along the interface of the two cells centered on the filament and advection playing a major role with it. An energy analysis reveals most of the mean kinetic energy is focused on the exterior of the cells, while the interface between the cells has a strong plume of TKE mostly being generated from the buoyant production caused by the warm ocean filament as well a transfer of energy from the mean flow.

Even on these short time scales, the dynamic coupling plays a large role in influencing the ocean flow. The surface winds and heat flux are enough to cause a response in the ocean strong enough to offset the balance of the jet and an adjustment follows. On top of the strong adjustment rearranging the kinetic energy and velocities of the jets interior, the momentum budget reveals that coupling also induces dissipation at the surface near the filament. This is further explored in the TKE where it is found that TKE generation occurs mostly due to production from the mean flow, and remains very close to the surface near where the atmospheric effects are strongest. The oceans lack of dramatic evolution in its temperature distribution allows for a lack of qualitative large scale differences in the atmosphere within the 24 hours of simulated time. However, an investigation of the resolved TKE of the coupled atmosphere reveals a noticeably weaker TKE signature as well as qualitatively different contributions in the budget.

## 4.2 Weak Winds

With an understanding of the coupled dynamics of the filament forcing the atmosphere and responding, we now can compare these responses with weak winds blowing across the front. All figures from the previous section will be recreated with the winds simulations UA3, UO3, CA3, and CO3.

### 4.2.1 Potential Temperature

Now investigating the cases with initial winds, the time evolution of the potential temperature can be seen in Figure 4.7. We start with the uncoupled cases seen in Figure 4.7 (a)-(d). For the atmosphere, early on in panel (a) we see results similar to that of the no-winds case, a plume of air rising from the filament until reaching a height of about  $z = 1$  km. However in the presence of winds the plume is offset from the center of the filament in the direction of the winds. In addition, there is another perturbation of the initial stratification aligned with the plume at about  $z = 2$  km. This perturbation was not seen in the no-winds case. Beyond this initial rise, the filament continues to generate warm air in the region of

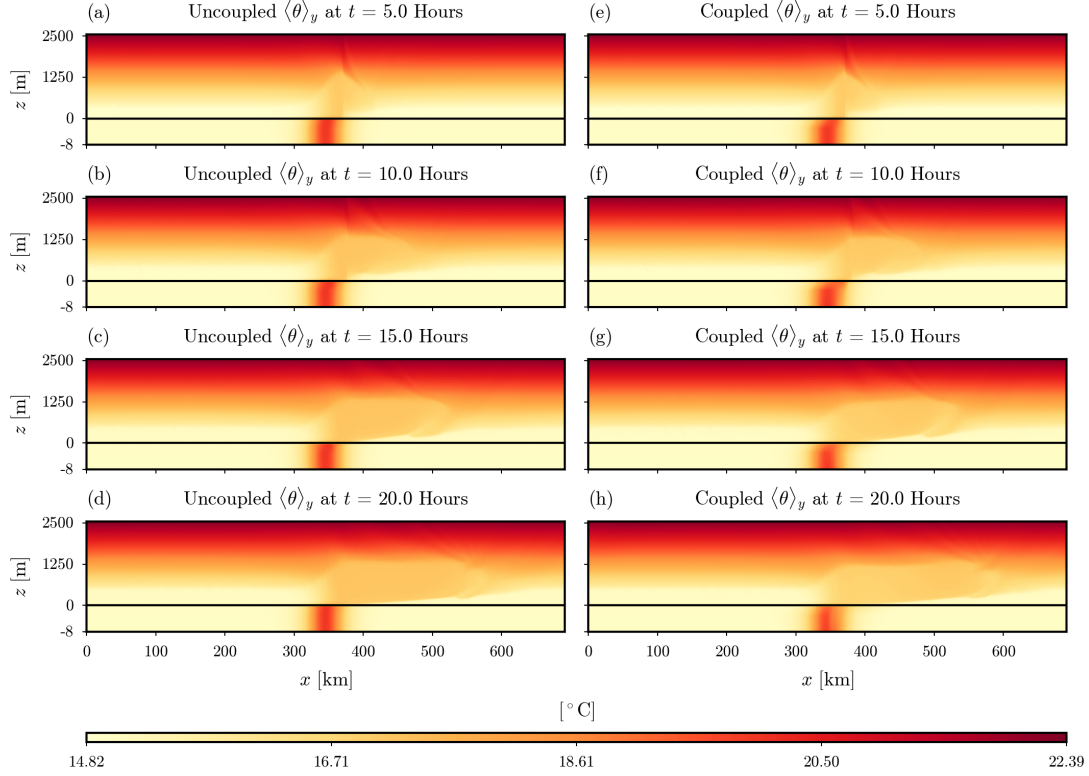


Figure 4.7: Evolution of the potential temperature  $\theta$  averaged in the homogeneous direction in both the coupled and uncoupled cases with weak  $3 \text{ m s}^{-1}$  zonal winds. The black line indicates where the datasets are separated between the ocean data (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere.

the plume, and the plume is advected and smeared out downwind. The perturbation in the upper layer is also smeared out over time and seemingly loses its significance. Within the ocean we once again see very little changes to the actual temperature however this time the constant wind stress along the surface allows for a slight deviation at the surface downwind.

Moving to the case of coupled initial winds CA3 and CO3, seen in Figure 4.7 (e)-(h), once again within the atmosphere we have very similar results to the uncoupled case. The only notable difference in the coupled atmosphere we see is the warm patch spread

farther downwind than the uncoupled case. Within the ocean we once again see a much stronger response than in the uncoupled case; the filament visibly cooling over time and also adjusting to the right near the surface, an expected result given the wind stress. Although this occurs in the uncoupled simulation, the effect is much stronger in the coupled simulation suggesting the wind stress of the coupled case to be stronger.

## 4.2.2 Velocity and Momentum Budget

Once again we now move to looking at the time evolution of the zonal velocity as well as the individual contributions of the momentum budget at the end of the simulated time. Starting with Figure 4.8, a depiction of the evolution of the zonal velocity over time. Note that in this depiction the initial  $3 \text{ m s}^{-1}$  winds were subtracted off the atmosphere.

Looking at the uncoupled data as seen in Figure 4.8 (a)-(d) for the atmosphere, early on in panel (a) we see a large disturbance in the winds around the filament with acceleration near surface winds over the first half of the filament. There is also a weaker surface winds going in the reverse direction of the mean winds over the far side of the filament similar to the surface convergence seen previously in the no-winds case. Above these reverse winds we see a patch of strongly accelerated winds once again in the direction of the mean flow. As time evolves the patch of reverse flow on the far side of the filament changes direction and becomes a patch of slightly weaker surface winds, and the strong patch of accelerated winds above gets advected downwind and stretched out while radiating waves. After the large disturbance is advected downwind in Figure 4.8 (d) we are left with a seemingly stable configuration with consistent strong surface winds in the vicinity of the filament as well as a patch of weak winds just above the filament, and a jet just beyond the filament at about 1250 m. This jet is consistent with findings of low level jets found over warm winds moving over cold ocean fronts as discussed in Small et al. (2008). Within the ocean, for the UO3 data, we see a very weak response, which contrasts the response seen in CO0 despite having similar wind speeds in certain regions. In UO3 we see that ocean response to  $3 \text{ m s}^{-1}$  winds is very relaxed, however, the near surface winds around the filament for the CO0 case were of similar magnitude but the response was much more dramatic. When comparing the quadratic wind stress approximation based off the  $2.7 \text{ m s}^{-1}$   $10 \text{ m}$  wind speed for the uncoupled ocean seen in equation (3.3) and the wind stress as calculated by WRF for the  $3 \text{ m s}^{-1}$  wind case away from the filament we see a difference of an order of magnitude suggesting the approximation may just be inappropriate at these scales. Within the UO3 data we also see very weak inertial oscillations at the surface, which are more dominant than the effects of the prescribed wind stress.

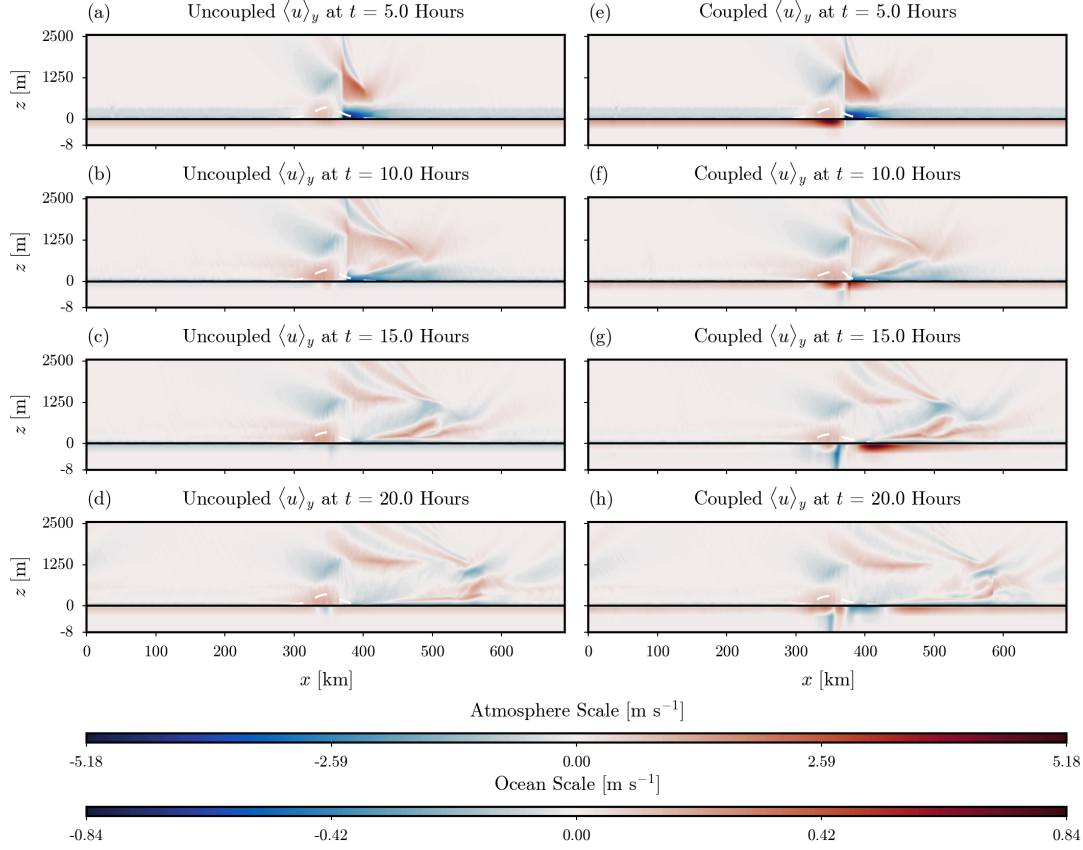


Figure 4.8: Evolution of the zonal velocity averaged in the homogeneous direction in both the coupled and uncoupled cases with weak  $3 \text{ m s}^{-1}$  initial zonal winds. Note that the initial winds were subtracted from the atmosphere. The black line indicates where the simulations are separated between the ocean (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere. The white dashed line represents the qualitative shape and location of the SST anomaly, however is unrelated to the scale of the  $z$  axis and should be seen as a visual aid only.

Moving to the coupled data seen in Figure 4.8 (e)-(h), we see results with similar patterns to that of the no-wind case. Once again in the atmosphere, the inclusion of dynamic coupling on these time scales does not make a qualitative impact; the structures

found in Figure 4.8 (a)-(d) are very similar to the results of panels (e)-(h). For the ocean the dynamic coupling has a significant impact. Similar to CO0, we see early on the surface ocean velocities directly affected by the accelerated surface winds above. Beyond this initial acceleration things once again get more unexpected with regions of surface convergence and divergence forming as time goes on, becoming less correlated to the winds above. Once again this is indicative of the ocean jet making adjustments.

To better understand the dominant contributions to these dynamics we look at the evolution of the zonal velocity equation term by term for the last simulated hour as seen in Figure 4.9. The layout is the same as previously described for 4.3.

Looking to the uncoupled data seen in Figure 4.9 (a)-(e), the atmosphere reveals very different patterns than what is seen in the no-winds case. The time derivative is somewhat strong over the filament as well as within the initial perturbation that is advected downwind, with small contributions that have wave like structures to the right of the filament. From the advection we see a strong unorganized contribution both over the filament as well as downwind following the initial perturbation and within the perturbation aligning with the time derivative. The Coriolis term weakly affects the region near the low level jet as well as the initial perturbation that is advected downwind. The Coriolis term also makes a consistent contribution along the surface of the ocean downwind of the filament. The pressure term loosely balances out the Coriolis term in the region above the filament and within the perturbation downwind, although is not as strongly in balance as seen in the no-winds case. The pressure term also appears dominant immediately above the filament which is consistent with the heat flux generating a pressure gradient. The dissipation is negligible which is similar to the no-winds case. For the ocean the time derivative and advection are negligible. The Coriolis term almost fully aligns with the jet as it did in the uncoupled case however deviates away from the surface on the left, showing more evolution than in the no-winds case. The Coriolis term also has a consistent contribution along the surface of the ocean which is further evidence of inertial oscillations in the top grid point. The pressure term remains consistent with the ocean jet, qualitatively similar to the no-winds case. The dissipation in the ocean is mostly negligible however there is a weak contribution that seems to counteract the Coriolis contribution at the surface of the ocean.

The coupled data is shown in Figure 4.9 (f)-(j). Once again the atmosphere is qualitatively very similar to that of the uncoupled data. For the ocean, the time derivative is no longer negligible, forming structures below and to the right of the filament that resemble the zonal velocity seen in Figure 4.8 (h). The advection contribution is weaker than the time derivative but covers a similar region. The Coriolis term is similar to that of the uncoupled case with an extra contribution to the right of the filament, this new Coriolis

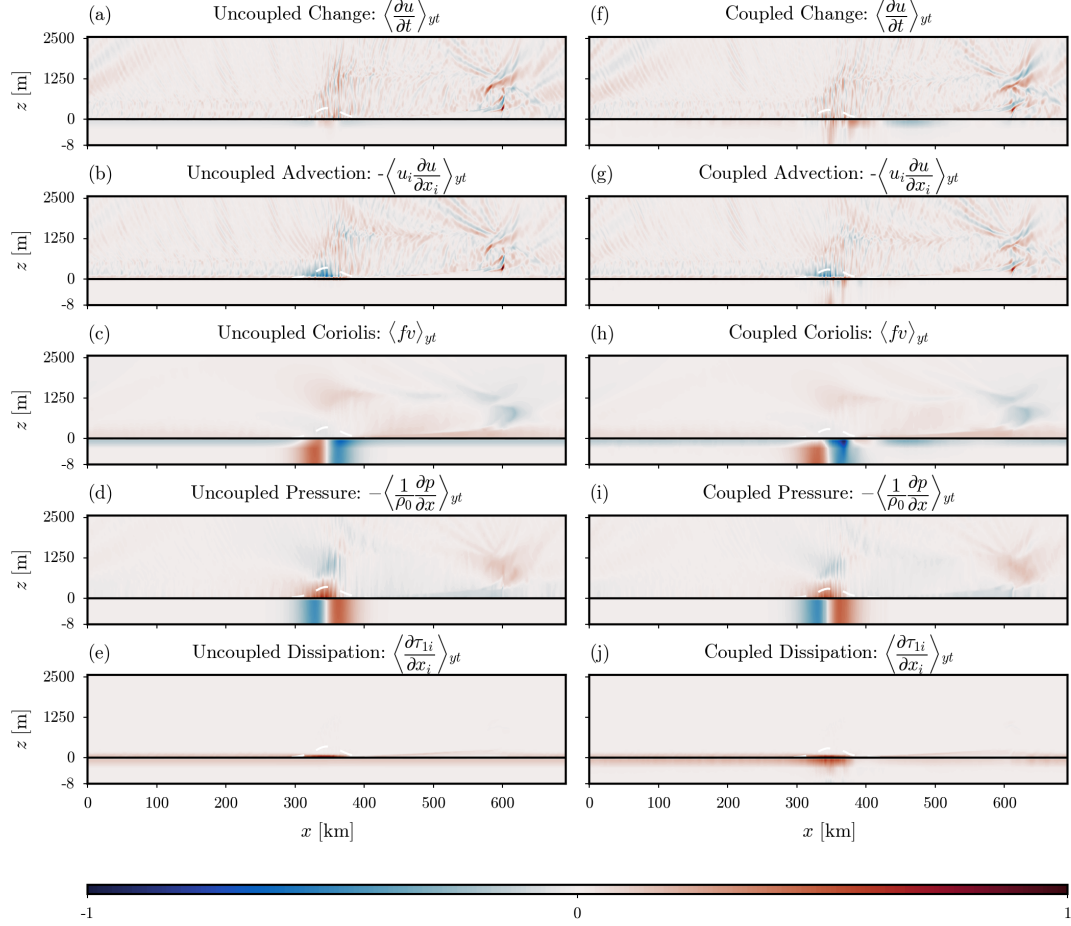


Figure 4.9: Normalized budget of the zonal momentum equation over time  $23 \text{ h} < t < 24 \text{ h}$  with  $3 \text{ m s}^{-1}$  zonal winds. The atmosphere and ocean are normalized with respect to the maximum magnitude budget term in each set of uncoupled and coupled simulations, and then scaled such that only a single color bar is needed to understand the relative importance of each term. The black line indicates where the simulations are separated between the ocean (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere. The white dashed line represents the qualitative shape and location of the SST anomaly, however is unrelated to the scale of the  $z$ -axis and should be seen as a visual aid only.

contribution aligns very well with the time derivative, suggesting the Coriolis term is the only contributor to the evolution of the flow in this region. The pressure is also similar to the uncoupled case, and once again the dissipation plays a much stronger role in the region of the filament just below the surface, a result consistent with the no-winds case (Figure 4.3 (j)).

### 4.2.3 Mean and Turbulent Kinetic Energies

Looking at the mean kinetic energies of the uncoupled data as seen in Figure 4.10 (a)-(d) we see that the spatial patterns align well with that of the zonal velocities in the atmosphere similar to the no-winds case. However, as seen in the momentum budget there are no clear regions of geostrophic balance where the Coriolis term dominates and makes a large contribution to the mean kinetic energy. This aligns well with the kinetic energy at the end of the simulated time in Figure 4.10 (d) which resembles the same structures and magnitudes of the zonal velocities (Figure 4.8 (d)), compared to the no-winds case (Figure 4.4 (d) and Figure 4.2 (d)) where the Coriolis term also contributed to the mean kinetic energy. Moreover, the patterns of the uncoupled ocean are similar to that of the ocean within the no-winds case, suggesting that the constant weak winds do little to disrupt the kinetic energy within the ocean jet. The ocean jet kinetic energy also appears relatively much weaker than in the no-winds case, which suggests that coupling in this case has a much larger impact on the mean kinetic energy of the jet near the surface.

For the coupled data in Figure 4.10 (e)-(h) we see similar results in the atmosphere to that of the uncoupled case. For the ocean the accelerated winds and stronger wind stress is enough to cause an adjustment as seen in the velocity field Figure 4.8 (e)-(h). However, unlike the no-winds case, mean kinetic energy is added to the surface and aligns with the position of the strong surface wind speeds. The kinetic energy added in this case is also quite strong and dominates the background kinetic energy of the jet. Despite the large amount of kinetic energy being added at the surface early on, it quickly dissipates almost in entirety by the end of the simulated time.

To better understand the evolution of the mean kinetic energy, especially this dissipation in the ocean, we now look at the turbulent kinetic energy as seen in Figure 4.11. In the uncoupled data (Figure 4.11 (a)-(d)) we see in the atmosphere TKE growing from a plume out of the center of the filament and spreading out before remaining in an irregular, but steady structure. This is unlike the no-winds case (Figure 4.5) which became less spread out and more concentrated in the center over time. We also see weak TKE within a 500 m layer near the surface of the ocean, an expected result given a shear stress put

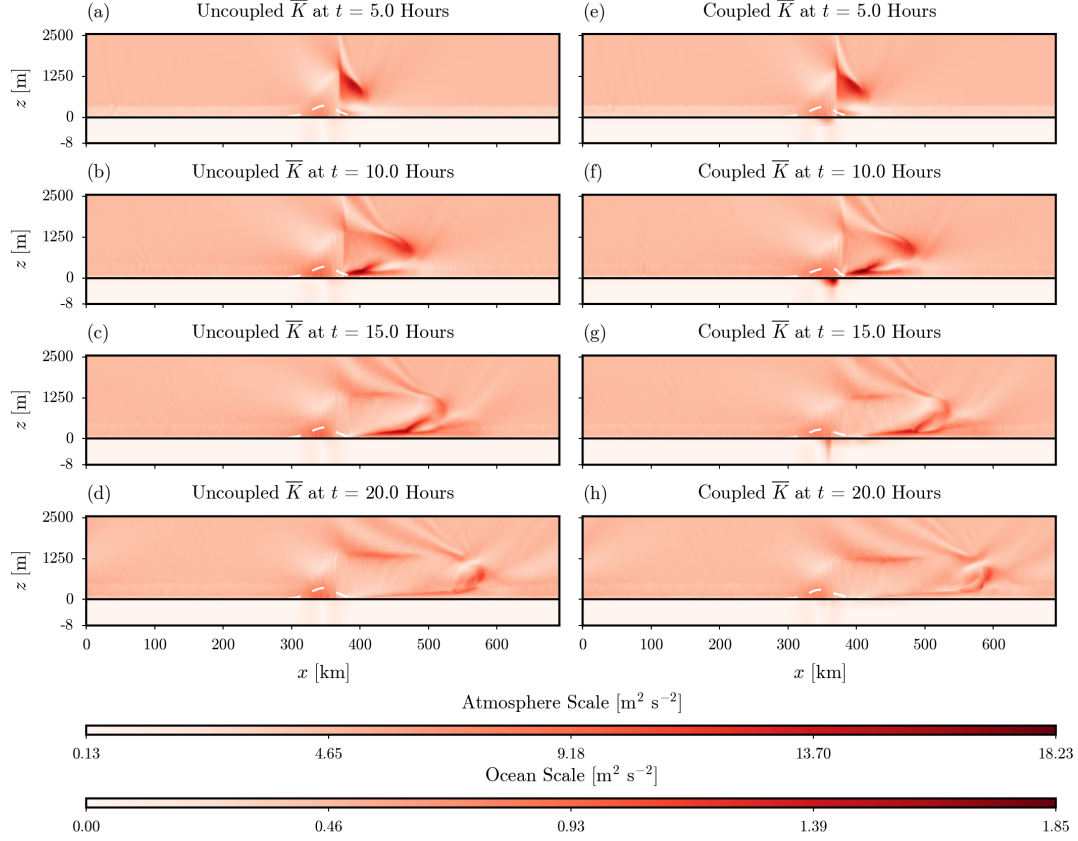


Figure 4.10: Evolution of the mean kinetic energy as described in (3.11) in both the coupled and uncoupled cases with  $3 \text{ m s}^{-1}$  zonal wind. The black line indicates where the simulations are separated between the ocean (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere. The white dashed line represents the qualitative shape and location of the SST anomaly, however is unrelated to the scale of the  $z$ -axis and should be seen as a visual aid only.

on the surface. However, the weak surface layer TKE is disrupted by the filament, and on the far side the surface TKE is reduced. This reduction in TKE seems analogous to previous works (Small et al., 2008; Sullivan et al., 2020) which have found that a cold front is capable of stabilizing winds blowing across it. In the uncoupled ocean once again there is no trace of TKE as expected from all previous results.

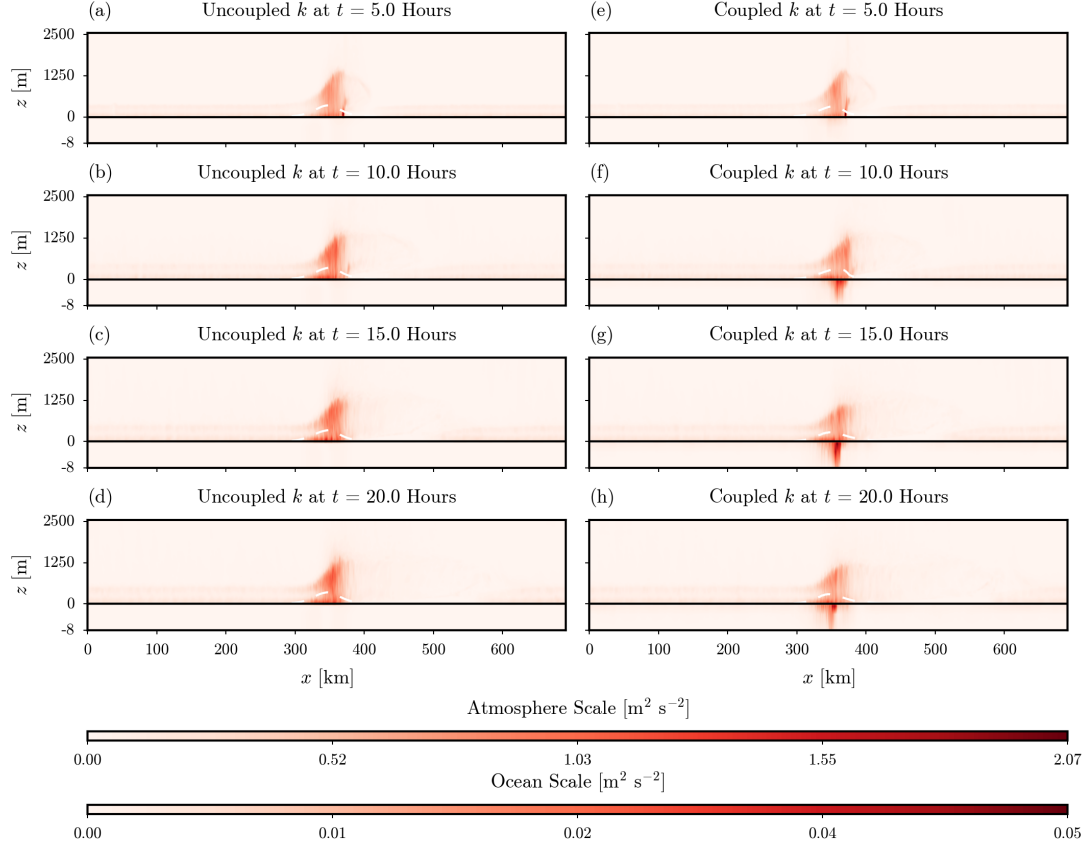


Figure 4.11: Evolution of the turbulent kinetic energy  $k$  as described in (3.12) in both the coupled and uncoupled cases with  $3 \text{ m s}^{-1}$ . The black line indicates where the simulations are separated between the ocean (below) and atmosphere (above). Note that the  $z$ -axis is stretched for the ocean as the significant response is much closer to the surface than for the atmosphere. The white dashed line represents the qualitative shape and location of the SST anomaly, however is unrelated to the scale of the  $z$  axis and should be seen as a visual aid only.

Within the coupled cases (Figure 4.11 (e)-(h)) for the atmosphere the overall pattern of the TKE plume is similar to that of the uncoupled case early on. As time goes on the concentration of TKE weakens over time, and by the end of the simulated time is notably weaker. The TKE of the coupled atmosphere was also found to be weaker in the no-winds

case (Figure 4.5 (e)-(h)), however in the presence of winds the effect is much more clear. In the ocean we see that the TKE aligns well with the TKE of the atmosphere on the surface and grows over time as the TKE of the atmosphere decreases, and also as the mean kinetic energy of the ocean decreases as seen in Figure 4.10.

To better understand the evolution of the TKE we look at the TKE budget as depicted in Figure 4.12. In the uncoupled data seen in panels (a)-(f) we keep our discussion to the atmosphere as the uncoupled ocean does not evolve. The time derivative and advection are weak, which is consistent to the no-wind case. The transport makes a notable contribution in reducing the TKE near the surface and increasing it above that, a similar effect seen to the no-winds case but much more present. Unlike the no-winds case the production plays a minor role making weak contributions about 500 m above the surface, while the buoyant production is much more defined and acts as the primary contribution. The dissipation is weak and seems to be of similar relative strength to that of the no-winds case.

The coupled simulations seen in Figure 4.12 (g)-(m) reveal very qualitatively similar results to the uncoupled simulations in the atmosphere, with the exception of the production term being notably weaker. Within the ocean the time derivative is quite strong with a notable feature occurring centered on the filament at the surface. The advection plays a minor role at the surface within the filament and the transport provides a weak contribution beneath it. The production completely dominates the evolution with a very strong contribution centered on the filament, and is only counteracted by the dissipation. The buoyant production is negligible.

#### 4.2.4 Discussion

Adding weak winds to the system allows us to better understand how the air sea interface near a filament-like warm SST anomaly is influenced. We are able to see how the dynamics of the system are comparable to that of the no-winds case. The atmosphere evolves with a warm plume of air consistently rising from the filament, but also being advected downwind, creating a growing region of warm air downwind. Moreover, we see a large initial perturbation above the filament in the zonal velocities when the surface convergence around the filament goes against that of the mean wind. However, as time goes on the mean wind overtakes the surface convergence and leaves behind a more stable configuration. The initial perturbation does not immediately dissipate and is instead advected downwind as it radiates gravity waves. The momentum budget reveals advection playing a more dominant role than in the no-winds case as well as reveals regions within the initial perturbation as well as above the filament where the Coriolis and pressure term somewhat

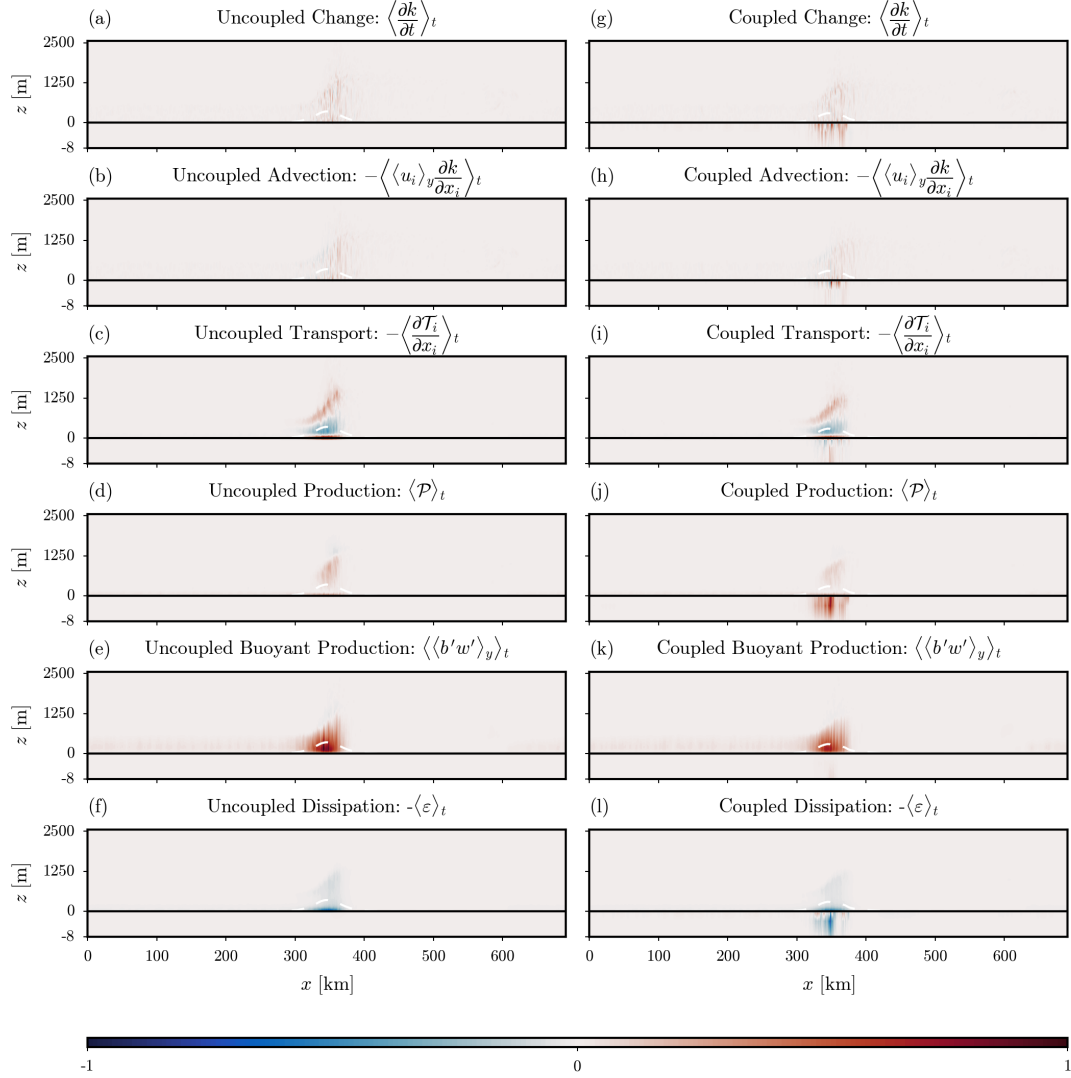


Figure 4.12: Normalized budget of the TKE equation (3.14) over time  $23 \text{ h} < t < 24 \text{ h}$ . The details of the average and layout are the same as the momentum budget in Figure 4.9.

cancel each other out. The energy analysis reveals the mean kinetic energy being significantly more similar to the zonal velocities than in the no-winds case, which is consistent with the Coriolis term playing less of a dominant role in the momentum budget. The TKE reveals a strong irregular shaped but consistent region above the filament where buoyant

production dominates alongside weak contributions from production and transport. We also see the TKE being reduced on the far side of the filament near the boundary which seems to point to a stabilization of the boundary layer with warm air crossing a cold front as seen in [Sullivan et al. \(2020\)](#).

Consistent with the no-winds case, the effects of coupling give qualitatively dramatic responses within the ocean as well as notable differences in the turbulent dynamics of the atmosphere. The mean winds prescribed are of similar magnitude to the surface winds seen in the coupled no-winds simulation, however the surface stress estimated by the quadratic drag was an order of magnitude weaker than the surface stress calculated by WRF. This resulted in a very weak, almost negligible response in the uncoupled ocean with winds compared to the coupled no-winds case, despite having similar magnitude surface winds. The coupled ocean response is also quite dramatic in the case with winds. The potential temperature noticeably decreases over the course of the simulation and the surface stress inducing an adjustment within the jet similar to the no-winds case. The momentum budget reveals a region just below the filament where dissipation is very strong, which is not present in the uncoupled ocean at all. The energy analysis shows a large gain in mean kinetic energy just below the region of the strongest surface winds from the atmosphere, but also a quick dissipation of this energy as time goes on. The turbulent kinetic energy has an almost inverse response, the region at the surface in the filament closest to the strongest winds grows in magnitude as the mean kinetic energy weakens. Note that the mean kinetic energy in the no-winds case seemed to be driven by the jet adjusting while here it seems to be imparted by the atmosphere above. The TKE budget in the ocean reveals that production completely dominates the evolution of the TKE which is expected from the evolution of the mean and turbulent kinetic energy.

# Chapter 5

## Conclusions and Future Work

### 5.1 Conclusions

This thesis shows the evolution and dynamics of the uncoupled and coupled air-sea interaction near a warm SST filament. The results can be broken up into two parts: the response of the atmosphere to the SST, and the coupled feedback on the ocean and back on the atmosphere. Previous works have found strong positive spatial and temporal correlation between near surface wind speeds and SST at ocean mesoscales which come with a conclusion that the SST drives the atmosphere (Small et al., 2008; Chelton and Xie, 2010; Seo et al., 2023). This thesis finds consistency with these results, showing a quick and strong atmospheric response that is able to achieve a somewhat steady state even within the short 24 hour time scale of the simulations. In the no winds case we saw surface convergence over the filament that became the source of two convective cells on either side. In the winds case we were able to see effects of wind over a warm and cold front. Similar to the no winds case surface convergence over the warm filament was observed driving air upward and then being advected downwind. In addition the winds case displayed a strong acceleration of the near surface winds over the filament with a deceleration after returning to the baseline SST on the far side. Alongside these phenomena the winds also allowed for the formation of an atmospheric jet at  $z \approx 1250$  m on the far side of the filament which is consistent with observation of low level jets seen in systems where warm air blows over a cold SST front (Small et al., 2008). Lastly a reduction in the TKE near the surface is observed on the far side of the filament which is potentially an example of the near surface stability increasing, a result seen in the submesoscale results of Sullivan et al. (2020) with air blowing over a cold SST front.

Alongside the consistency with previous works in the atmospheric response we also have the exploration of the coupled effects of these responses on the ocean, and then also the feedback into the atmosphere. Across the no winds and winds cases a dramatic difference was found between the uncoupled and coupled ocean. In the no winds case the uncoupled ocean displayed no evolution which is expected as the ocean jet was initialized in an inertially and gravitationally stable configuration. In contrast, within the coupled simulations the immediate wind stress provided by the atmosphere was able to accelerate the ocean surface causing an adjustment in the otherwise stable ocean jet. After the adjustment took place mean kinetic energy was generated and even on these short time scales the jet started to break down. Despite the adjustment in the jet, the potential temperature in the ocean was able to stay fairly consistent throughout the 24 hours, which resulted in a lack of qualitative differences between the coupled and uncoupled atmosphere. However, even with the minimal SST adjustment an analysis of the TKE revealed there to be differences in the turbulent structures of the atmosphere, showing sensitivity to this SST evolution.

The winds case showed an even more notable difference between the coupled and uncoupled simulations. First, an important note that the uncoupled ocean was provided a uniform constant in time wind stress based on the initial  $3 \text{ m s}^{-1}$  zonal wind and a quadratic wind stress and dimensionless drag coefficient based on the findings of Kara et al. (2007). Similar wind speeds were found over the filament in the no winds case, however, when the ocean was coupled to these winds of similar magnitude a much stronger stress was induced on the ocean as calculated by WRF. The stress calculated by the quadratic wind stress was an order of magnitude smaller than the stress calculated by WRF, despite the very similar wind speeds. This resulted in the weak winds case of the uncoupled ocean having very little evolution and no strong adjustment of the jet as seen previously. Given that there was very little evolution of the jet, the evolution of the coupled ocean in the winds case was once again dramatically different. The coupled ocean experienced a significant amount of cooling near the surface over the 24 hours and also immediately strongly reacted to the stress induced by the accelerated winds over the surface. This once again led to the jet adjusting. Despite the jet adjusting internally, the surface winds were able to impart a large amount of mean kinetic energy into the top level of the ocean early on. The mean kinetic energy however did not persist and was quickly transferred into a turbulent kinetic energy just below the surface of the filament. Similar to the no winds case the coupled atmosphere did not have a large scale qualitative difference to the uncoupled atmosphere. However, an exploration of the TKE revealed a weaker generation of TKE above the filament.

Overall these results depict a complicated air-sea interaction near a filament-like SST distribution with results similar to that of both winds moving over a cold to warm front,

as well as a warm to cold front seen in previous works, (Spall, 2007; Small et al., 2008; Chelton and Xie, 2010; Sullivan et al., 2020). In addition, the significance of dynamic air-sea coupling near SST fronts are shown, with the coupling playing a significant role in the immediate evolution of the ocean even within short time scales and weak winds. The coupling is less impactful of the qualitative structures seen in the atmosphere, however still affects the turbulent structures in the vicinity of the front.

## 5.2 Future Work

Many aspects of this thesis can be expanded upon, the first and foremost being the inclusion of ocean surface currents to the coupling system. As mentioned, the surface parameterization WRF uses to calculate the surface stresses assume the ocean surface is static. This resulted in a coupled system where the atmosphere ignored the presence of ocean surface currents, and likely often overestimated the stress on the ocean after the surface had been influenced by the winds. The thesis used the assumption that the surface currents would remain much smaller than the near surface winds, however more recent works have shown the importance of surface ocean currents even playing a role in atmospheric evolution as seen in Seo et al. (2023).

A limitation of this work was due to the computational inefficiency of the coupling between the models. Despite having qualitative differences within the 24 hours of simulated time, the dramatic changes occurring in the ocean suggest running the simulations for longer could reveal an even more impactful effect of coupling. However, running the simulations for longer became computationally infeasible within the time frame of this thesis. In addition a more efficient coupling model that fully utilized the parallelizing capability of Oceananigans could allow for the exploration a larger parameter space including more variations of the wind speeds, variations of the length scale of the filament, and variations of the maximum temperature of the filament, common parameter spaces seen in similar works (Spall, 2007; Sullivan et al., 2020, 2021).

Finally the analysis of a single front as opposed to a filament like SST distribution could be performed as a simpler experiment to better understand mechanics involved. A filament like SST distribution was partly chosen due to a limitation of WRF requiring periodic lateral boundaries to avoid artifacts of waves bouncing off open boundaries. The single front analysis could provide better comparison of other previous works (Spall, 2007; Small et al., 2008; Sullivan et al., 2020) as well as more studied regions of the ocean including the north edge of the Gulf Stream, the Kuroshio, and the equatorial tongues of the Atlantic and Pacific (Small et al., 2008).

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# APPENDICES

| Notation                       |                                               |
|--------------------------------|-----------------------------------------------|
| $u_i$                          | Velocity Components for Index Notation        |
| $u, v, w$                      | Velocity Components                           |
| $u_*$                          | Surface Friction Velocity                     |
| $\theta$                       | Potential temperature                         |
| $T$                            | Absolute temperature                          |
| $Q_h$                          | Surface Upward Heat Flux                      |
| $p$                            | Pressure                                      |
| $\Phi$                         | Normalized Pressure                           |
| $\tau_{ij}$                    | Kinematic Stress Tensor                       |
| $\mathcal{T}_i$                | Transport Component $i$                       |
| $\mathcal{P}$                  | Turbulent Production                          |
| $\varepsilon$                  | Turbulent Dissipation                         |
| $X^{(a)}$                      | ‘ $X$ ’ in the Atmosphere                     |
| $X^{(o)}$                      | ‘ $X$ ’ in the Ocean                          |
| $\langle X \rangle_{ab...}$    | ‘ $X$ ’ Averaged in $a, b...$                 |
| $\langle X'^2 \rangle_{ab...}$ | Variance of $X$ in $a, b...$                  |
| $\langle X'Y' \rangle_{ab...}$ | Covariance of $X, Y$ in $a, b...$             |
| $X_{10}$                       | ‘ $X$ ’ at 10m                                |
| $X_0$                          | A Prescribed Constant Associated with ‘ $X$ ’ |

Table 1: Notation

| Symbol           | Value                | Meaning                                                       |
|------------------|----------------------|---------------------------------------------------------------|
| $L_x$            | 691.2                | Domain Size [km]                                              |
| $L_y$            | 76.8                 | Domain Size [km]                                              |
| $L_z^{(a)}$      | 20                   | Atmosphere Depth [km]                                         |
| $L_z^{(o)}$      | 1.024                | Ocean Depth [km]                                              |
| $\ell_x$         | 153.6                | Filament Scale [km]                                           |
| $\ell_z$         | 1.024                | Filament Scale [km]                                           |
| $c_j$            | 6                    | Filament Scale Constant                                       |
| $\rho_0^{(a)}$   | 1.225                | Average Surface Density [kg m <sup>-3</sup> ]                 |
| $\rho_0^{(o)}$   | 1026                 | Average Surface Density [kg m <sup>-3</sup> ]                 |
| $\Gamma_d^{(a)}$ | $9.8 \times 10^{-3}$ | Adiabatic Lapse Rate [°C m <sup>-1</sup> ]                    |
| $\Gamma^{(a)}$   | $6.7 \times 10^{-3}$ | Prescribed Lapse Rate [°C m <sup>-1</sup> ]                   |
| $N^{(o)}$        | $3.7 \times 10^{-3}$ | Background Buoyancy Frequency [s <sup>-1</sup> ]              |
| $\alpha^{(o)}$   | $2 \times 10^{-4}$   | Thermal Expansion Coefficient [°C <sup>-1</sup> ]             |
| $C_D$            | $1.6 \times 10^{-3}$ | Drag Coefficient                                              |
| $C_p^{(o)}$      | 3991                 | Specific Heat Capacity [J °C <sup>-1</sup> kg <sup>-1</sup> ] |
| $g$              | 9.8                  | Gravitational Acceleration [m s <sup>-2</sup> ]               |
| $f$              | $1 \times 10^{-4}$   | Coriolis Parameter [s <sup>-1</sup> ]                         |

Table 2: Parameters and Constants