Transport and Mixing of Water Masses Across the Southeast Caribbean Ocean Imaged by Seismic Reflection Data

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TRANSPORT AND MIXING OF WATER MASSES ACROSS THE SOUTHEAST CARIBBEAN OCEAN IMAGED BY SEISMIC REFLECTION DATA

Approved by:

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TRANSPORT AND MIXING OF WATER MASSES
ACROSS THE SOUTHEAST CARIBBEAN OCEAN
IMAGED BY SEISMIC REFLECTION DATA

A Thesis Presented to the Graduate Faculty of the
Moody School
Southern Methodist University
in
Partial Fulfillment of the Requirements
for the degree of
Master of Science
with a
Major in Geophysics
by
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December 16, 2023
ACKNOWLEDGMENTS

The culmination of this work would not have been possible without my advisor and now friend, Prof. Beatrice Magnani, who cheered on my successes, questioned my every scientific decision, and built me up from my failures. This thesis turned into much more than we originally planned and I am incredibly thankful for that, as it led to more discoveries and an abundance of challenges. All of these avenues of exploration broadened my knowledge of both science and the power of persistence as we explored many failed directions, which eventually led to the thesis here today. Furthermore, I am incredibly grateful to the rest of my committee members, Dr. Matt Hornbach and Dr. Chris Hayward, for contributing important insight into ocean dynamics and seismic methods and showing a genuine curiosity in a study outside your general realm of expertise.

Additionally, this research would not have been accomplished without the help of my family and friends who supported me along the way. Throughout the ups and downs of life, they helped me see this through and grow in the process. I express my dearest gratitude to my mom, dad, and sisters. Although they wondered if I would ever finish this study, they continued to motivate me to keep going. Finally, I want to thank my lovely girlfriend Annie, who without her, I might not have developed a new perspective on life and... not eaten as many great dinners! She showed me a life filled with joy, motivation, intelligence, creativity, and experiences that I will carry with me for the rest of my life.
The Caribbean Sea serves as a major pathway for global thermohaline circulation (THC), which is a complex and vital component of the Earth’s climate system, influencing global heat distribution and oceanic circulation. Though relatively stratified, it is the boundary layer that distributes mass and temperature between the surface waters and the deep ocean where we observe various multiscale mixing processes from mesoscale to fine-scale. In regions where bathymetry is shallower and mechanical mixing forces, such as winds and tides, are more dominant, diapycnal diffusivity is typically stronger, driving vertical mixing. This type of mixing occurs at small scales, typically as internal waves break within the internal ocean, making it difficult to quantify and observe. Through the combination of seismic images and oceanographic data, known as seismic oceanography, we can qualitatively and quantitatively observe the variability of the ocean’s internal wave field and its diverse components, which include the turbulent and internal wave subranges from vertical displacement spectra. Exploiting these subranges allows us to quantify vertical mixing behaviors across isopycnal layers, effectively representing the cascade of energy for mixing. Quantitatively constraining these energy components is essential to comprehensively understand the total energy budget of the THC.

This research focuses on mapping and quantifying diapycnal diffusivity in the southeastern Caribbean Sea, a region characterized by the convergence of two primary water
masses, North Atlantic water (NAW) and South Atlantic water (SAW), as they spill into the Caribbean Sea through the Lesser Antilles passages. This convergence introduces perturbations in temperature, salinity, and nutrients, resulting in the formation of the Caribbean Current. The current’s predominant westward direction, driven by surface winds, is influenced at depth by interactions with deeper water masses and the irregular coastal bathymetry.

We utilize five seismic profiles, totaling approximately 1000 km in length, collected in 2004 in the southeastern Caribbean Sea, in conjunction with oceanographic data and models, to identify fine-scale and mesoscale structures in the thermocline (300-1200m). Additionally, we quantify diffusivity along sections of each seismic profile using displacement slope spectra techniques calculated from seismic reflectors.

Our analysis reveals average diffusivity values of $\sim 10^{-5.2} \text{ m}^2/\text{s}$, which are comparable to the global average of $10^{-5} \text{ m}^2/\text{s}$ of the open ocean. Notably, we observe an increase in diffusivity to values of $>10^{-4} \text{ m}^2/\text{s}$ in regions characterized by elevated or rugged bathymetry, confirming the hypothesis of the existence of regions of concentrated, high diffusivity (i.e., hot spots) and greater mixing. The existence of these regions has been advocated to explain the discrepancy between the observed average diapycnal diffusivity value of $>10^{-5} \text{ m}^2/\text{s}$ and the theoretical globally averaged values of $>10^{-4} \text{ m}^2/\text{s}$ derived from an advective diffusive balance. Our findings contribute to a better understanding of mixing processes in the SE Caribbean Sea, shedding light on the quantification of diapycnal diffusivity and its spatial variations, ultimately enhancing our grasp of the THC’s energy budget.
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6.8 The full BOL30 seismic profile with black box identifying the area of interest for subsequently tracked reflectors and resulting diffusivity calculations. Estimated turbulent diffusivity, labeled $\log k_t$, was calculated for each individually tracked reflector. The reflector results were blended using a 2km by 20m, half overlapping window, to create a better qualitative understanding of larger regions of observation.
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This is dedicated to everyone who motivated, challenged, and supported me throughout this journey to mold me into a good geophysicist and even better person.
1.1. Motivation

Oceanic mixing in and around the thermocline is crucial to oceanic dynamics, as it redistributes mass, nutrients, heat, and salt across isopycnals or density layers of the ocean while contributing to global thermohaline circulation (Wunsch, 2002; Waterhouse et al., 2014). This circulation system, which encompasses the vertical movement of heat and salt, exerts a profound influence on climate patterns and mass transport on a global scale as it contributes to meridional overturning and the heat budget of the ocean (Kunze et al., 2006; Wüst, 1964; Rahmstorf, 2006; Wunsch, 2002). As energy transfers from large-scale to small-scale flow, turbulence induces irreversible changes to the water masses (Wunsch, 2002). Small-scale turbulent mixing or overturning across density layers in the ocean induces vertical and horizontal motions, resulting in the mixing of water masses across stratified density isopycnals, commonly described as diapycnal diffusivity ($K_{\rho}$) (Munk and Wunsch, 1998; Osborn, 1980). This irreversible mixing process acts as a driving force in mediating the transfer of heat, salt, and nutrients to the deeper ocean (Figure 1.1) (Rahmstorf, 2006; Holbrook et al., 2003; Wüst, 1964). It is then important to understand the trends of diapycnal diffusivity as it is fundamental in the dynamics of the deep ocean and has broader implications for global ocean circulation patterns (Fortin et al., 2016).

The spatial distribution of diapycnal mixing in and around the thermocline exhibits high variability, being influenced by both depth and location (Kunze et al., 2006; Waterhouse et al., 2014). It is recognized that where diffusivity occurs in the open ocean is spatially heterogeneous or relatively patchy, and typically has values at or below $10^{-5} \text{ m}^2/\text{s}$ in the main thermocline or oceanic interior (Kunze et al., 2006; Waterhouse et al., 2014). However,
an average global diapycnal diffusivity of $10^{-4} \, m^2/s$ is required to maintain stratification and produce overturning within the thermocline (Munk and Wunsch, 1998; Waterhouse et al., 2014; Wunsch, 2002). The magnitude difference between diffusivity levels indicates there must be regions of increased mixing to balance the oceans (Waterhouse et al., 2014).

It has been proven in the past few decades that diffusivity levels can be multiple magnitudes greater in areas with shallow or sharp topography, through the use of high-resolution vertical profilers that measure microscale variations of temperature and particle velocity (Lueck and Osborn, 1985; Lueck and Mudge, 1997; Polzin et al., 1997). While patches of heightened diffusivity are more likely to appear in close proximity to shallow bathymetry, their spatial distribution remains fragmented, displaying variations influenced by both depth and geographical location (Kunze et al., 2006; Tang et al., 2021; Waterhouse et al., 2014). This variable distribution of enhanced diffusivity necessitates high resolution, laterally continuous observations to clearly identify the spatial extent of these areas of enhanced diffusivity, often referred to as "hot-spots," which contribute to vertical shearing, resulting in the dissipation of internal and turbulent wave energy (Lueck and Mudge, 1997; Waterhouse et al., 2014). Oceanic passages controlled by shallow topography, such as the Luzon Strait, Faroe Islands, Strait of Gibraltar and the Lesser Antilles, are ideal locations to study mixing hot spots and related oceanic mixing features within the thermocline, as the internal waves in these regions are affected by a combination of mechanical forces, such as wind and tides along with shallowing bathymetry (Gregg, 1998; Ivey, 2004; Mauritzen et al., 2005; Munk and Wunsch, 1998; Tian et al., 2009). The complexity results in a variety of mixing behaviors that require dense, laterally continuous measurements, to properly observe the extent and location of the diverse mixing features.

The Caribbean Sea serves as a major pathway for the meridional overturning circulation (MOC), which transports heat and salt from the fresher South Atlantic water (SAW) and is augmented by the more saline North Atlantic water (NAW) as part of the global ocean conveyor belt (Kirchner et al., 2008). The waters enter the Caribbean Sea driven by the retroflection of the North Brazil Current (NBC), which moves northwestward towards the
Lesser Antilles, as a boundary current along the South American Coast (Schott et al., 1998; Johns et al., 2002). The westward inflow of the north and south Atlantic water, which enters the basin through the narrow and shallow passages bounded by the volcanic islands of the Lesser Antilles which form a bathymetric barrier extending from north to south (Figure 1.2) (Morrison and Nowlin Jr, 1982; Johns et al., 2002). The total inflow of water into the Caribbean has been the subject of various studies, resulting in some uncertainty. To address this, I rely on the findings of Rhein et al. (2005) and Kirchner et al. (2008), which estimate an average total inflow of approximately 19.4 Sv (13.4 Sv of SAW). Among the southern islands of the Lesser Antilles, namely Grenada, St. Vincent, St. Lucia, Dominica, and Guadalupe, the majority of water (12.4 Sv) enters the Caribbean, with decreasing mean inflow observed in the mentioned order (Kirchner et al., 2008). The islands act as stirring rods on the waters flowing into the Caribbean Sea (Mukherjee et al., 2023; Tang et al., 2022), and the combination of complex open-ocean eddy formation and shallow bathymetry produces submesoscale turbulence in the upper ocean (Mukherjee et al., 2023), making this an excellent site to study mechanical mixing within the ocean. The focus of this research is to identify oceanic mixing features that contribute to the diffusivity of heat, salt, and nutrients in the Caribbean Sea, to quantify diffusivity levels in the region, to highlight areas of increased mixing, and to test whether these areas are associated seafloor features and/or other mixing-driving processes. My analysis of seismic reflection data combined with oceanographic data and models provide evidence of various internal oceanic mixing features such as eddies, lee waves, and thermohaline staircases, present in the southeastern Caribbean Sea. We also estimate an average diapycnal diffusivity similar to the global average of the order of $10^{-5} \, m^2/s$ for the upper $\sim 1000m$ and identify regions of increased diffusivity associated with sharp bathymetry (Munk and Wunsch, 1998; Kunze et al., 2006).

I use a combination of seismic reflection data and oceanographic models to investigate the dynamic fine-scale mixing process of the thermocline and its interaction with mesoscale features in the southeastern Caribbean Sea. The seismic data consists of five north-south striking, seismic profiles that intersect the generally westward-flowing Caribbean current at
a high angle. These profiles are complemented by in-situ Expendable Bathythermograph (XBT) casts collected along specific seismic lines data from the Hybrid Coordinate Ocean Model (HYCOM) database. This study aims to investigate regions of increased diffusivity driven primarily by bathymetry and identify mesoscale features that contribute to the complex interaction of the thermohaline circulation (THC) and mixing across the southeastern Caribbean Sea.

Figure 1.1: General mixing schematic showing the interaction between surface-driven forces, thermohaline processes, and bathymetry on the exchange of mass and heat within the ocean. Each process affects the energy levels driving the global overturning of the ocean.
Figure 1.2: Map of the study area showing the seismic profiles used in this study. Top: bathymetric map showing in-situ XBT cast locations, with key physiographic regional features. Bottom: colors show surface water velocity averaged for the month of May 2004 overlain by northward velocity (contours). Data extracted from HYCOM database generated in Ocean Data Viewer (ODV). Most of the color-shaded values are negative indicating a general westward flowing direction, with peaks in westward velocity at the sills between the Antilles islands, which allow Atlantic inflow into the Caribbean Sea.
1.2. Study Area and Previous Work

Stratified waters in the Caribbean are characterized and classified by their relative temperature, salinity, oxygen, and other oceanic properties within specific isopycnal ranges and layers (Morrison and Nowlin Jr, 1982; Rhein et al., 2005). These waters retain their original water mass characteristics for variable time frames based on mechanical forces (winds and tides) and density-driven forces such as thermohaline circulation and internal waves (Wüst, 1964; Wunsch, 2002). The isopycnal layers in are defined by neutral density, $\sigma$ (kg/m$^3$), which is scaled to 1000 decibars. In the Caribbean sea the water masses include from shallow to deeper depths: the warm tropical surface water (SW) with a neutral density less than 24.5 ($\sigma < 24.5$), followed by the salinity maximum water (SMW) ranging from $\sigma = 24.5$ to 26.3. Between depths of $\sim 200$-400m, is the Tropical Atlantic Central Water (TACW) with a density between $\sigma = 26.3$ and 27.1. The Antarctic intermediate water (AIW), characterized by its relatively fresher composition, occupies the density range of $\sigma = 27.1$ to 27.6 from $\sim 400$-1000m. Beyond a density of 27.6, is the Atlantic deep water (ADW), which exhibits relatively uniform and well-mixed characteristics (Figure 1.3).

The interaction and mass transport between these different water masses and across the Caribbean is not fully understood. Our understanding of mass transport is most comprehensive for surficial processes, which can be observed through measurements of sea surface temperature (SST), sea surface salinity (SSS), and sea surface height (SSH). These measurements provide evidence of heat and nutrient transport at the surface and are routinely monitored through remote sensing techniques in both spatial and temporal domains. However, the internal portion of the ocean, where localized and mesoscale features interact and contribute to small-scale mixing, is less sampled and presents more significant challenges for investigation (Jouanno and Sheinbaum, 2013).

Previous studies of the internal ocean carried out using hydrographic and velocity measurements, have mapped physical oceanographic features, such as temperature, salinity, and silicates, along single transects crossing the Caribbean from north to south (Joyce et al., 2001; Casanova-Masjoan et al., 2018). While not explicitly addressing diffusivity, these studies ob-
tained velocity measurements to observe general mass transport. These analyses do not explain dynamical mixing processes but rather provide localized observations at sparse horizontal resolution, to provide general mass movement (Joyce et al., 2001; Casanova-Masjoan et al., 2018; Parra et al., 2023). The only estimates to constrain vertical diffusivity in the area are derived from velocity profiles and conductivity, temperature, depth (CTD) sensors (Kunze et al., 2006). The estimates by Kunze et al. (2006) are based on four individual calculations derived from velocity and CTD casts acquired in 2003 along a singular transect at 66°W, and resulted in an average diapycnal diffusivity estimate for the internal ocean in the upper 1000m of \( \log_{10}(K_\rho) = -5.3 \ m^2/s \ (0.5 \times 10^{-5} \ m^2/s) \).
Figure 1.3: (a) General water mass depths entering through the Windward Antilles Islands below 15°N derived from HYCOM. The layers represent the water mass density ranges for \( \sigma = 24.5-26.3 \) (Salinity Maximum Water), \( \sigma = 26.3-27.1 \) (Tropical Atlantic Central Waters), and \( \sigma = 27.1-27.6 \) (Antarctic Intermediate Waters), and \( \sigma = 27.6 \) (Atlantic Deep Waters). (b) Highlights the mean inflow, measured in Sverdrups, through key passages, driven by wind and the Meridional Overturning Cell (MOC). Figure adapted from (Johns et al., 2002). The inflow of the different water masses is heavily influenced by the seafloor depth, and current velocity.
2.1. Seismic Oceanography

Marine seismic reflection surveys in the modern age are usually performed using an acquisition vessel that typically fires an airgun (or an array of airguns) that generates a source signal, and one (or multiple) streamer(s) (cable-containing hydrophones spaced in groups) to record the signal and noise at a relatively constant depth. Often an array of different-sized airguns is used to produce different frequencies and waveforms that are used to reduce specific source-generated noise (Cox et al., 2020). In its simplest form, the seismic reflection technique is designed to generate seismic waves and record the wavefield that is reflected off subsurface interfaces. Reflection seismology, therefore, relies on changes in density and seismic wave velocity between layers for imaging. When seismic waves traveling through the subsurface encounter an interface with changes in density and velocity, some of the wave energy will be reflected off the interface and some will transmit through the boundary. The amount of energy (and therefore the amplitude) of the reflected wave is determined by the reflection coefficient $RC$, which, for near vertical incident angles, is defined as:

$$RC = \frac{\rho_2 v_2 - \rho_1 v_1}{\rho_2 v_2 + \rho_1 v_1}$$ \hspace{1cm} (2.1)$$

where $\rho$ and $v$ are the density and seismic wave velocity of each medium, respectively. The product of density $\rho$ and velocity $v$ is also referred to as the impedance contrast $Z$ (Sheriff and Geldart, 1995):

$$Z = \rho v$$ \hspace{1cm} (2.2)$$

As reflected arrivals are recorded and mapped through the seismic reflection survey, an image of the subsurface is built. Reflector continuity, amplitude, and attribute changes can
be analyzed and related to variations in density and velocity properties of the media at interfaces by means of seismic inversion.

Though designed to image geological features below the seafloor, data acquired with seismic reflection methods have recently been used to identify fine-scale thermohaline structures, internal mixing, and other density or velocity variable oceanic features, with a method known as seismic oceanography (SO) (Holbrook et al., 2003). SO combines seismic reflection data with oceanographic properties to qualitatively and quantitatively assess energy and mass exchange resulting from various mechanical forces (winds and tides), density-driven forces, such as thermohaline circulation and internal waves, and physical features, such as sharp bathymetry. In SO, boundary layers are driven primarily by temperature and salinity contrasts, which directly relate to the amplitude of the reflected seismic waves, assuming the layers are stable for a period long enough to image (Holbrook et al., 2003; Buffett et al., 2009). Because SO applies identical collection and processing techniques of conventional marine seismic surveys while focusing on the water column, it provides a unique opportunity to reprocess any existing marine seismic reflection dataset to study the water column.

Oceanographic interpretations strongly depend upon in-situ oceanographic measurements and remote sensing of sea surface properties (Klaeschen et al., 2009). Similar to borehole data, in-situ measurements of salinity, density, and temperature provide direct information only at or near the site, and their ability to map the lateral variability of internal oceanic features is limited by data availability (Nakamura et al., 2006). On the other hand, remote sensing measurements provide real-time salinity (SSS), sea surface height (SSH), and temperature (SST) values across extensive areas, while their information is depth-limited. Measurements of SST and SSS have been combined with in-situ measurements to make predictions of upwelling and downwelling behavior in the ocean (Correa-Ramirez et al., 2020; Rueda-Roa and Muller-Karger, 2013; Rueda-Roa et al., 2018).

By collecting data over large areas, reflection seismology can map a physical contrast as a seismic reflector with a lateral resolution of fewer than 10m, and relate the reflection amplitude to an expected acoustic impedance contrast and to the physical properties (Páramo and
Holbrook, 2005), thereby filling the data gap between in-situ measurements, connecting surface information with deeper features, and in general providing the framework within which a comprehensive interpretation of processes is possible. In the water column, the acoustic impedance contrasts are likely caused by temperature and salinity contrasts introduced by various water masses (Holbrook et al., 2003; Tang and Zheng, 2011).

Coincident and simultaneous acquisition of expendable bathythermograph (XBT) and conductivity, temperature, depth (CTD) profiles with seismic reflection data was first tested by (Nandi et al., 2004) in the Norwegian Sea. The results of the study verified that the impedance contrasts as derived from temperature and salinity variability coincide in the water column with seismic reflection impedance contrasts (Nandi et al., 2004) proving that the seismic oceanographic method can image water mass boundaries, resolve the lateral and vertical extent of the fine structure in the ocean (Nandi et al., 2004), and therefore represents a powerful tool to start addressing the processes at the root of these features.

SO has evolved as a successful technique for identifying boundaries of mesoscale features, including bathymetric disruptions in mixing, eddies, water masses, thermohaline staircases, as well as finer-scale features like internal waves, oceanic fronts, and water mass boundaries. It provides a quasi-snapshot of the internal structure of the ocean (Holbrook et al., 2003; Nandi et al., 2004; Holbrook and Fer, 2005; Buffett et al., 2009; Eakin et al., 2011; Tang and Zheng, 2011; Alford et al., 2015; Quentel et al., 2010; Dickinson et al., 2020; Gunn et al., 2021; Tang et al., 2022). Since its inception, seismic oceanography has become a bridge between the oceanographic and geophysical communities and has been playing an increasing role.

2.1.1. Quantifying Reflection Data: Isopycnal Slope Spectra

Oceanic circulation involves intricate interactions of mechanical and density-driven forces across various scales, from large-scale flows to minute turbulent mixing (Batchelor, 1959; Munk and Wunsch, 1998). Energy transfer occurs in both vertical and horizontal directions. As energy cascades from larger to smaller scales, it transitions into small-scale variance, eventually leading to molecular diffusion (Klymak and Moum, 2007). One of the primary
mechanisms for vertical heat and salinity exchange across isopycnal surfaces is through the breaking of internal waves. This breaking induces turbulent motions, establishing a robust density stratification in the process (Wunsch, 2002). However, it is worth noting that turbulent motions can arise from various sources, not just the breaking of internal waves. Examples include shallow wind-driven turbulence and shear instabilities (Wunsch, 2002).

The understanding of these interactions has been greatly aided by spectral analysis, a mathematical technique first introduced by Fourier to break down functions into their constituent frequencies. In oceanography, spectral analysis has been instrumental in disentangling the intricate relationship between spatial and temporal scales, particularly in the context of energy cascades (Batchelor, 1959; Garrett and Munk, 1975; Klymak and Moum, 2007). Obtaining in-situ measurements such as moorings or towed instruments to track temperature or velocity fluctuations collected over space and time, two unique subranges associated with internal wave mixing and turbulent mixing become apparent (Kolmogorov, 1941; Batchelor, 1959; Garrett and Munk, 1975). The energy distribution, often from vertical or horizontal displacements of isotherms, is plotted along the wavenumber \( k_x \), which defines the spatial frequency of a wave (Klymak and Moum, 2007). The subranges are identified based on their respective power-law slope that best fits the data.

Beginning with internal waves, which occupy the lower wavenumbers \( (k_x < 10^{-2} \text{ m}^{-1}) \) on horizontal lengths of 100-1000m, the Garrett-Munk 75 (GM75) model is used to identify the internal wave subrange (Garrett and Munk, 1975). This model was one of the first, robust, kinematical models to identify a specific internal wave subrange that fits along a singular power law in relation to wavenumber from towed instruments. They identified a power law exponent of -2.5 or \( k_x^{-2.5} \), where \( k_x \) is generally \( < 10^{-2} \text{ m}^{-1} \) using towed and moored instruments to track vertical displacements of isotherms (Figure 2.1). This exponent reflects the scaling behavior of internal wave energy as a function of wavenumber, providing crucial insights into the distribution of energy across different spatial scales within the ocean (Klymak and Moum, 2007). Moreover, the GM75 model has paved the way for subsequent research and the development of more sophisticated models and observational techniques.
for studying internal waves (Gregg, 1989; Henyey et al., 1986). These advancements have enhanced our ability to comprehend the complex interplay of forces, such as buoyancy, stratification, and bathymetry, that govern the generation, propagation, and dissipation of internal waves in the world’s oceans.

Additionally, the turbulent subrange, located within the ocean’s energy spectrum, becomes pronounced at higher wavenumbers \(k_x > 10^{-2} \text{ m}^{-1}\) and corresponds to shorter length scales. Where turbulent forces become dominant, it reflects different physical processes primarily associated with small-scale mixing (Klymak and Moum, 2007). We identify the subrange using the Kolmogorov power-law constant of -5/3, which serves as a pivotal theoretical framework for comprehending turbulence at small scales. This finding describes how the turbulent energy of fluids varies with wavenumber, with the energy spectrum following a power-law distribution characterized by an exponent of -5/3. Central to this idea is the notion of the inertial subrange, a region where energy cascades from larger to smaller scales, governed primarily by inertial forces (Batchelor, 1959; Klymak and Moum, 2007). This understanding of cascading energy led to the formation of the power-law constant used for identifying turbulence.

Subsequently, in 1959, the Batchelor model expanded upon the Kolomogrov constant to provide the theoretical framework to quantify small-scale variations of quantities like temperature and velocity in a turbulent fluid (Batchelor, 1959). The Batchelor model related the turbulent subrange to the rate at which turbulence dissipates energy or the turbulent dissipation rate \(\epsilon\) \((W/kg)\), as it cascades to smaller eddies. This finding laid the groundwork for quantifying energy and mass exchange from wave displacement spectra in the internal ocean.
Figure 2.1: Figure adapted from (Nikurashin et al., 2013) showing the signature GM75 -2.5 power law slope associated with modeled depth and rough or flat topography response. The general horizontal wave spectra response for mesoscale feature and submesoscale features are highlighted. Notice that the energy level of the overall slope decays with depth but is 1 to 2 magnitudes greater in rough topography for the same depth besides the surface.

More recently, seismic oceanographic profiles, have emerged as a robust tool for estimating both internal wave and turbulent mixing phenomena by leveraging frequency-wavenumber power spectra extracted from seismic reflectors (Holbrook and Fer, 2005; Sheen et al., 2009; Fortin et al., 2016). The interplay between power slope spectra stands out as a pivotal methodology for bridging the gap between energy exchanges captured in seismic data and those documented in oceanographic measurements (Holbrook and Fer, 2005). These phenomena manifest as subtle oscillations along seismic reflections, characterized by sinusoidal
patterns spanning a wide range of wavelengths and amplitudes (Holbrook and Fer, 2005). Notably, seismic reflectors exhibiting lower wavenumbers align with the Garrett and Munk 75 (GM75) model for internal waves, adhering to a power law with a -2.5 exponent (Garrett and Munk, 1975). Conversely, higher wavenumbers are dominated by turbulence, characterized by the Kolmogorov -5/3 power exponent, closely matching subranges typically identified by oceanographic instruments. (Kolmogorov, 1941; Batchelor, 1959; Sheen et al., 2009). We focus on the turbulent subrange of seismic oceanography as it offers a more robust and accurate means of estimating mixing rates from horizontal measurements alone (Klymak and Moum, 2007; Sheen et al., 2009).

The application of spectral analysis in seismic oceanography operates on the premise that seismic reflectors align with isopycnal surfaces, where reflector displacements are defined by $\zeta$ (Holbrook and Fer, 2005). As a result, any deviations identified within the reflectors can be interpreted as perturbations in the isopycnals, provided that the data is not noise-contaminated (Holbrook et al., 2013). We estimate the power of vertical displacements from tracked reflectors, $\phi_{\zeta}$, across a range of wavenumbers $k_x$ using the work of Klymak and Moum (2007), where displacement from reflectors is calculated using multi-taper Fourier Transform, $|F(k_x)|^2$, to produce periodograms with less variability and bias than standard approaches (Thomson, 1982; Sheen et al., 2009).

Under the assumption that the rate of energy cascades from large to small scales, we can then quantify the rate of energy dissipation and subsequently diffusivity (Holbrook and Fer, 2005; Klymak and Moum, 2007; Sheen et al., 2009). Turbulent dissipation rate $\epsilon$ (W/kg) is calculated using a variation of the Batchelor model, by finding the best-fit model through least squares linear regression of the turbulent subrange (Klymak and Moum, 2007; Dickinson et al., 2017). The power of vertical displacement $\phi_{\zeta}$ is therefore defined as:

$$\phi_{\zeta} = \frac{4\pi \Gamma}{N_0^2} \left[ C_T \epsilon^{2/3} (2\pi k_x)^{-5/3} \right]$$  \hspace{1cm} (2.3)

where several variables are introduced, though most are assumed to be constant (Holbrook et al., 2013). $\Gamma$ is turbulent flux, generally, a constant value of 0.2, that relates the kinetic energy dissipation rate to an average buoyancy flux (Osborn, 1980; Falder et al., 2016).
$C_T$ is the Obukhov-Corrsin constant, equal to 0.4, used to describe how efficiently turbulent flow mixes a passive scalar like heat or salinity (Sreenivasan, 1996). $N_0$ denotes the mean buoyancy frequency, further explained below and used in dissipation and diffusivity calculations.

The concept of buoyancy frequency, also known as the Brunt-Väisälä frequency, delineates a vertical restoring force, describing the oscillatory motions triggered when lighter fluids enter denser regions, striving to reestablish equilibrium (Gargett, 1984). Buoyancy frequency $N$ is defined as

$$N = \sqrt{-\frac{g}{\rho} \frac{dp}{dz}}$$

(2.4)

where $g$ is the gravitational acceleration, and $\rho$ and $(dp/dz)$ represent the mean fluid density and its depth-related variation, respectively. $N$ is commonly calculated from CTD casts through the Gibbs-SeaWater (GSW) Oceanographic Toolbox which uses the Thermodynamics Equation of Seawater - 2010 (TEOS-10) approach (Wunsch, 1970; McDougall and Barker, 2011).

To emphasize the transition between the internal wave and the turbulent subrange, spectra are commonly multiplied by $(2\pi k_x)^2$, which converts vertical displacement power spectra into horizontal slope spectra, $\phi_{\zeta x}(k_x)$ (Klymak and Moum, 2007; Dickinson et al., 2017).

$$\phi_{\zeta x} = \frac{4\pi \Gamma}{N_0^2} \left[ C_T \epsilon^{2/3} (2\pi k_x)^{1/3} \right]$$

(2.5)

Following this transformation, the turbulent spectrum, when mapped from vertical displacement to slope spectra, now manifests a positive power exponent of $1/3$ instead of the previous $-5/3$ (Figure 2.2) (Klymak and Moum, 2007; Falder et al., 2016).

Once the best-fit model is calculated, I use $\epsilon$ to compute diapycnal diffusivity, $K_\rho$ defined as:

$$K_\rho = \frac{\Gamma \epsilon}{N^2 (m^2/s)}$$

(2.6)

Based on the energy level or where the slope falls along the displacement spectra, the level of diffusivity is calculated using either slope models (Figure 2.2) (Tang et al., 2021).
Figure 2.2: Figure adapted from (Tang et al., 2021) showing (a) tracked reflections and (b) mean horizontal spectrum of the tracked reflections where the x-axis shows horizontal wavenumber, $k_x$, and y-axis is the slope spectrum, $\phi_\zeta$, scaled by the normalized buoyancy frequency $N/N_0$. Slope -1/2 shows the fitted range for internal waves using the Garrett-Munk model Garrett and Munk (1975) and 1/3 slope is the best-fit line using the adjusted Batchelor model from Klymak and Moum (2007). The gray dashed lines (b) highlight the spectral levels for $K_p$ values. Solid vertical gray lines (b) denote the turbulent subrange used for diffusivity calculations.

Since we assume $\Gamma = 0.2$, and because the buoyancy frequency, $N$, generally has limited variability below the mixed layer, this approach provides a relatively accurate estimation of diffusivity, even with approximation of the use of constant values. (Osborn, 1980; Dickinson et al., 2017; Tang et al., 2021).
The seismic data used in this study were collected west to east over a region more than 800 km wide and during a time interval of more than a month (April 22 to May 25, 2004). These data cross the highly variable bathymetry of the South American continental slope and margin (Figure 1.2), and image the turbulent mixing of oceanic water masses merging and spilling into the Caribbean basin through the Lesser Antilles passages (Figure 1.3). I integrate seismic reflection data, in-situ XBT casts, and oceanographic models to map fine-scale density and velocity contrasts to identify regions of water mass mixing, quantify diffusivity across the region, and determine localized areas of increased mixing.

It is important to note that the ocean is a dynamic and rapidly changing environment. The seismic data recorded represent a "quasi-snapshot" of the ocean, capturing its structure at a specific moment in time. While the ocean's characteristics can vary temporally at a scale not fully understood, the data provide valuable information for understanding the general mixing behavior of the ocean. In the subsequent sections, I provide a detailed description of the data utilized to map and quantify oceanic mixing processes and features along each seismic profile.

3.1. Seismic Reflection Data

The seismic dataset used in this research was originally acquired in the southeast Caribbean, as part of the BOLIVAR (Broadband Ocean-Land Investigation of Venezuela and the Antilles Arc Region) focused on the Caribbean-South American plate boundary tectonics (Levander et al., 2006). For my study, I selected five seismic lines for a total length > 1800 km. The five profiles were selected based on their extent, location, and direction with respect to the water mass inflow into the Caribbean from the Atlantic (east to west). The five profiles
strike generally N-S, at a high angle relative to the western flowing water masses, and they are positioned at increasing distance from the Grenada Passage (Figure 1.2).

All data were acquired using the R/V Ewing, with a 114 l, 20-element airgun array, tuned to minimize the effects of source reverberations. Data were recorded on a 6 km-long digital streamer towed at a depth of 7 m. A shot interval of 50 m and a sampling rate of 4 ms resulted in a nominal fold of 60 and a frequency range up to 125 Hz (Table 3.1). The vertical resolution of the seismic data can resolve fine-scale mixing down to 10 m. The acquisition of seismic data posed challenges that affected the quality of the data, including source power-downs to minimize disturbance to marine life, which resulted in regions of low signal-to-noise data. Additionally, swells caused by storms led to increased wave movement, which caused tugging on the streamer and higher noise levels across a wide range of frequencies. This noise adversely affected the signal-to-noise ratio and our ability to accurately map coherent fine-scale boundaries in the intermediate ocean. Processing steps were taken to attenuate these and other noise sources and enhance coherent signal before and after stacking.

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<td>Fold</td>
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</tr>
</tbody>
</table>

Table 3.1: Acquisition parameters for the marine seismic data.
3.2. Oceanographic Data

3.2.1. In-Situ Oceanographic Measurements

One of the most accurate measurements of physical oceanic properties with depth often involves the use of expendable bathythermograph (XBT) and conductivity-temperature-depth (CTD) instruments. These instruments, deployed from vessels, sample the water column at discrete intervals. Compared to standard seismic reflection profiles, XBT/CTD instruments offer a vertical resolution of approximately 0.7 m, which is an order of magnitude higher. They provide temperature measurements with an accuracy of approximately $\pm 0.1^\circ C$ (Kizu and Hanawa, 2002).

XBT casts were acquired during the seismic acquisition campaign, coincident in space and time with the seismic reflection data (Figure 1.2). Out of all the XBT data, I identified 4 suitable casts to incorporate in the analysis (Figure 3.1). The casts comprise temperature-depth and velocity-depth pairs where depth values are derived from pressure measurements taken during the descent of the XBT probes. Salinity is assumed to have a constant value of 36 parts per thousand (ppt) for velocity estimates. According to Heinmiller et al. (1983), the standard margin of error in pressure-to-depth conversions is less than $\pm 2\%$. 
3.2.2. Oceanographic Model: HYCOM

Over the past few decades, numerous hydrographic measurements have been recorded in the Caribbean Sea. While these datasets provide insights into the general distribution of temperature and salinity in the region, they may not necessarily reflect the exact conditions at the time and location of the acquisition of the seismic data presented in this work.

In order to constrain my seismic interpretations with oceanographic data, I used the publicly available 3D assimilated HYCOM ocean model, which offers temperature, salinity, and water velocity at 1/12° gridded data resolution on a daily time scale (Figure 3.2). This model fills the information gaps where our in-situ measurements are not available.
HYCOM is defined as a Lagrangian Vertical Direction (LVD) model, which has the freedom to adjust the vertical spacing of the coordinate type (the three coordinate types used by ocean circulation models being depth, density, and terrain-following) to account for sophisticated physical processes of vertical and horizontal mixing (Chassignet et al., 2007). The model relies on isopycnal layers, resulting in higher resolution in highly mixed regions of the ocean typically present in the upper 1000m (Chassignet et al., 2009). Furthermore, the model removes any generated data that is more than three standard deviations from the mean standardized temperature or salinity (Thacker and Esenkov, 2002). As a result, the model has been proven successful in correcting for appropriate depth and thickness of model layers to accurately portray potential temperature, density, salinity and velocity within a useful resolution (Thacker and Esenkov, 2002).
Prior to seismic data interpretation, the data is processed both before summing individual ensembles (i.e. pre-stack), and after (i.e. post-stack), to produce coherent reflectivity, that ideally represents boundary layers in the ocean. In general, raw data are typically challenging to interpret due to the presence of noise, inadequate amplitude recovery, and the absence of migration, therefore processing techniques are employed to rectify these factors while minimizing undesirable artifacts and generating a sharper and more coherent image of the subsurface. Data processing is indispensable for all seismic exploration investigations, especially in oceanic environments where significant impedance contrasts at the seafloor tend to overshadow subtle density and velocity variations within the water column. The goal of the following chapter is to discuss the primary seismic processing steps used in this study and to highlight the impact of the processes employed on the final imaging. Additionally, I discuss minor data filtering applied to the XBT casts to remove long-wavelength temperature variations and allow a comparison of the XBT data with the fine-scale features observed in the seismic data.

4.1. Seismic Processing

4.1.1. Data Conditioning (Preprocessing)

The initial necessary step to process seismic reflection data is to build a database of the relative relationship between all sources and receivers throughout the survey (a process known as the geometry calculation). The data for each seismic line were downloaded from the Lamont Marine Data Center (http://marine-geo.org/) in multiple SEGY files, which were then concatenated according to the acquisition sequence. Receiver and source locations
were derived from the acquisition navigation metadata, used to calculate midpoint locations and to bin the dataset into Common Mid Point (CMP) bins. The geometry information generated was then loaded on the headers of the data, and data were displayed for QC.

4.1.2. Filtering

Since I am focused on the data within the water column, a key first step in processing the data is to remove all data outside of my area of interest, e.g. above the direct arrival and below the seafloor reflector. This has the advantage of calibrating amplitudes to enhance the faint reflectors in the water column, which are otherwise overpowered by the seafloor and subsurface reflectors. Upon first inspection it is evident that the data are affected by two main types of noise: 1) pervasive low-frequency ambient noise, and 2) source reverberation (Figure 4.1).

---

**Figure 4.1:** Unprocessed shot gather (SIN: 3000) from BOL3 profile before processing, highlighting the characteristics of signal and noise present in the raw data. The direct arrival, bubble pulse, faint reflections, and low-frequency, background noise can be identified. The inset shows the power spectra in dBPower from 0-125 Hz of the entire data window shown. Note the energy at 7 Hz, 19 Hz, and 32 Hz frequency (marked by the arrows) resulting from source noise reverberation (i.e., bubble pulse). Energy drops rapidly beyond 110 Hz.

Power spectra analysis of the data shows that the background noise is restricted to frequencies below 10 Hz (Figure 4.2), while signal decays rapidly beyond 110 Hz (Figure 4.3). I therefore applied a broad minimum phase Ormsby bandpass filter with 10-40-120-125
Hz corner frequencies, to remove the ambient noise and retain as broad a frequency band as possible.

Figure 4.2: (Left) Unprocessed shot gather (SIN: 3000) from BOL3 profile in absolute offset (m) vs two-way travel time (ms). (Right) Frequency spectra of selected portion of the data shown by the blue rectangle on (left) scaled by percent power, showing the dominant frequency of the noise.
Signal Frequency Spectra

Figure 4.3: (Left) Unprocessed shot gather (SIN: 3000) from BOL3 profile in absolute offset (m) vs two-way travel time (ms). (Right) Frequency spectra of selected portion of the data shown by the blue rectangle on (left), scaled by percent power. Data selection is narrow to avoid as much ambient noise as possible while falling within the hyperbolic energy region, which represents the reflected signal response of the water column. Frequency content in this region spans from background noise (i.e., $\sim 10$ Hz) to 110 Hz, with most of the energy between 40-80 Hz.

In order to retain the broadest frequency spectrum, particularly in the higher frequency ranges, I set the upper-frequency corner 5 Hz below the Nyquist (125 Hz). The frequency filter greatly improves the signal-to-noise ratio, by suppressing the background incoherent noise and highlighting reflections at depth (Figure 4.4). The coherent noise intertwined with the signal is addressed with additional tailored processing.
Figure 4.4: Shot gather (SIN: 3000) from BOL3 profile in absolute offset (m) vs time (ms), after the application of an Ormsby bandpass filter (10-40-120-125 Hz) and top and bottom mutes. Frequency filtering enhanced the presence of signal (i.e., hyperbolic arrivals between 600-1200 ms). Still visible are the direct arrival and the train of bubble pulse oscillations.

One of the most common types of noise in marine reflection seismology is source noise in the form of direct arrival and source reverberations, known as bubble pulse. The bubble pulse is the result of a sequence of expansions and contractions of the airgun bubble after its initial release. Each time the bubble expands, it generates a steep fronted shock wave in the water near the bubble (i.e., a pulse), that is recorded by the streamer (Figure 4.5). The bubble pulse oscillations typically lose energy over time, which creates a slightly variable pulsing wavelet. This noise is generally greatly reduced by using a tuned array, as during the acquisition of this dataset, but mild reverberations can be left over from the tuning process. In the dataset analyzed here, the bubble pulse dominates the first ~500 ms of data of all shots gathers (Figure 4.4), and needs to be attenuated/mitigated to extract shallower, near-offset reflectors in the water column and properly balance seismic amplitudes through the water column.
Figure 4.5: A general schematic of a bubble pulse along pressure-time. The pressure pulses result from the sequence of expansion and contraction of the gas bubble following an initial airgun blast. Each expansion generates a shock wave that follows the initial pulse, resulting in a train of reverberations (Lance et al., 2015).

Two filtering methods were used to remove this noise: predictive deconvolution and F-K (frequency-wavenumber) filtering. The first method exploits the repetitive, predictable nature of the bubble reverberations to attenuate the noise based on the wavelet characteristics and repeating time interval (Yilmaz, 2001). Since marine seismic data are minimum phase, I used a minimum phase predictive deconvolution filter to attenuate the bubble pulse noise. The parameters for deconvolution were established through autocorrelation of the first 0.5 seconds of data for each seismic shot (Figure 4.6). The best operator length (time interval between pulses) was determined to be 150 ms and the predictive lag (wavelet length) was 16 ms. Since the repeating pulse was not perfectly consistent in structure, predictive deconvolution was only partially successful at removing the source reverberation. I therefore relied on F-K filtering to remove the remaining noise.
F-K filtering uses a 2D Fourier transform to convert the data into the frequency-wavenumber (F-K) domains, which allows identification of events based on sound speed and/or event dip (i.e., cycle/distance) vs frequency. This type of filter is most effective with coherent noise, that can be easily isolated in the F-K domain, such as direct arrival energy and source reverberations, as well as additional coherent, ambient noise with small moveout (i.e., horizontally traveling energy) that is present in numerous shot gathers (Figure 4.7). This noise is particularly damaging to the final image as it can stack and disrupt the continuity of the hyperbolic structure of the reflectivity in the water column even in the Common Depth Point (CDP) domain and after NMO correction. Once the unwanted energy is identified, a filter is designed to remove the energy, and filtering is applied to all shots for each seismic line. The resulting processing successfully removes low-frequency noise, enhances signal at near and far offsets, and attenuates residual reverberations (Figure 4.8).
Figure 4.6: (a) Shot gather 3000 from BOL3 in absolute offset (m) vs time (ms). (b) Autocorrelation of the same shot showing 500ms of data, with the beginning of the direct arrival set at zero lag. The bubble pulse is present in both images and highlights the effect of the airgun reverberation on the data. The pulse appears every $\sim 160$ms in the upper 500ms of data.
Figure 4.7: FK analysis of SIN 6200 from BOL3 highlights the noise from the direct arrival and intermittent near zero wavenumbers. The correlating features are identified on the adjacent shot gather to the left. Manually drawn polygons are designed around the noisy regions to reject or attenuate the noise with a 5% taper on the edge of the polygon filter to reduce ringing. Amplitudes that appear above \( \sim 60 \) Hz in the positive wavenumber are aliased signals that wrapped around due to exceeding the Nyquist wavenumber (40 cpkm) as a result of trace spacing (12.5m).
4.1.3. Velocity Analysis and NMO Correction

Repeatedly imaging the same region from different source-receiver pairs creates redundancy, known as fold \( n_f \), defined as:

\[
  n_f = \left( \frac{n_g \Delta g}{2 \Delta s} \right) \quad (4.1)
\]

where \( \Delta g \) and \( n_g \) are group spacing and number of channels, respectively, and \( \Delta s \) is the shot interval.

CDP stacks are generated by summing traces that are imaging the same point in the subsurface (common depth points - CDPs) so that signal constructively interferes and uncorrelated noise attenuates, under the assumption that noise is not coherent. Because these traces are different source-receiver pairs with different source-receiver distances, the travel-time differences between the zero offset and all non-zero offset traces need to be corrected before the traces can be summed together to create a zero offset section and increase the signal-to-noise ratio (Yilmaz, 2001). A velocity field is, therefore, necessary to correct for
the hyperbolic normal moveout (NMO) of affected arrival times of CDP gathers. In seismic oceanographic studies, a constant velocity of 1500 m/s effectively corrects reflection times in the water column for stacking. Though, alternatively, velocity analysis is used to provide more robust NMO corrections (Fortin and Holbrook, 2009). For the purpose of this study, I relied on a constant velocity of 1500 m/s, since there was little effect observed in reflectivity between velocity picking and constant velocity. Following NMO correction, a CDP stack is generated by summing the mean trace response so that signal constructively interferes and attenuates uncorrelated noise (Yilmaz, 2001).

4.1.4. Migration and Depth Conversion

A seismic stack before migration does not represent the subsurface in its true subsurface position, so migration is applied to correct for dipping events and collapse diffractions that occur in complex geology and/or by abrupt truncations of reflective surfaces (Yilmaz, 2001). The goal of migration is then to make the stacked profile appear as similar to the true subsurface structure as possible. Though velocity picking is often employed to migrate seismic data, as long as lateral velocity variations are mild, I can use a constant velocity for migration. Since velocities in the ocean typically vary from 1470 m/s to 1530 m/s, I use a mean velocity of 1500 m/s to time migrate the seismic data (Tsuji et al., 2005).

Following time migration, data are converted to depth using the same constant velocity of 1500 m/s. One problem with depth conversion is that velocity values are generally estimated or limited in range, so depth conversion is not entirely accurate (Yilmaz, 2001). Though accuracy is a concern, it is still a widely employed process to interpret seismic profiles in depth since it most accurately represents the true subsurface positioning. Additionally, depth-converted sections make the correlation between different data types, such as XBTs and oceanographic models, possible.

4.2. Oceanographic Data

To compare the seismic reflection data with temperature data from the XBT casts I filtered the original temperature casts to remove longer wavelengths and reveal the finer
thermal structure of the water column at the time of seismic acquisition. In order to do so, I normalized the temperature casts by detrending them to zero and applied a high-pass filter to remove wavelengths greater than the dominant wavelength of the seismic data (Tsuji et al., 2005).

Wavelength is found using the equation:

\[ \lambda = \frac{v}{f} \]  

(4.2)

Where \( v \) is velocity and \( f \) is frequency. Since our data on average has a dominant frequency of 40 Hz and an average velocity of 1500 m/s, I filtered the casts in the coding language Python by applying a high-pass filter to pass waves between \((1/42 \text{ m})\) and the Nyquist wavenumber (Nandi et al., 2004). The filtered cast should then closely match the fine-structure wavelength resolution of the seismic data (Figure 4.9).
Figure 4.9: Four filtered XBT casts along BOL19 (Cast 1 and 2), BOL30 (Cast 3) and BOL12 (Cast 5) used in the analysis with the raw, unedited casts to the right. The filtered casts show distinct differences in variability, that relate to stratified layers in the ocean, with the exception of Cast 2 (located in the Cariaco Basin), which lacks relatively strong temperature variations below 250m.
I begin by identifying the presence of different water masses along each profile using water mass characteristics defined by Rhein et al. (2005) using temperature-salinity (T-S) plots from HYCOM (Figure 5.1). This allows us to determine the vertical extent of each water mass by considering their isopycnal density boundaries derived from HYCOM data (Tang et al., 2013).

Based on a modified version from Rhein et al. (2005) (Table 5.1), I identify three primary water masses along our seismic profiles: the Tropical Atlantic Central Water (TACW), the Antarctic Intermediate Water (AIW), and the North Atlantic Deep Water (NADW). Classification of the NADW relies on Morrison and Nowlin Jr. (1982), since most studies do not typically investigate water mass movement below the AIW in the Caribbean. The water masses are classified by isopycnal ranges according to individual water mass characteristics and approximate depths (Rhein et al., 2005).

Marine seismic reflection imaging typically does not contain resolvable data in the upper 200m, where Rhein et al. (2005) previously identified the Surface and Salinity Maximum water (SMW). I acknowledge this shallow water as an important mode of mass transport, but our study relies heavily on seismically observable regions to understand the nuances of internal mass movement within the thermocline, primarily the TACW and AIW. These two water mass zones have a mean South Atlantic Water (SAW) percentage of 56 - 59%, indicating a strong THC presence via the MOC through diffusion and advection (Figure 5.2) (Kirchner et al., 2008; Buffett et al., 2009).

The seismic profiles traverse the complex bathymetry of the Caribbean-South American plate boundary, which is characterized by deep basins (such as the Cariaco and Bonaire basins) separated by steep bathymetric highs of volcanic (e.g., Aves ridge, Leeward Antilles)
<table>
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<td>1000 - 2000</td>
<td>Relatively Homogeneous</td>
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Table 5.1: Water Masses Present in the Caribbean Sea With Density Distributions and Identifying Features (Adapted from Rhein et al. 2005)
or tectonic (e.g., the Southern Caribbean Deformed Belt) origin. These bathymetric features, along with factors like water velocities, eddies, and shallow bathymetry, contribute to the highly variable reflectivity observed in the seismic images.

It is important to note that the seismic lines were acquired at different times, with several days of difference between portions of the same profiles. When interpreting these profiles in terms of water column structures, it is necessary to consider the temporal scale of ocean circulation, which, in order to optimally stack the seismic reflections, must be on the same temporal scale as the seismic acquisition.

To constrain the seismic data interpretation, I identify water masses along seismic profiles based on HYCOM isopycnal depths and oceanographic parameters (temperature, salinity, and water flow velocity), and correlate them with reflection trends along each profile.

In the following sections, I describe the five seismic profiles from east to west with the related oceanographic data and identify the correlation between reflectivity and in-situ temperature casts.
Figure 5.1: Source water profiles used for the temperature/salinity (T/S) analysis for identifying water masses adapted from (Rhein et al., 2005; Kirchner et al., 2008). The northern source waters (18°N/55°W, 24°N/50°W), appear on the right side of the T/S diagram in red. The other source from the southern hemisphere (10°S/35°W, 3°S/23°W) is fresher than the northern source and is located on the left side of the T/S diagram. The blue profiles were extracted from their oceanographic model. The shorter green lines represent the salty South Atlantic source with green data points from the observed profile and yellow data points from their model. Salinity maximum water (SMW), Tropical Atlantic Central Water (TACW), Antarctic Intermediate Water (AIW), and North Atlantic Deep Water (NADW) are all labeled and identified based on the neutral density value range.
Figure 5.2: (a) General current flow patterns and major currents entering and circulating in the tropical Atlantic adapted from NOAA. Strong westward currents originating from the south and northern waters enter through the southern islands of the Caribbean. (b) The annual mean SAW distribution in central water (CW or ACW) and (c) the annual mean SAW distribution in Intermediate Water (IW or AIW) adapted from (Kirchner et al., 2009). Note the more dominant concentration of South Atlantic Water (SAW) in the southeastern region of the Caribbean Sea which decreases northwestern and is more concentrated in the intermediate water mass region. Yellow dashed box highlights the region shown in b and c.
5.1. Seismic Profiles

5.1.1. BOL 30

Previous studies (Johns et al., 2002; Schmitz Jr and Richardson, 1991; Wilson and Johns, 1997) have determined that the majority of water enters the Caribbean through the southeast of the Windward Antilles via the Grenada, St. Vincent, and St. Lucia Passages. This is supported by HYCOM surface water velocity in the region, which shows a strong northwestern flow during seismic data acquisition (Figure 5.3b).

BOL30 was acquired in a northwest-to-southeast direction over a continuous period of three days, starting on May 22nd and concluding on May 25th, 2004. The profile provides a unique opportunity to observe the acoustic response of the relatively undisturbed South Atlantic Current as it flows into the Caribbean and subsequently interacts with the North Atlantic Water (NAW) and upwelling Caribbean water influenced by surface currents and bathymetry (Correa-Ramirez et al., 2020; Lueck and Mudge, 1997; Wunsch, 2002). Temperature-salinity (T-S) plots, from two regions inboard (Figure 5.3c) and outboard (Figure 5.3d) of the Grenada passage, display a distinct difference of approximately 2°C in salinity minimum core in the AIW between the Caribbean and the Atlantic Ocean. This is likely due to the shallowing of fresh water, which is warmed by the near-surface or mixed layer as it spills over the Grenada Passage. The salinity minimum core, present in the T-S plot, and subsequent seismic profiles, is predominately formed in the SAW and weak or not present in the NAW (Kirchner et al., 2008). The fresher waters forming the salinity minimum core, originate from the South Atlantic Waters (SAW) and dominate the Antarctic Intermediate Waters (AIW) in the Atlantic Ocean and the southernmost part of the Caribbean (Figure 5.2) (Rhein et al., 2005; Kirchner et al., 2008).
Figure 5.3: (a) Location map of BOL30 profile showing the region of assimilated HYCOM data used for analysis of this profile. (b) Average eastward surface (<100 m) velocity (colors) overlain by northward velocity (contours) for the time of acquisition of BOL30 with profile location shown in red. (c) Temperature-Salinity (T-S) plot of the portion of the HYCOM data along the BOL30 profile west of the Grenada Passage (>11.5°N, >62°W), overlain by neutral density values for specific temperature, salinity, and depth. I identify the water masses based on neutral density boundaries from Rhein et al. (2005). (d) Temperature-Salinity (T-S) plot of the portion of HYCOM data along the BOL30 profile in the Atlantic Ocean (<11.5°N, <62°W), prior to entering the Caribbean, overlain by neutral density values for specific temperature, salinity, and depth. Waters masses labeled and separated by a solid black line for (c) and (d) are Salinity Maximum Water (SMW), Tropical Atlantic Central Water (TACW), Antarctic Intermediate Water (AIW), and North Atlantic Deep Water (NADW) for the respective neutral density range.

The complex interactions of nearly parallel surface currents, thermohaline mixing of Caribbean water and inflowing Atlantic water, and rugged bathymetry in this region, result in stark reflectivity and layering of both the internal Caribbean basin and the Atlantic water column (Figure 5.4). Beginning at the southeast end of BOL30 (CDP ~104000 - 84000), the
seismic reflection data in the Atlantic Ocean reveal a distinct pattern of short, undulating reflectors as the waters move to the northwest (Figure 5.4b and c). HYCOM velocities suggest that the NADW water flows south between CDP ∼104000 - 96000, opposite to the shallower water masses (Figure 5.4c). Notably, there are two vertically adjacent eddy-like features with strong, laterally continuous reflections that form lens structures near the edge of the rising seafloor (CDP 104000 - 96000), similar to features identified by Gorman et al. (2018). HYCOM velocity data indicates the presence of a northwestern current associated with the eddies, starting from the surface (Figure 5.4). Seismic imaging provides a valuable link between surface measurements and the subsurface by revealing the extent of the near-surface eddy extending into the deeper mixed layer. Furthermore, in the northern region (CDP 96000 - 88000), minor stratified layers extend toward the seafloor, but the area is dominated by scattered reflectivity (Figure 5.4b).

In the portion of the Grenada passage imaged by the seismic reflection data (CDP 56000 - 48000), the predominant inflow of water into the Caribbean is limited to the upper ∼750 meters, where the presence of relatively saline shallow waters is observed (Figure 5.4). Notably, slightly north, between Tobago and St. Vincent, a significant portion of intermediate waters containing the fresher South Atlantic Waters (SAW), enters the region (Rhein et al., 2005). Within the passage, the seismic data image a strongly layered water column, with continuous high amplitude reflectors bounding regions of semi-transparent reflectivity, particularly visible at a depth of ∼500m.

In the Grenada Basin (CDP 48000 - 20000), the seismic reflection data show high amplitude, laterally continuous (>30km-long), undulating reflectors that dominate the water column from 500m to 1km depth (Figure 5.4b). This depth range is primarily associated with the salinity minimum core and the isopycnal range of the AIW (Figure 5.3f), indicating that the laterally continuous reflectivity is correlated with the zone of minimum salinity and with the AIW. This correlation, in turn, supports the presence of a broader AIW in deeper waters (Figure 5.4e). A deeper AIW has been primarily attributed to the structural boundaries of the Grenada Passage and Aves Ridge, which influence thermocline depth (Wunsch,
2002). Additionally, Johns et al. (2002), identified this region has an increased influx of SAW through the Grenada and St. Vincent passages within the AIW, overlain by a higher concentration of NAW in the shallower waters (Johns et al., 2002). The two stratified waters likely encounter horizontal and vertical diapycnal diffusivity, leading to a variety of reflection characteristics. I observe shorter, lesser-defined reflectivity between 300-500m, where most energy is lost above 300m due to processing of the shallow noise. Below 500m, there is greater stratification, as the waters diffuse more slowly, allowing for the formation of longer, continuous, stacked reflectors, commonly associated with double diffusivity (Buffett et al., 2017) (Figure 5.4b).

Toward the NW, beyond this region, the continuous, subhorizontal, undulating reflectivity exhibits a sharp upward turn at the Aves Ridge (CDP 20000 - 16000), showing characteristics typical of lee waves, which are high amplitude, continuous, steeply dipping reflectors above shallowing seafloor as defined by Eakin et al. (2011) (Figure 5.4). Lee waves are observed in seismic profiles when deep, stratified water flows over steep topography, making the Aves Ridge an ideal location for such occurrence (Eakin et al., 2011). HYCOM water velocities support the influence of the shallow, rough bathymetry of the Aves Ridge, which is marked by a water velocity flows toward the northwest (Figure 5.4c) (Lueck and Mudge, 1997).

Northwest of Aves Ridge (CDP<16000) the seismic profile enters the Venezuelan Basin. Here the long, strong, reflectors that characterize the water column in the Grenada basin and spill over the Aves Ridge lose their continuity, and reflectivity is dominated by dense, short, highly stratified, randomly dipping reflectors (Figure 5.4b). Densely stratified reflectors may indicate the presence of intense double-diffusive mixing (Biescas et al., 2010), with the highest amplitude reflectors occurring between 500-800m. XBT data reveal temperature variations between 300-800m depth, that support the predominantly stratified region observed in the seismic profile from CDP 16000 - 0 (Figure 5.4a). A sharp decrease of approximately 1°C around 500m depth can be observed on the unfiltered cast, which correlates poorly with individual reflectors in the seismic profile (Figure 5.4a). The lack of correlation may suggest
other factors impacting reflectivity such as salinity. The HYCOM model further indicates the presence of deeper, surface and central waters that extend to approximately 500m (Figure 5.4). These higher salinity and warmer waters appear to be forced downward by the westward flowing surface current at -0.3 m/s before reaching the cooler, less saline AIW.
Figure 5.4: (a) The unfiltered in-situ temperature cast is shown to highlight the staircase-like gradients with the sharpest temperature changes marked in the box; (b) The seismic reflection profile BOL30 with interpreted features (see Figure 5.3 for location). HYCOM (c) water flow velocity and direction with positive contours indicating northward direction; (d) Temperature, and (e) Salinity along the seismic profile during acquisition.
Temperature and Salinity are used to show the relative concentrations for the oceanographic properties largely responsible for impedance contrasts. HYCOM predicted neutral density boundaries on (c), (d), and (e) delineate the general boundary of the subsequent water masses (TACW, AIW, and NADW). (f) Identified water masses (TACW, AIW, and NADW), based on HYCOM predicted isopycnal depths and reflectivity characteristics across each predicted water mass boundary (Buffett et al., 2009).

5.1.2. BOL 19

Profile BOL19 was acquired to the west of BOL30, following a predominantly north-south trajectory from May 13-14, 2004 (Figure 5.5). The data along this profile are strongly affected by weather-generated noise (i.e. swell noise), which impacts data from CDP 9700 through CDP 40000, just north of the Cariaco Basin. Additionally, poor weather conditions during data acquisition likely increased mixing, reducing the capability of extracting continuous reflectivity, but was improved through careful processing. This profile offers the opportunity to investigate the water column structure of the Cariaco Basin, one of the largest anoxic basins with true oceanic characteristics (Muller-Karger et al., 2001), and to compare it with the nearby Caribbean sea water structure. The Cariaco Basin is fully anoxic below a depth of 300 meters, primarily due to separation from large-scale Caribbean mixing by two shallow sills with depths of less than 150 meters. Consequently, the Cariaco Basin stands as one of the few isolated regions in the ocean predominantly driven by winds, specifically the Caribbean Low-Level Jet (CLLJ) (Jouanno and Sheinbaum, 2013; Alvera-Azcarate et al., 2011).

During data collection, HYCOM data show a distinct cold water filament at the surface (<100m water depth), originating from the Cariaco Basin in the south, which likely subducts toward the north (Figure 5.5b). T-S plots and in-situ XBT casts obtained from the relatively isolated Cariaco Basin (Figure 5.5d and 5.6b) show minimal temperature variation below 250m, suggesting a shallow thermocline. HYCOM data is sparse in Cariaco, though it still represents the traditional salinity maximum and linear decrease in temperature and salinity within the TACW but lacks the characteristic salinity minimum in the AIW (Figure 5.5d). The T-S plot for the Caribbean sea region north of the Cariaco Basin (Figure 5.5c) presents
a distinctive salinity minimum supporting the notion that stratified water masses exist in this area, a contrast to the conditions found in the Cariaco Basin.

Figure 5.5: (a) Location map of BOL19 showing the region of assimilated HYCOM data used for oceanographic data; (b) Average surface (<100 m) temperature for the time of acquisition of BOL19 showing locations of both BOL47 and BOL19; (c) T-S plot of the Caribbean Sea north of Cariaco (>11°N), overlain by neutral density values for specific temperature, salinity, and depth. I identify the water masses based on neutral density boundaries from Rhein et al. (2005); (d) T-S plot of the Cariaco Basin (<11°N), overlain by neutral density values for specific temperature, salinity, and depth. Waters masses labeled and separated by a solid black line for (c) and (d) Salinity Maximum Waters (SMW), Tropical Atlantic Central Water (TACW), Antarctic Intermediate Water (AIW), and North Atlantic Deep Water (NADW) for the respective neutral density range.

There is distinctive layering in the seismic profile, which is expected based on HYCOM (Figure 5.6c). The strong continuous layer between 300-400m is interpreted as the TACW, based on isopycnal predictive layers and the stark change in reflectivity observed along the
profile. The region from 400-1000m is interpreted using the same method but is associated with the AIW. Beyond this depth, the reflectivity fades and the dominant water mass is likely the NADW.

The profile intersects the Cariaco Basin, from the southern end of the line to approximately CDP 42000 (Figure 5.6c). Remarkably, the basin appears transparent with minimal recognizable reflectors reaching depths of only ∼350-400m. The seismic response in the basin diverges significantly from other seismo-acoustic images obtained. HYCOM also shows a drastically different density layering, likely due to the relatively constant, warmer waters below 400m, as supported by our Cast 2 data (Figure 3.1 and 5.6). The absence of reflectivity highlights the result of little to no internal ocean mixing, such as thermohaline mixing, and the related acoustic impedance response in the ocean.

Northward, outside of the Cariaco Basin, the seismic profile is highly variable. The zone between CDP 40000 and 30000, is characterized by short, undulating reflectors interspersed with subhorizontal, laterally continuous reflectors (Figure 5.6c). HYCOM water velocities indicate a 0.3 m/s eastward flow, underlain by 0.1 m/s westward flow (Figure 5.6d). Short, deepening, undulating reflectors, occur below the strong eastward flow, which induces stratification instability (Figure 5.6c and d) (Wenegrat et al., 2020).

Along the northern end, a semi-transparent region of approximately 25 kilometers in width (CDP 30000 - 26000) extends to the near-surface. This blanking region coincides with a significant change in eastward directional velocity and a near-zero northward component. Around CDP 23000, the amplitude of reflectivity increases (Figure 5.6c), and the presence of a strong southwestern current extends from the surface to a depth of 1500 meters (Figure 5.6d).

A transition in reflectivity is observed at CDP 8000 between 300m and 1km, where stacked, laterally coherent reflectors converge in the south and short, sloping reflectors dominate north of CDP 8000 (Figure 5.6c). The stacked reflectors exhibit characteristics similar to density-driven, diapycnal mixing, likely generated by the overlying cooler water and strong surface currents. These stacked reflectors remain identifiable to a depth of approx-
imately 1000 m. Beyond that region, isopycnals from HYCOM indicate the transition to non-reflectivity is the beginning of the North Atlantic Deep Water (NADW). Oceanographic models suggest that the location where the stacked reflectors change characteristics corresponds to a shift in water velocity directions, indicating the presence of a waterfront boundary. This boundary is similar to the apparent gyre observed on the surface temperature map between 12°N and 13°N (Figure 5.5b). Notably, a minor blanking zone around CDP 16000 between 600 meters and 750 meters is supported by a dominant eastward flowing water velocity and a lack of northward velocity (Figure 5.6).

Assuming that temperature is the primary factor, contributing to reflectivity, as shown by Sallarès et al. (2009), I would expect XBT cast 1 (Figure 5.6) to closely match changes in the observed reflectivity. However, this cast was collected at the northern end of the blanking zone (Figure 5.6a) and shows minimal correlation with the reflectivity in that area. This suggests that increased mixing in the region might lead to impedance contrasts that are too temporally short to correlate to the instantaneous cast data, which does not require internal layers to persist longer than a few seconds.
Figure 5.6: (a) XBT Cast 1 and (b) XBT Cast 2 (b) filtered to remove wavelengths longer than 42 m, plotted over a 19km-wide portion of the seismic reflection image. XBT Cast 2 in the Cariaco Basin displays minimal temperature changes, below the threshold of 0.04°C, at a depth greater than 300 m, correlating with a lack of reflectivity in the seismic reflection data; (c) Seismic profile of BOL19 with labeled water flow direction, indicating the presence of a local gyre boundary about 40km north of the Cariaco Basin; (d) HYCOM predicted eastward water velocity (color) overlain by northward velocity (contours) during seismic acquisition.
HYCOM predicted (e) temperature and (f) salinity spatial distribution along profile at the time of acquisition. HYCOM predicted neutral density boundaries on 5.5c, (e), and (f) delineate the boundaries of the water masses TACW, AIW, and NADW. (g) Identified water masses (TACW, AIW, and NADW), based on HYCOM-predicted isopycnal depths and reflectivity characteristics changes across each predicted water mass boundary.

5.1.3. BOL 47

Profile BOL47 was acquired south-to-north from May 12-13, 2004. The near-surface temperature (Figure 5.7c) averaged over the days of acquisition is roughly 0.5°C cooler during the BOL47 acquisition compared to BOL19. Particularly, north of 13°N, the surface water is cooler, and both HYCOM data and (Jouanno et al., 2008; Christianson, 2015) show that these waters sink below the more buoyant, warmer waters in the days following the seismic data acquisition.

Figure 5.7: (a) Location of BOL47 profile showing region of assimilated HYCOM data used for oceanographic analysis. (b) T-S plot of the region, overlain by neutral density values for a specific temperature, salinity, and depth. I identify the water masses based on neutral density boundaries from Rhein et al. (2005). (c) Average surface temperature for the time of acquisition of BOL47 showing locations of both BOL47 and BOL19. Waters masses labeled and separated by a solid black line for (b) Salinity Maximum Waters (SMW), Tropical Atlantic Central Waters (TACW), Antarctic Intermediate Waters (AIW), and North Atlantic Deep Waters (NADW) for the respective neutral density range

The BOL47 seismic profile exhibits prominent, long, continuous reflectors between 250-500m depth (Figure 5.8a). Based on HYCOM isopycnals (Figure 5.8), which identify the boundary between the AIW below and the TACW above at ~500m depth, I identify this
region as the TACW. The TACW/AIW isopycnal boundary correlates with a remarkable change in reflectivity, with the water column below the TACW being characterized by discontinuous, variably dipping reflectors. To the south, at about CDP 4000, a semi-transparent region extends from ~500m depth to the base of the thermocline (1000m) (Figure 5.8a). Profile BOL47 intersects a distinct cold water filament observed at the surface, extending from approximately CDP 1000 to 12000 (Figure 5.8). Similar filaments have been identified in previous studies, such as Andrade and Barton (2005) and Jouanno and Sheinbaum (2013).

Further north, between CDP 12000 and 20000, a less-defined semi-transparent region is observed below 750 meters. When comparing this location with eastward flowing velocities, it directly corresponds to a velocity of approximately 0.1 m/s, similar to the previously mentioned zone. However, in this case, the water is flowing in the eastward direction. Between these two zones, a change in eastward velocity occurs, but it does not result in a noticeable difference in seismic reflectivity. Here it appears that I capture reflection imaging on an intersecting, northeastern flowing filament, that appears to extend to the base of the thermocline (Figure 5.8). It is worth noting that the strong lateral reflectors in the upper 500 meters remain relatively consistent across the entire profile and seem to truncate any upward dipping reflectors. This indicates the presence of a well-defined and robust water mass at the surface.
Figure 5.8: (a) Seismic profile of BOL47 with labeled thermohaline staircases and (b) manually highlighted water masses (TACW, AIW, and NADW) represent the features and water masses expected across the profile. HYCOM generated (b) eastward velocity profiles overlain by northward velocity contours (c) Temperature, and (d) salinity. HYCOM predicted neutral density boundaries on (c), (d), and (e) delineate the general boundary of the subsequent water masses (TACW, AIW, and NADW). (e) Manually highlighted water masses (TACW, AIW, and NADW), based on HYCOM-predicted isopycnal depths and reflection characteristics that change roughly at the boundary of each predicted water mass.

5.1.4. BOL 12

Seismic profile BOL12 was acquired in two distinct continuous sections. The first section was collected from south to north from May 7th to May 8th, 2004, while the second section was collected from north to south between May 9th and May 10th, 2004. The original processing of this seismic line was performed by Christianson (2015), but it was reprocessed
with the same processing flow as the rest of the seismic data in this study for consistency. The resulting combined seismic image, shown in Figure 5.10, clearly displays two physiographic regions along the profile: the Bonaire Basin and the Venezuelan Basin. These two basins are separated by the Leeward Antilles, which extends a few hundred meters below the surface.

![Image](image.png)

Figure 5.9: (a) Location of BOL12 showing the region of assimilated HYCOM data used for oceanographic data. (b) T-S plot of the region, overlain by neutral density values for a specific temperature, salinity, and depth. Waters masses labeled and separated by solid black line for (b) Salinity Maximum Water (SMW), Tropical Atlantic Central Water (TACW), Antarctic Intermediate Water (AIW), and North Atlantic Deep Water (NADW) for the respective neutral density range.

The seismic profile exhibits clear layering, consistent with HYCOM predictions (see Figure 5.10c). Notably, a robust and continuous layer is observed to the 400m depth range, which aligns with the predicted position of the TACW based on isopycnal modeling. Further down, approximately between 400-1000m, a similar approach is used to interpret seismic features where reflectivity is highly variable, which is likely indicative of the AIW. At greater depths the reflectivity abates, suggesting that the predominant water mass in this region is likely the NADW.

South of the Leeward Antilles, in the Bonaire Basin, the seismic profile exhibits subhorizontal, gently southward dipping reflectors down to 800m, extending from the southern
edge of the Leeward Antilles high, at CDP 44000, to CDP 56000, where reflectors shorten and change dip direction (north dipping), above the shallowing bathymetry. The reflectivity within the basin appears to have a circular structure and suggests the presence of an eddy extending to ~750m depth and laterally over a distance of ~50km (Figure 5.10a). HYCOM data show a strong southeastward flowing surface current at the southernmost end of the profile. This strong current in the first 200m of water may induce Ekman pumping and upwelling of cooler, deeper water to the surface along the continental shelf (Figure 5.10b) (Christianson, 2015), resulting in the formation of an eddy feature in the basin.

North of the Leeward Antilles, two prominent, continuous, sigmoidal reflectors, likely filaments, extend from the northern edge of the Antilles Arc at the surface to a depth of 1km in the Venezuelan Basin, near the northern end of the profile (Figure 5.10a) (Christianson, 2015). Both reflectors/filaments sole out at about 1km depth, and mark the bottom of the reflective water column. Based on HYCOM isopycnal and temperature models (Figure 5.7b), this region of transparent reflectivity corresponds to the NADW (Figure 5.10). Temperature and salinity models reveal a shallower NADW, beginning in the Bonaire Basin, and moving toward the open waters around the same area as the southern filament. Though it has been found that cool water is upwelled within the Bonaire Basin and moves northward, it is more likely that the surface waters are subducting beyond the Leeward Antilles (Rueda-Roa and Muller-Karger, 2013; Christianson, 2015). As the filament subducts, it interacts with the strong eastward flowing waters, producing the lack of reflectivity between CDP 36000-28000.

A second eddy is interpreted on the northern section of the profile, beginning around CDP 16000 to the end. The eddy is marked by a semi-transparent lens-like structure (Gorman et al., 2018) and extends for 30km and occupies the full thermocline, being defined at the base by the strong, laterally continuous reflector that defines the northernmost filament. The eddy structure is visible in the HYCOM water velocity flow (Figure 5.10b) which shows a circular feature in the water velocity pattern.
Figure 5.10: (a) BOL12 Seismic profile with labeled features. HYCOM generated (b) eastward velocity profiles overlain by northward velocity contours (c) temperature, and (d) salinity. HYCOM predicted neutral density boundaries on (c) and (d) define the general boundaries of the water masses (TACW, AIW, and NADW) along the profile. (e) Water masses identified (TACW, AIW, and NADW), based on HYCOM-predicted isopycnal depths and reflectivity characteristic changes across each predicted water mass boundary.

5.1.5. BOL 3

Profile BOL3 was acquired during three non-sequential time intervals: April 29-30, followed by April 22, and then April 30 - May 1, 2004. During acquisition, the near-surface waters along the profile averaged \( \sim 0.3 \, \text{m/s} \) northwestern velocity (Figure 5.11c). Although the profile was acquired over \( \sim 10 \) days, the strong westward flowing current remained constant. Regardless, I separated the seismic observations into their discrete acquisition time, dividing them into three zones from south to north. Even so, many large-scale features
appear to persist across each temporal section (Figure 5.12a).

Figure 5.11: (a) Location of BOL3 showing the region of assimilated HYCOM data used for oceanographic analysis. (b) T-S plot of the region, overlain by neutral density values temperature, salinity, and depth characteristic of the water masses identified in the Caribbean Sea. I identify water masses based on neutral density boundaries from Rhein et al. (2005). Waters masses are labeled and separated by solid black line and are Salinity Maximum Water (SMW), Tropical Atlantic Central Water (TACW), Antarctic Intermediate Water (AIW), and North Atlantic Deep Water (NADW) for the respective neutral density range. (c) Average eastward surface velocity overlain by northward velocity contours averaged over the time of acquisition of BOL3 seismic profile.

The seismic profile has short, and mostly semi-transparent reflectivity within the HYCOM predicted TACW range. I interpret this region of seismic data as the TACW and indicative of more diffusive conditions (see Figure 5.12c). Beginning where more continuous reflectivity is observed around 400m, isopycnals support this region as the top of the AIW. Similar to previous interpretations, reflectivity diminishes, suggesting the transition to the NADW.

Zone 1 is characterized by densely stratified reflectors, extending from the farthest north to CDP ∼40000 and from a depth of 400m to 600m, exhibiting thinning and vertical undulations of continuous reflectors towards the south. Assuming the reflectors represent isopycnal layers (Figure 5.12a) (Biescas et al., 2010; Holbrook and Fer, 2005), they may be indicative of both turbulence and double-diffusivity. Within this region, a shallow layer (100-300m depth) with salinity levels above 36.8 PSU is observed, in combination with a high northwestern water velocity flow bounded at the base by a sharp vertical gradient (above 500m depth) (Fig-
ure 5.12b). The interaction between the warmer, more saline surface water and the cooler, fresher underlying water mass could result in the occurrence of salt fingering (Schmitt et al., 2005). As the water velocity gradient drops to the south, so does the continuity of reflectors.

Zone 2 is dominated by a region of laterally continuous and densely stacked reflectors, starting around CDP 20000, which tapers towards the north, accompanied by a semi-transparent region in the upper 600m. The upper boundary, marked by a dashed line in Figure 5.12a, is a change in reflector continuity, angle, and stacked appearance. This boundary aligns with a region of relatively higher westward and northward-flowing water, exhibiting fairly homogeneous salinity and temperature. Reflectivity patterns suggest a water mass boundary and the potential presence of a diffusive thermohaline staircase, similar to the characteristics reported by Tang and Zheng (2011), where blanking zones indicate the boundaries of adjacent water masses. The waterfront boundary does not correspond to an abrupt change in temperature or salinity but is strongly supported by water velocities that extend from the near-surface into the deeper ocean, gradually weakening towards the north.

Zone 3, though less resolvable given the narrow space, exhibits a similar structure to Zone 2, with a stratified region of reflectors tapering towards the south near the waterfront boundary, accompanied by a strong northwestern velocity and a high salinity region that also tapers in a similar manner (Figure 5.12b and d). Along the shallowing coast (CDP 16000-8000), water velocity and the swallowing bathymetry, indicate a flow pattern that likely introduces the more saline near-surface water to the cooler, less saline TACW, resulting in a double-diffusive process within the AIW, giving rise to diapycnal mixing.
Figure 5.12: (a) BOL3 seismic profile with labeled oceanographic interpreted features. HYCOM generated (b) eastward velocity (color) overlain by northward velocity (contours), (c) temperature, and (d) salinity predicted model at the time of acquisition. HYCOM predicted neutral density boundaries on (c), and (d) delineate the general boundary of the TACW, AIW, and NADW. (e) Water masses (TACW, AIW, and NADW) identified based on HYCOM predicted isopycnal depths and reflectivity characteristic changes across each predicted water mass boundary.

The presence of double-diffusive structures in the seismic profile exhibits a strong association with higher salinity near the surface. The characteristics of zone 2, on the other hand, are primarily correlated with changes in water velocity. While the temperature gradient in the oceanographic models remains relatively consistent across the profile, the salinity maximum exhibits more variability among the three zones, with the broadest salinity maximum
zone observed in the northernmost 100km of the profile.

5.2. XBT Casts

Five in-situ Expendable Bathythermograph (XBT) casts were acquired during the seismic survey (Figure 1.2). In general, the temperature is responsible for approximately 80% of seismic impedance, therefore I concentrated on temperature as the leading factor impacting seismic reflectivity (Sallarès et al., 2009). After filtering (see Chapter 5), four XBT profiles were correlated with seismic reflection data, with the exception of Cast 4, due to an approximate offset of 40 km from the BOL30 transect.

XBT cast 5 (Figure 5.13a and b) exhibits a significant correlation with thermohaline staircases, observed in the seismic data between 300m and 700m depth, above the depth of the interpreted filament, which is located at ∼700m at the location of the cast (Christianson, 2015) (Figure 5.13a). Although our XBT measurements are limited to sparse in-situ data, where available, particularly for Cast 5, they display a distinct "step-like" structure, as described by Tait and Howe (1971) (Figure 5.13b). The temperature cast reveals a pronounced temperature change near the base of the filament at around 700m depth, characterized by a spike in colder temperature overlaying warmer water. The presence of high amplitude reflectors is attributed to temperature contrasts of approximately 0.1°C. Notably, temperature contrasts below 0.03°C, below 900 m, do not seem to produce reflectivity, similar to the findings of Nandi et al. (2004). Temperature casts 1 and 3 indicated an average fluctuation of 0.08°C within the thermocline (<1000m) (Figure 4.9). However, these casts were more difficult to correlate with specific reflectors and showed a better association with the general region of the thermocline.

In-situ XBT cast 2 (Figure 4.9b), collected in the Cariaco Basin, shows temperature fluctuations of 1 magnitude less than the other casts within the typical central and intermediate water regions. The variability drops below the impedance contrast threshold of 0.03°C at ∼250m, which is significantly shallower than the other regions (Figure 5.13c). I estimate the base of thermocline at ∼250m during acquisition, which is ∼30m deeper than Alvera-
Azcarate et al. (2011) estimates for the Cariaco in the same month and year (May 2004). Both the temperature cast and seismic reflection data corroborate this lower boundary of the oxic-anoxic boundary. However, it is important to recognize that the variability of the mixed layer depth in the Cariaco Basin is primarily driven by winds and eddies in the Caribbean (Alvera-Azcarate et al., 2011).

Most importantly, the key finding from the cast analysis and seismic, is that in some cases, amplitudes present a robust correlation to temperature, (Figure 5.13). Furthermore, our analysis shows that the primary region of stratified mixing, with the exception of the Cariaco Basin, is confined predominantly within the upper 1 km of the water column.
Figure 5.13: (a) Filtered XBT Cast 5 plotted over a 19km-wide seismic section of BOL12. Temperature variations greater than 0.1°C correlate strongly with the high amplitude reflectors shallower than the Antarctic Intermediate Waters (AIW). The zone associated with the AIW is where temperature fluctuations of ~0.04 - 0.1°C occur, which is intersected by the reflections corresponding to the filament identified along the seismic profile. The North Atlantic Deep Waters, denoted NADW, show decreased reflectivity where temperature variations approach 0. (b) Unfiltered XBT profile with zoomed-in region focused on the area of interpreted staircase features, highlighting the step-like structures present in the area. Interface and layer thickness identify the temperature response of thermocline boundaries and layers; (c) Filtered XBT Cast 2 plotted over a 19km-wide seismic section of BOL19, in the Cariaco Basin. The cast is displayed using a different temperature scale to emphasize the dense, subtle, temperature changes. Reflectivity is only present in the upper 250 meters where temperature undulations are greater than 0.04°C.
6.1. Pre-Conditioning and Spectra Analysis

Prior to performing diffusivity calculations, steps need to be taken to determine the suitability of the seismic data for spectra analysis by verifying that the seismic data contains useful horizontal wave number, $k_x$ content within the turbulent slope range (Holbrook et al., 2013). In this section, we outline the preconditioning steps using data from the BOL30 profile as an example, although we apply these steps to data of all seismic profiles in this study.

First, direct data transforms are used to extract amplitude and frequency content of a selected data window and to display the average response on a power-wavenumber space to identify whether the seismic data contains the turbulent slope of $k_x^{1/3}$, used for turbulent diffusivity calculation (Figure 6.1) (Klymak and Moum, 2007; Falder et al., 2016).
Figure 6.1: (Left) Sample seismic data window from profile BOL30; (right) direct data transform of data shown left plotted as a function of amplitude and wavenumber ($k_x$). Internal wave field, turbulence, and noise are identified on the spectra using the best-fit slope for the power-slope of $-1/2$, $1/3$, and 2, respectively. The turbulent subrange is identifiable from the data transform, indicating that diffusivity calculations are possible on these data (Holbrook et al., 2013).

Once the presence of the turbulent slope is verified within the data, we evaluate the frequency content of the data and calculate the signal-to-noise ratio (SNR). The SNR for a given data window is assessed using the mean SNR value, with a target minimum SNR of 4, to determine the quality of the seismic data (Holbrook et al., 2013; Falder et al., 2016). SNR is evaluated by calculating the maximum cross-correlation value ($c$) between traces and the difference between the zero-lag autocorrelation of the first trace ($a$) from the direct data transform as (Eq. 6.1) (Holbrook et al., 2013; Dickinson et al., 2017).

$$ SNR = \sqrt{\frac{|c|}{|a - c|}}, \quad (6.1) $$

Provided the signal-to-noise ratio (SNR) is sufficient, we auto-track reflectors using the instantaneous phase angle, $\phi$, determined from the Hilbert transform of the seismic profile (Barnes, 2007; Holbrook et al., 2013). To accomplish this, we begin by normalizing the amplitude of each reflection to the range of -1 to +1 by calculating the cosine of the instantaneous phase angle, which enhances the continuity of reflections and eliminates the phase
angle discontinuity at \( \pi/2 \) (Figure 6.2) (Barnes, 2007; Holbrook et al., 2013). Instantaneous phase values are contoured with a constant value to discern individual continuous reflections. The chosen contour value does not alter the geometry but does influence the count of tracked reflectors, potentially leading to disparities in more noise-prone areas (Gunn et al., 2021). Holbrook et al., 2013 recommend a contour value of 0.6 to preserve an adequate SNR while still capturing a satisfactory number of reflectors for subsequent analysis. However, we conducted tests with various contour values and determined that \( \pm 0.8 \) yields the maximum number of reflectors while preserving accurate tracking, similar to findings by Gunn et al. (2021).

Subsequently, the contoured regions are subdivided into segments corresponding to continuous reflectors, with each segment containing a minimum of \( n \) data points (we set \( n=200 \)). This configuration ensures a minimum tracking distance of 1.25 km, considering a common depth point (CDP) spacing of 6.25 m, allowing us to observe turbulent features (Figure 6.2) (Holbrook et al., 2013; Falder et al., 2016; Gunn et al., 2021). From the tracked reflectors, we derive displacement spectra, denoted as \( \phi \), which are then converted into slope spectra by multiplication with \((2\pi k_x)^2\), enabling subsequent analysis (Klymak and Moum, 2007; Sheen et al., 2009).

This methodology of utilizing reflector tracking based on instantaneous phase has shown notable benefits. It yields precise reflector tracks that are more reliable compared to tracking reflection amplitudes and less time-consuming than manual selections guided by human intervention (Holbrook et al., 2013; Dickinson et al., 2017).

It is important to note that under the assumption of sufficient SNR in the seismic data, the tracked reflectors are presumed to align with isopycnal surfaces. This assumption implies that any observed displacements in the reflectors can be interpreted as displacements of isopycnals. While permanent fine structures, which may appear as seismic reflections, can be attributed to phenomena such as intrusions or thermohaline staircases, their slopes with respect to density surfaces are relatively small (Holbrook and Fer, 2005; Sheen et al., 2009). The displacement of these relatively stable density gradients offers valuable insights into the
spatial characteristics of these semi-permanent thermohaline features (Holbrook and Fer, 2005).
Figure 6.2: (Top) Portion of BOL30 seismic profile used for diffusivity calculations displayed with no amplitude correction; (Middle) Same data window as in top panel, with normalized amplitudes and reflections converted to $\cos(\phi)$ for clearer continuity that is not amplitude dependent (Barnes, 2007); (Bottom) Same data window as in top panel showing auto-tracked reflections. Reflections of a minimum length of 1km are tracked, as they are more dependent on SNR than the original amplitude of reflectivity.
In addition to these preconditioning steps, actions were taken to improve the signal-to-noise ratio (SNR) of the seismic profiles in order to enhance reflection continuity. In particular, we implemented two distinct processing steps, as proposed by Holbrook et al. (2013). The first step entailed a subtle refinement to the original bandpass filter parameters (10-40-110-125 Hz), where the adjustment involved modifying the bandpass filter to span 30-100 Hz, accompanied by a 15 Hz taper on both ends. This adjustment led to a marginal enhancement in the visibility of internal wave features within the spectra, simultaneously mitigating the influence of high-frequency noise \( (k_x > 0.02 \text{ cpm}) \) beyond the turbulent subrange.

The second processing step, of greater significance, utilized notch filters at intervals of \( k_x = 0.02 \text{ cpm} \) to counteract harmonic noise resulting from a 50m shot spacing (Holbrook et al., 2013). Harmonic noise was pervasive across each seismic line and negatively affected reflection tracking if left unaddressed. Within the original seismic profile, the recurrent spikes at 0.02 cpm disrupted the continuity of reflection tracking and reduced the median tracking length by nearly 15% (Figure 6.3). We produced a band-stop filter centered over each notch with a width of \( \pm 0.002 k_x \), effectively attenuating the detrimental spike from \( \pm 0.02, \pm 0.04 \) and \( \pm 0.06 \text{ cpm} \) (Figure 6.4). Post-application of the filtering process, the median length of tracked reflectors increased from 2.25 km to 2.6 km, accompanied by an increase of total length tracked of \( \sim 50 \text{ km} \) (Figure 6.4).
Figure 6.3: (Top) Horizontal wavenumber, $k_x$ (cpm) spectrum prior to processing steps, produced by summing the amplitudes from the 2D Fourier Transform spectrum for the BOL30 seismic data window (Bottom Left); (Bottom Right) Averaged slope spectrum, $\phi_{\zeta x}$, of the data window, highlighting the best-fit turbulent range. Noticeable spikes are present in both spectrum images at 0.02 cpm.
Figure 6.4: (Top) Horizontal wavenumber, $k_x$ (cpm) spectrum after the application of processing steps, produced by summing the amplitudes from the 2D Fourier Transform spectrum for the BOL30 seismic data window (Bottom Left); (Bottom Right) Averaged slope spectrum, $\phi_{\zeta x}$, of the entire window, highlighting the best-fit turbulent range. Note the improved coherency in the reflectors after the preprocessing steps (Tang et al., 2021).

Spectral characteristics of the turbulent subrange from both data transform and slope spectrum indicate a general range of 0.004-0.0155 ($10^{-2.4}$-$10^{-1.8}$) cpm for the data analyzed (Figure 6.5). This subrange is consistent for both extraction methods throughout the profiles, given a sufficient SNR. This result ensures confidence in implementing the processing methods from Holbrook et al., 2013, and provides assurance that we can extract meaningful information from reflection tracking.
Figure 6.5: (A) BOL30 seismic profile showing the sample region where the direct data transform (DDT) was extracted (black box); (B) portion of the seismic profile and (C) DDT spectrum, shown using arbitrary units for amplitude. A clear change in slope occurs at 0.004 cpm and typically ends around 0.02 cpm, where the data becomes dominated by noise.
The slope spectra are calculated directly from detrended (D) tracked seismic lines (black lines), on the same portion of the seismic profile used for the DDT extraction. The shaded region highlights the standard deviation range calculated from 2 degrees of freedom. The turbulent subrange is identified as a best-fit line (red dashed line) on both, DDT and Displacement Spectra, showing a similar subrange.

To determine turbulence from tracked reflectors, we rely on the equation for the slope spectrum of the turbulent regime $\phi_\zeta$, for seismic data (see Chapter 2), defined as:

$$\phi_\zeta x = \frac{4\pi \Gamma}{N_0^2} \left[ C_T \epsilon^{2/3} (2\pi k_x)^{1/3} \right]$$

(6.2)

Where, $\Gamma$ is turbulent dissipation flux, $\epsilon$ is the turbulent dissipation rate, $N_0$ is the mean buoyancy frequency, and $C_T$ is the Obukhov-Corrsin constant (Klymak and Moum, 2007; Osborn and Cox, 1972; Sreenivasan, 1996; Osborn, 1980). Where the reflector slope spectra, $\phi_\zeta x (k_x)$, yield a 1/3 slope in $\phi_\zeta x - k_x$ space, turbulent dissipation, $\epsilon$, and subsequently diapycnal diffusivity, $k_\rho$ is calculated by aligning the mean reflector slope spectra with the Klymak and Moum (2007) subrange model, $\phi_\zeta x$, to estimate the $\epsilon$ (Figure 6.6). We rely on Dickinson et al. (2017) for best multi-taper parameters, where wavenumber resolution ($J_{\text{res}}$) is 4 and power resolution ($K_{\text{res}}$) is 6. We then apply the process to all tracked reflectors for each seismic line, where SNR is greater than 4 (Figures 6.8, 7.2- 7.5). As global open ocean mean is defined as logkt=-5 $m^2/s$, we convert values to logarithm to compare our results with global ocean values (Figure 6.6) (Munk and Wunsch, 1998).

In the diffusivity equation, $k_\rho$,

$$K_\rho = \Gamma \epsilon / N^2$$

(6.3)

we assume a constant dissipation flux, $\Gamma=0.2$, conventionally used in seismic oceanographic studies and described by Osborn (1980). Additionally, we calculate a buoyancy frequency, $N$, which is assumed to be constant ($N=1.8$), to remove localized dependencies (Sheen et al., 2009). We used a constant $N$ as no CTD casts were collected during the seismic acquisition to constrain buoyancy frequency variations during the seismic survey. The average buoyancy value was calculated following TEOS-10 guidelines, using 17 CTD casts extracted from the World Ocean Circulation Experiment (WOCE) database, collected in August 1997 and
October 2003, along 66°W, between 11-14°N (McDougall and Barker, 2011). Since our primary region of reflectivity falls within the Antarctic Intermediate Water (AIW), the average buoyancy frequency was calculated from depths 300m-1000m, commonly associated with the depth of the AIW. At depths > 300m, $N$ varies ±0.3, while the greatest variability is in the upper 200m (Figure 6.7).
Figure 6.6: (Top) Portion of BOL30 seismic profile with auto-tracked reflectors shown in black. Red labeled reflectors are selected for spectra analysis shown on the bottom plots; (Bottom) The three plots show the resulting individual reflector spectra along $\log \phi_x - \log k_x$ space to better represent the displacement spectra extent. The region highlighted in red identifies the turbulent subrange. The best-fit model is shown as a green dashed line. The calculated diffusivity is shown in log form as $\log k_t$ for turbulent diffusivity. Slope spectra calculations were performed on every tracked reflector for all seismic profiles in this study.
Figure 6.7: Buoyancy frequency calculations for 17 CTD casts from the World Ocean Circulation Experiment (WOCE) database collected during April 1997-October 2003 along 66°W. Left side shows each individual cast calculation, indicating a significantly higher $N$ in the upper 200m. The right plot shows the average of calculation from the 17 casts with the shaded region indicating one standard deviation from the mean estimation at 100m intervals. The red dashed line shows the value of $N=1.8$ used in this study.

### 6.2. Diffusivity Results

We applied spectral analysis by tracking reflectors on all five seismic lines, to study the turbulent mixing and lateral continuity of isopycnals below the mixed layer down to approximately the top of the NADW (1.2 km depth). Smoothed diffusivity plots were created to enhance observations of spatial variability by averaging half-overlapping windows of 2km width and 20m high (Figure 6.8) (Tang et al., 2021). The smoothed maps show greater macro trends of mixing along the seismic lines.

In this section, we highlight results from BOL30. Results from diffusivity analysis of all other seismic lines (BOL19, BOL47, BOL12, and BOL3) are discussed in the following chapter. For all seismic lines, analysis begins below the mixed layer (>300m) due to poor resolution near the airgun source noise. Along BOL30, the area of interest encompasses the Grenada Passage through to the Venezuelan Basin, a ∼280km-long region extending from 0.3-1.2 km depth, and 12000-57000 CDP. Along BOL30, we auto-tracked 2703 reflections for
a total distance of 6365.52 km, with a median length of 1.88 km per tracked reflector (Figure 6.8). The average SNR was 6 for the analysis, well above the necessary threshold (Holbrook et al., 2013).

We capture isopycnal layers of water masses as they enter the Grenada Passage and spill into the Grenada and Venezuelan Basin (Figure 6.9). BOL30 is the only seismic line acquired parallel to the current inflow, and captures multiple regions likely impacted by topographic effective mixing (Lueck and Mudge, 1997). The average diffusivity for the entire region is $\log_{10}(K_\rho)=-5.26 \text{ m}^2/\text{s}$, similar to $-5.3 \text{ m}^2/\text{s}$ by Kunze et al. (2006), which is still below the global average of open ocean diffusivity, $-5 \text{ m}^2/\text{s}$ (Munk and Wunsch, 1998). Even so, diffusivity levels were spatially variable and ranged by approximately 6 orders of magnitude (Figure 6.10).

Key areas of increased diffusivity ($K_\rho>-4 \text{ m}^2/\text{s}$) are visible near the Grenada Passage (CDP 40000-55000) and as the waters flow over the Aves Ridge (CDP 15000-20000) (Figure 6.8). Both of these areas correlate to shallower bathymetry, which disrupts the general current flow direction and produces greater vertical forces (Lueck and Mudge, 1997). Reflections leading up to the spillway in the Grenada Passage (CDP 50000-55000) tend to be longer, and have lower diffusivity, before shortening and mixing increases near the slope. Above the Aves Ridge, a sharp change in reflector angle occurs, and a subsequent increase in diffusivity vertically extends from the seafloor to the mixed layer (Figure 6.8).

The deeper waters ($>1\text{km}$) in the Grenada Basin (CDP 25000-35000), have longer and more stratified reflectors of lower diffusivity ($K_\rho<-6 \text{ m}^2/\text{s}$). Where the waters are more stratified, diffusivity tends to decrease, and regions of increased mixing tend to have shorter, discontinuous reflectors (Holbrook et al., 2013; Tang et al., 2021). Throughout the profile, increased diffusivity patches are spatially variable. However, diffusivity appears to increase around bathymetric highs and steep slopes, where mixing is favored.
Figure 6.8: The full BOL30 seismic profile with black box identifying the area of interest for subsequently tracked reflectors and resulting diffusivity calculations. Estimated turbulent diffusivity, labeled Logkt, was calculated for each individually tracked reflector. The reflector results were blended using a 2km by 20m, half overlapping window, to create a better qualitative understanding of larger regions of observation.
Figure 6.9: (a) 3D vertically exaggerated map of the study area highlighting bathymetric variability across the region. Key structural features are labeled for reference. Individual seismic lines and casts are labeled on the (b) 2D map for correlation with the structural features. (c) Average absolute current velocity for the upper 100m of the ocean during the seismic acquisition showing the general flow direction and magnitude of the shallow water mass. Note that seismic profile BOL30, highlighted in black, is the only seismic profile roughly parallel to the general water flow direction. All other profiles strike at high angle to the water current.
The application of the automated tracking algorithm to each seismic profile yielded an average of approximately 2078.2 tracked reflectors, with an average cumulative length of 5139.93 km. The median length of tracked reflectors was 1.92 km, and the mean signal-to-noise ratio (SNR) was 5.6 (Table 6.1). I calculated diffusivity values for a total of 10,391 reflectors spanning five seismic profiles. The range of diffusivity values extended over six orders of magnitude, from approximately \(-8\) to \(-2\) m\(^2\)/s, which is generally consistent with findings from other seismic oceanographic studies (Dickinson et al., 2017; Gunn et al., 2021). The average diffusivity across all profiles was \(-5.2\) m\(^2\)/s, resembling the average calculated by Kunze et al. (Kunze et al., 2006) (\(\log_{10}(K_\rho) = -5.3\) m\(^2\)/s) for the internal ocean along 66°W. Histogram distributions of diffusivity values for each profile are presented in Appendix A.

It is essential to acknowledge that the use of fixed constants for parameters such as \(\Gamma\), \(N\), and \(C_T\) introduces simplifications into the calculations, leading to an inherent level of uncertainty. Following the approach of Dickinson et al. (2017), the maximum likely uncertainty in \(\log_{10}(K_\rho)\) is ±0.4 m\(^2\)/s in log units. This study uncertainty estimate for \(N\) is ±0.3 cph, contributing to an average uncertainty estimate of ±0.15 m\(^2\)/s in log units.
(Dickinson et al., 2017; Gunn et al., 2021). Taking into account typical upper and lower bounds for $\Gamma$ (0.1, 0.4) and $C_T$ (0.3, 0.5), the overall uncertainty estimate is approximately $\pm 0.25$ (Dickinson et al., 2020).

<table>
<thead>
<tr>
<th>Profile</th>
<th>Log$K_\rho$ (mean)</th>
<th>Number of Reflections Tracked</th>
<th>Cumulative Length of Tracked Reflections (km)</th>
<th>Median Length (km)</th>
<th>SNR (mean)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BOL30</td>
<td>-5.26</td>
<td>2703</td>
<td>6365.52</td>
<td>1.88</td>
<td>6.0</td>
</tr>
<tr>
<td>BOL19</td>
<td>-5.28</td>
<td>1350</td>
<td>3262.39</td>
<td>1.91</td>
<td>4.0</td>
</tr>
<tr>
<td>BOL47</td>
<td>-4.71</td>
<td>1918</td>
<td>4390.43</td>
<td>1.86</td>
<td>7.0</td>
</tr>
<tr>
<td>BOL12</td>
<td>-5.23</td>
<td>2503</td>
<td>6274.21</td>
<td>1.93</td>
<td>4.0</td>
</tr>
<tr>
<td>BOL3</td>
<td>-5.50</td>
<td>1917</td>
<td>5407.11</td>
<td>2.04</td>
<td>7.0</td>
</tr>
<tr>
<td>Average</td>
<td>-5.20</td>
<td>2078</td>
<td>5139.93</td>
<td>1.92</td>
<td>5.6</td>
</tr>
</tbody>
</table>

Table 6.1: Table showing number of reflections tracked, cumulative length of tracked reflections, and median length for each dataset.
Chapter 7

DISCUSSION

The analysis conducted by this study highlights the presence of mixing features predominantly within the Antarctic Intermediate Water (AIW), as it flows from east to west and interacts both with other water masses and with the bathymetric structures in the region. The distinctly stratified water masses are identified by their unique temperature and salinity traits within specific isopycnal boundaries (Morrison and Nowlin Jr, 1982; Rhein et al., 2005).

The water masses entering the region are primarily sourced from the South Atlantic Water (SAW), accounting for about 65% of the total (Rhein et al., 2005; Kirchner et al., 2008). The fresher SAW flows in a northwesterly direction, with the predominant influx pathway through Grenada and St. Vincent of the Lesser Antilles (Morrison and Nowlin Jr, 1982; Rhein et al., 2005). The transportation of SAW into the Caribbean mainly occurs along a boundary current along the South American Coast and by the retroflective North Brazil Current (NBC) (Johns et al., 2002; Rhein et al., 2005). Conversely, the North Atlantic Water (NAW) is primarily propelled by the westward flowing North Equatorial Current, under the influence of the Caribbean Low-Level Jet (CLLJ) (Schott et al., 1998).

Water inflow exhibits seasonal variability, and during seismic acquisition, the influx of waters was closer to the salinity maximum period observed in March from along-shore transport by the NBC (Johns et al., 2002; Chérubin and Richardson, 2007). Despite the increased salinity from the SAW, the Temperature-Salinity (T-S) plots along all five seismic lines indicate a discernible salinity minimum present in the AIW (Morrison and Nowlin Jr, 1982).

Turbulent phenomena within oceanic environments exert a profound influence on various processes and phenomena. An intriguing observation involves the diffusivity patterns, particularly in the proximity of rough topographical structures, which can exhibit values
substantially higher, potentially two orders of magnitude or greater, than those found in the open ocean (Munk and Wunsch, 1998; Tang et al., 2022). This phenomenon has piqued our interest in investigating variations in diffusivity near bathymetric features, offering insights into augmented mixing processes.

Our more focused investigation centers on regions of heightened diffusivity adjacent to shallow topography, commonly referred to as ”hot-spots” of intense mixing (Tang et al., 2022). The primary study area encompasses the BOL30 region, where we aim to elucidate the variability in diffusivity near sharp, bathymetric changes, specifically, in the Grenada Basin and above the Aves Ridge (Figure 6.9).

Generally, we observe increased diffusivity the seafloor, with more pronounced diffusive zones occurring in regions characterized by steep or rough topographic structures (Waterhouse et al., 2014). Our measurements reveal diffusivity values, $K_\rho$, exceeding $-4 \text{m}^2/\text{s}$, primarily within a proximity of 300 meters to the rough, steep seafloor. Notably, the area above the Aves Ridge (CDP 13000-18000) emerges as a conspicuous zone of enhanced mixing, displaying elevated levels of diffusivity in comparison to the surrounding waters at equivalent depths (Figure 7.1). Furthermore, the distinct, upward undulations above the ridge are interpreted as large-amplitude Lee waves propagating vertically (Figure 7.1A) (Eakin et al., 2011). These undulations manifest their sharpest slopes near the summit of the Aves Ridge and gradually level off approximately 400 meters above the top of the ridge. Enhanced diffusivity becomes evident beneath the elongated, continuous reflectors marking the boundary of the more stratified thermocline.

A comparison of diffusivity values in the Grenada Basin and above the Aves Ridge at analogous depths shows that the latter exhibits higher diffusivity levels, particularly near the slope. However, it is crucial to recognize the substantial spatial heterogeneity in diffusivity. The deeper zones of the Grenada Basin (Figure 7.1B) manifest lower diffusivity levels, approximating $K_\rho = -5.5 \text{m}^2/\text{s}$, in contrast to the open ocean estimates of $K_\rho = -5 \text{m}^2/\text{s}$ (Munk and Wunsch, 1998). These findings support the notion of increased diffusivity in the vicinity of and above shallower topographical features.
Of noteworthy significance are the observations of the minimal diffusivity values within the central domain of the Antarctic Intermediate Water (AIW) throughout the entire section, where the stratification of waters is markedly pronounced (Figure 7.1). In this region, the diffusivity patterns are predominantly influenced by density-driven processes, in contrast to mechanical processes (Wunsch, 2002).
Figure 7.1: (Top) BOL30 seismic profile with overlain diffusivity map, smoothed from tracked reflectors using a 2km by 20m, half overlapping window. Rectangular boxes show regions of data for diffusivity calculations. (A) Shows the analysis for the region above the Aves Ridge at the northwest end of the profile. The seismic data with the smoothed diffusivity overlain in color (left), and the average diffusivity vs depth for the data window (right). The darker blue line and the shaded region highlight the mean and the one standard deviation, respectively (B) Shows the analysis for the Grenada Basin at the same depth range as in (A). The seismic data with the smoothed diffusivity overlain in color (left), and the average diffusivity vs depth for the data window (right). The darker blue line and the shaded region highlight the mean and one standard deviation from the mean, respectively. The analysis shows the average diffusivity decreasing from \( K_{p} = -4.5 \) to \(-5.5\) m\(^2\)/s with depth in the Grenada Basin, and increasing near and above the steeped sloped of the Aves Ridge.
While tracked reflections offer valuable insights, it is important to acknowledge the limitations of this method to calculate diffusivity values using seismic reflection data. For example, in regions where reflectors have lengths shorter than 1.25 km, tracking becomes unfeasible. Nevertheless, areas characterized by semi-transparency and shorter reflectors hold significant information (Sheen et al., 2009). These regions are generally indicative of enhanced diffusivity and can be addressed using alternative methodologies like Direct Data Transform (DDT). Employing DDT often leads to higher-resolution diffusivity measurements, that are scaled by the diffusivities recorded by tracked reflectors, resulting in improved resolutions for diffusivity estimates (Fortin et al., 2016). This outcome is expected, given that the presence of mixing waters tends to diminish the fine structure associated with thermohaline variations, subsequently reducing seismic energy reflection capabilities at boundaries and gradients. While this particular method is not employed in this thesis, it offers potential avenues for further investigations into diffusivity based on the findings of this study.

7.0.1. Variability Across the southeast Caribbean Sea

As the Antarctic Intermediate Water (AIW) advances westward beyond the Lesser Antilles, the mixing processes become intricate and highly unpredictable (Figure 6.9). Consequently, the salinity minimum core of the AIW decreases due to both vertical and horizontal mixing, and this effect can be observed only until 20°N (Tsuchiya, 1989). The mixing dynamics are influenced not only by turbulence but also by double-diffusive mixing, surface upwelling, and eddy forces, as discussed in Chapter 5. These mechanisms collectively contribute to a dynamic and variable environment characterized by fine-scale mixing (Figures 7.2-7.5).

Van der Boog et al. (2022) showed that the temperature, salinity, and density of the AIW increase along its westward trajectory, and these changes align with modifications induced by vertical fluxes from turbulence and double-diffusive mixing. In the Caribbean Sea, heightened salinity prevails, resulting in the occurrence of double-diffusive salt fingers within the lower Central Water (ACW) and the upper AIW (Schmitt et al., 2005). These features manifest as layered reflectors and stem from the 2 magnitudes larger molecular
diffusivity of heat compared to that of salt (Schmitt et al., 2005).

The estimation of diapycnal diffusivity across each seismic line provides further evidence of prevalent vertical mass exchange (van der Boog et al., 2022). Observable disparities in diffusive characteristics are discernible throughout the profiles, warranting closer examination for future studies. For our current investigation, we aim to present a brief analysis and primarily focus on offering diffusivity measurements for each line.

Of notable importance are the BOL19 and BOL47 seismic lines, located at the center of our study area, which exhibit less densely quantified tracked reflectors (Figures 7.2 and 7.3). In these instances, it is crucial to acknowledge that the tracking algorithm follows coherent reflectors, and broader regions of non-tracked reflectors could indicate heightened turbulence levels, leading to shorter temporal and spatial isopycnal boundaries (Holbrook et al., 2013). Additionally, during the acquisition of BOL19, adverse weather conditions persisted during the data acquisition for CDPs greater than 9700 up to the Cariaco Basin (CDP 40000), necessitating the truncation of the seismic section at 9700 for quantitative analysis. Nonetheless, a distinct feature of elongated reflectors is evident in the upper 600 meters along BOL19 (CDP 10000-19700), with an average $K_\rho$ of -6 m$^2$/s. A similar feature appears across the entirety of BOL47, but primarily in the upper 300m, which is likely associated with the ACW.

Beyond these two seismic lines, more distinct features come into play. Notably, the two filaments observed on BOL12 exhibit an average $K_\rho$ ranging between -4.5 to -5.5 m$^2$/s. However, beneath the filament closest to the Leeward Antilles (CDP 25000-35000), diffusivity is significantly higher, surpassing -4 m$^2$/s (Figure 7.4). This value is closer to the canonical abyssal diffusivity and necessitates further investigation into the mechanisms driving increased mixing in this region (Munk and Wunsch, 1998). Additionally, the deeper water filament (CDP 0-12000) appears to induce shear between the waters above and below it, resulting in fewer tracked reflectors and diffusivity levels an order of magnitude greater (Figure 7.4). While it is likely that the diffusivity within the semi-transparent lens is higher, precise quantification would necessitate detailed DDT analysis.
The BOL3 seismic line exhibits the longest tracked reflectors and, consequently, the lowest $K_\rho$ values (Table 6.1). This line affords a significant opportunity to quantify double-diffusive structures and observe the vertical and horizontal variations toward a gently sloping ridge (Figure 7.5). Notably, we observe $K_\rho$ values dropping below $-6 \text{ m}^2/\text{s}$ at the heart of the double-diffusive structure (CDP 20000-30000), with an increase generally noted towards the southern slope. Around CDP 20000, below 700 meters, the increase in $K_\rho$ can be attributed to turbulence induced by the Southern Caribbean Deformed Belt, which is associated with a prominent bathymetric steep profile (Figure 6.9). In general, sporadic areas of increased diffusivity are observable throughout and are progressively less predictable north of the double-diffusive structures (CDP > 28000) as the profile extends into the smoothed bottom Venezuelan Basin (Figure 7.5).
Figure 7.2: The full BOL19 seismic profile with black box identifying the area of analysis for subsequently tracked reflectors and resulting diffusivity calculations. Estimated turbulent diffusivity, labeled Logkt, was calculated for each individually tracked reflector. The reflector results were blended using a 2km by 20m, half overlapping window, to create a better qualitative understanding of larger regions of observation.
Figure 7.3: The full BOL47 seismic profile with black box identifying the area of analysis for subsequently tracked reflectors and resulting diffusivity calculations. Estimated turbulent diffusivity, labeled Logkt, was calculated for each individually tracked reflector. The reflector results were blended using a 2km by 20m, half overlapping window, to create a better qualitative understanding of larger regions of observation.
Figure 7.4: The full BOL12 seismic profile with black box identifying the area of analysis for subsequently tracked reflectors and resulting diffusivity calculations. Estimated turbulent diffusivity, labeled Logkt, was calculated for each individually tracked reflector. The reflector results were blended using a 2km by 20m, half overlapping window, to create a better qualitative understanding of larger regions of observation.
Figure 7.5: The full BOL3 seismic profile with black box identifying the area of analysis for subsequently tracked reflectors and resulting diffusivity calculations. Estimated turbulent diffusivity, labeled Logkt, was calculated for each individually tracked reflector. The reflector results were blended using a 2km by 20m, half overlapping window, to create a better qualitative understanding of larger regions of observation.
7.1. Spatial and Temporal Variability

Spatial and temporal blurring is a ubiquitous issue in seismic oceanography, although it is generally perceived to be insignificant within the time frames observed for fine-scale features. An investigation by Falder et al. (2016) examined multiple stacks of varying streamer lengths to investigate spatial variability and discovered that as stack lengths decreased, the transition between internal waves and turbulent regimes became more pronounced, but spectral degradation occurred due to a diminished signal-to-noise ratio. The study concluded that within typical seismic acquisition parameters, such as those utilized in this study, the spatial and temporal variability are largely immune to the effects of dispersion. Further supporting this, a two-day timelapse analysis of BOL12 by Christianson (2015) revealed that the vertical movement range for major oceanic features within the AIW is 7-40m. These findings suggest that the spatial and temporal fluctuations of fine-scale features are generally unimpeded by smearing. Nevertheless, we are unable to address regions along water mass boundaries and rough topography, where the mixing rates are 1-2 orders of magnitude greater, resulting in fainter reflectivity and shorter, discontinuous reflectors (Lueck and Mudge, 1997; Sheen et al., 2009; Wunsch, 2002).

In order to estimate a vertical and horizontal diffusive timescale for the AIW, we utilize a similar method from (Buffett et al., 2009). This calculation makes broad assumptions about the AIW properties, since the waters exhibit high variability, the estimation represents an average timescale for diffusivity to fully occur. The vertical diffusive timescale is calculated as $T_d = \frac{sh^2}{kv}$, where $kv$ is the vertical diffusivity and $h$ is the thickness of the reflectors (approximately 8m). We use 1 for s, representing the proportion of property transferred to the adjacent layer as detailed in Buffett et al. (2009). We use our average diffusivity of $kv=10^{-5.2}$ m$^2$/s, based on (van der Boog et al., 2022) for the AIW, who performed a comparable calculation. Hence, we arrive at $T_d \sim 117$ days. For our seismic profiles, the longest coherent reflective layer appears along BOL3 interpreted as a salt finger, yielding $L \sim 50$km. The horizontal diffusivity timescale is calculated using the equation $T_h = \frac{sL^2}{kh}$. Using $kh=125$ m$^2$/s, a generic horizontal diffusivity from (Van Der Boog et al., 2019), we
find $T_h \sim 231$ days for the longest observed horizontal layer.

The horizontal length of these features is typically orthogonal to the Caribbean current general flow. Since the longest horizontal layer was observed in our data and no distinctive features are visible among the profiles, except for the northernmost section of BOL47 and BOL19, we anticipate that the effective diffusivity would produce layers shorter than the distance between each seismic profile (roughly 50km). Using the effective diffusivity equation, $L = \sqrt{D_e \cdot t}$, for an average time scale of subtropical regions ($t=4$ days) and average effective diffusivity ($D_e=10^{-4}$), the layers would last roughly 60km (Chiswell, 2013; Qian et al., 2019). This general calculation supports the reason why we don’t observe analogous features at similar latitudes along different seismic lines.
Chapter 8
CONCLUSIONS

Legacy seismic data collected from April 22 to May 25, 2004, spanning approximately 1000km, East-West, across the southeastern Caribbean Sea were analyzed. The seismic lines clearly display a constant thermocline, detectable water masses, and structures reaching depths of approximately 1000m and sometimes beyond, with the exception of the anoxic Cariaco Basin. Reflective water masses in the southeastern Caribbean Sea, such as the ACW and AIW, can be identified before reaching the relatively homogeneous NADW. Undulating reflectors, ranging from meters to kilometers in length, are imaged in all seismic lines.

Similar to previous seismic oceanography studies (Holbrook et al., 2003; Tsuji et al., 2005; Buffett et al., 2009; Quentel et al., 2010; Eakin et al., 2011), this research successfully identified water mass boundaries and oceanographic features such as thermohaline stair-cases, lee waves, turbulent waves and eddies within the seismic data. Moreover, areas with relatively consistently stacked reflectors are typically associated with double diffusion of salt and temperature variations (Fer and Holbrook, 2008; Biescas et al., 2010). These diffusive features within the AIW are recognizable along BOL3, BOL12, and BOL30 (Figures 5.4, 5.10, 5.12). The interpretations were further augmented by HYCOM oceanographic models to compensate for the lack of in-situ oceanographic measurements, thus providing a deeper insight into the fine-scale mixing within and around mesoscale features. The strong resemblance between time-resolvable models like HYCOM’s 1/12° daily gridded model and seismic images attests to the efficacy of using seismic reflection images and the HYCOM model for tracking lateral continuity of water mass boundaries, fronts, and oceanic features. Even in regions where seismic zones are semi-transparent, the model provides valuable insights, enabling interpretation that would not be possible with in-situ oceanographic measurements alone.
Most importantly, this study successfully implemented spectra analysis using track seismic reflectors to quantify diffusivity in the highly variable Caribbean sea. The average diffusivity measurements from all seismic lines indicate a similar magnitude to Kunze (2003) and Wunsch (2002) with clear regions of increased mixing. Though there is observed heterogeneous diffusivity along all seismic lines, this study confirmed the hypothesis that increased diffusivity strongly correlates with locations of rigid, shallowing seafloor morphology. This key finding supports the idea that these regions can be identified and quantified through the use of seismic reflectivity, particularly, in the lesser-documented southeastern Caribbean sea (Lueck and Mudge, 1997; Fortin et al., 2016; Tang et al., 2022). Additionally, the turbulent subrange is present and similar when derived from both tracked reflectors and direct data transform (DDT) for each seismic profile, providing an opportunity for further investigations through the use of DDT (Holbrook et al., 2013; Fortin et al., 2016).

In conclusion, evidence from a combination of seismic imaging, oceanographic models, and spectra analysis supports that legacy seismic data can be used to quantify and observe fine-scale mixing behaviors. Given the abundant legacy marine seismic data, there are vast opportunities to further explore fine-scale mixing trends throughout the world’s oceans.
Appendix A

A.1. Histograms of Diffusivity

Figure A.1: BOL19 histogram of diffusivity distribution from tracked reflectors

Figure A.2: BOL47 histogram of diffusivity distribution from tracked reflectors
Figure A.3: BOL12 histogram of diffusivity distribution from tracked reflectors

Figure A.4: BOL3 histogram of diffusivity distribution from tracked reflectors
A.2. Additional DDT and Slope Spectrum

Figure A.5: Seismic profile BOL19 showing section where DDT was extracted. Zoom of seismic profile in the top right next to the horizontal wavenumber spectra. The horizontal wavenumber $k_x$ spectrum is calculated directly from seismic data. This is accomplished by performing a 2D Fourier transform of the seismic section, multiplying by $(2\pi k_x)^2$, and summing spectral levels at each $k_x$ value. Amp is the arbitrary units for amplitude. A clear change in slope occurs at 0.004 cpm and typically ends around 0.02 cpm, where the data becomes noise dominant. Tracked seismic reflectors (black lines) on the same zoomed seismic section. Slope spectrum calculated from n=200 tracked reflectors. The shaded region highlights the standard deviation range calculated from 2 degrees of freedom.
Figure A.6: Seismic profile BOL47 showing section where DDT was extracted. Zoom of seismic profile in the top right next to the horizontal wavenumber spectra. The horizontal wavenumber $k_x$ spectrum is calculated directly from seismic data. This is accomplished by performing a 2D Fourier transform of the seismic section, multiplying by $(2\pi k_x)^2$, and summing spectral levels at each $k_x$ value. Amp is the arbitrary units for amplitude. A clear change in slope occurs at 0.004 cpm and typically ends around 0.02 cpm, where the data becomes noise dominant. Tracked seismic reflectors (black lines) on the same zoomed seismic section. Slope spectrum calculated from $n=200$ tracked reflectors. The shaded region highlights the standard deviation range calculated from 2 degrees of freedom.
Figure A.7: Seismic profile BOL12 showing section where DDT was extracted. Zoom of seismic profile in the top right next to the horizontal wavenumber spectra. The horizontal wavenumber $k_x$ spectrum is calculated directly from seismic data. This is accomplished by performing a 2D Fourier transform of the seismic section, multiplying by $(2\pi k_x)^2$, and summing spectral levels at each $k_x$ value. Amp is the arbitrary units for amplitude. A clear change in slope occurs at 0.004 cpm and typically ends around 0.02 cpm, where the data becomes noise dominant. Tracked seismic reflectors (black lines) on the same zoomed seismic section. Slope spectrum calculated from $n=200$ tracked reflectors. The shaded region highlights the standard deviation range calculated from 2 degrees of freedom.
Figure A.8: Seismic profile BOL3 showing section where DDT was extracted. Zoom of seismic profile in the top right next to the horizontal wavenumber spectra. The horizontal wavenumber $k_x$ spectrum is calculated directly from seismic data. This is accomplished by performing a 2D Fourier transform of the seismic section, multiplying by $(2\pi k_x)^2$, and summing spectral levels at each $k_x$ value. Amp is the arbitrary units for amplitude. A clear change in slope occurs at 0.004 cpm and typically ends around 0.02 cpm, where the data becomes noise dominant. Tracked seismic reflectors (black lines) on the same zoomed seismic section. Slope spectrum calculated from $n=200$ tracked reflectors. The shaded region highlights the standard deviation range calculated from 2 degrees of freedom.
Bibliography


