Role of mixing in the structure and evolution of a buoyant discharge plume

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A field study to examine the near-field structure of a buoyant discharge plume, and the turbulent mixing associated with its evolution, was conducted in Mount Hope Bay near Somerset, Massachusetts. The study focused on a 50 m$^3$/s thermal discharge, approximately 5$^\circ$C warmer than ambient waters, emanating from an electrical generating facility. The discharge enters the Bay through a 10 m wide surface canal creating a plume that has many attributes in common with larger scale river plumes. At a distance of 200 m, the plume was characterized by a core with high velocity and 2 $\sim$ 3$^\circ$C higher temperature than ambient water. At this location the core of the plume was observed with a width of approximately 150 m during ebb tide, remaining bottom attached along the plume centerline. Analysis of the temperature budget with respect to specific isotherms yielded values of Reynolds temperature flux $<u'\theta'>$, on the order of $10^{-2}$C m/s, suggesting buoyancy flux values ($B = g\alpha<w'\theta'>$) of order $10^{-5}$ m$^2$/s$^3$. These values are consistent with the turbulent energy expected to enter the system as a result of bottom stress, implying that mixing across the first 200 m is driven by bottom friction and is most active along the edges of the plume where thermal gradients are strong. The industrial plume studied here is dynamically similar to larger geophysical plumes, such as river plumes. However, the aspect ratio is critical in determining whether mixing driven by bottom friction is important in the overall evolution of the plume.


1. Introduction
1.1. General Introduction

Buoyant plumes are generated in the ocean when a lower density fluid is discharged into a higher density environment [Wright and Coleman, 1971; Jirka et al., 1981]. River plumes, the most common type of geophysical plume, form when fresh water flows into an estuary or from an estuary into the coastal ocean [e.g., Fong and Geyer, 2002]. On a smaller-scale, industrial plumes, such as the thermal discharge from a power plant [e.g., Atkinson, 1993] may be dynamically similar to geophysical plumes in many respects.

The spatio-temporal evolution of a plume immediately following discharge is controlled by a complicated and poorly understood relationship between spreading of the plume in the lateral direction, and mixing of the plume with ambient waters (R. D. Hetland, Relating river plume structure to vertical mixing, submitted to Journal of Physical Oceanography, 2006, hereinafter referred to as Hetland, submitted manuscript, 2006). The manner in which a plume evolves, and mixes with the surrounding waters is a major factor affecting local circulation and the fate of the discharge in general, which may have significant implications for the transport of sediments [e.g., Geyer et al., 2004], nutrients [Yin et al., 1995] and pollutants, including heat.

The dynamics of any buoyant plume can be generalized by dividing the plume into two regions: the near field, and the far field. Jirka et al. [1981] define the near field as the region where momentum dominates over buoyancy. In this region the plume behaves much like a buoyant jet, driven by the enhanced velocities of the discharge as it initially enters the coastal region. The momentum surplus of the jet generates significant turbulence, which can result in rapid mixing of the plume with ambient waters. Hetland (submitted manuscript, 2006) suggests that the boundary between the near and far fields be delineated by the point where the composite Froude number of the flow drops below one, implying a transition from supercritical to subcritical flow. Far field dynamics are dominated by buoyancy effects [e.g., Fong and Geyer, 2002], and mixing is controlled primarily by the local wind stress [Lentz, 2004]. The effect of the Earth’s rotation may also play a critical role in the far field [Horner-Devine et al., 2006].

Understanding the nature of near-field plume dynamics will allow us to relate the estuarine outflow to the larger scale coastal environment. The physics in the near-field are responsible for setting up the initial conditions of the far-field plume, such as the water mass properties, including
buoyancy, and essentially act as a bridge between the river or industrial discharge channel and the ocean. With respect to numerical modeling of buoyant plumes, most models of the coastal ocean are computationally capable of modeling the far field, but cannot constrain the inherently small scales and high energy present in the near field. Thus, understanding the near-field is crucial for linking the dynamics of the estuarine and coastal regions.

The similarities between industrial plumes and geophysical plumes are significant, and, given the smaller scales, consistent flows, and enhanced accessibility of industrial plumes, they may play a key role in furthering knowledge of buoyant plumes in general, with much that can be applied at larger scales to geophysical plumes. Similarly, much can be drawn from the essentially separate bodies of literature that exist for the two types of plumes, with industrial plumes historically treated in the engineering literature [e.g., Jirka et al., 1981; Adams and Stolzenbach, 1977], while geophysical plumes have typically been discussed within the oceanographic literature [e.g., Wright and Coleman, 1971; MacDonald and Geyer, 2004; Lentz, 2004]. In order to further such comparison and understanding, this paper evaluates observational data from the near field of an industrial thermal plume. Overall, we seek to characterize the structure and spatio-temporal evolution of an industrial thermal plume, and relate those findings to an understanding of geophysical plumes. Specifically, we also attempt to evaluate the role played by mixing in affecting the evolution of the plume, and identify the processes responsible for generating turbulence and mixing. The following section first provides a brief review of the current understanding of buoyant plume structure.

1.2. Plume Structure

A number of studies on buoyant plumes have been made over the last several decades. Ultimately, most positively buoyant plumes detach from the bottom and become surface trapped. This detachment is often referred to as the “lift-off” of the plume. The nature of this lift-off, and the role of the near field in this detachment is a key component of understanding near field structure. Observations of river plumes have found that lift-off typically occurs uniformly across the cross section [e.g., Wright and Coleman, 1971], consistent with two-layer hydraulic theory [i.e., Armi and Farmer, 1986; Farmer and Armi, 1986], or at an angle to the cross section, consistent with a two-dimensional (2-D) extension of hydraulic theory [MacDonald and Geyer, 2005]. Landward of the lift-off location, the discharging water mass is essentially unmodified, but substantial mixing occurs seaward of the lift-off in both the Mississippi [Wright and Coleman, 1971] and Fraser River [MacDonald and Geyer, 2004] outflows.

Much of the significant work related to industrial discharges has been performed in the laboratory, or using numerical simulations [e.g., Adams and Stolzenbach, 1977]. The location of plume detachment is discussed by Atkinson [1993], for discharges flowing along a uniformly sloping bottom, based on the results of a control volume approach for evaluating the momentum equation in the near field. The water depth at detachment is found to increase with bottom slope and decrease with increasing bottom friction, entrain-

2. Observational Program

2.1. Study Site: Mount Hope Bay

Mount Hope Bay (MHB) makes up the northwest corner of Narragansett Bay, lying within both Rhode Island to the south and west and Massachusetts to the north and east (Figure 1). It covers an area of approximately 36 square kilometers. MHB has a depth at mean low water of about 5 meters or less over much of its area. Tides in this region are predominately semidiurnal, and tidal range varies from 0.9 to 2.2 meters between extreme neap and spring tides. The bay adjoins the east passage of Narragansett Bay to the southwest through a 1 km wide passage. The Sakonnet River provides a direct connection south to Rhode Island sound. The Taunton River is the major source of fresh water to the bay, and four smaller rivers provide an additional input of fresh water.

In recent years questions have been raised concerning the effect of the 1600 MW fossil fuel-fired power station at Brayton Point, Massachusetts, on the MHB ecosystem (Figure 2). Within the Brayton Point Power Station (BPPS), steam is condensed by a once through cooling water system, drawing cold water from MHB at an average maximum rate of approximately 50 m$^3$/s. The discharge of heated wastewater occurs through a canal designed with a 10 m wide venturi structure to enhance exit velocities and encourage mixing with ambient waters. The bottom of the discharge channel is at the same elevation as the bottom of the bay, resulting in a depth of approximately 5 m at mean low water. The near field to far field transition occurs on the order of 1 km from the discharge location, and presumably there is significant mixing of the discharge within the first kilometer. Only sparse data are available within the near-field region, however, and no estimates of turbulence and mixing exist within that region.

Several previous studies have focused on the thermal plume in MHB, generally oriented towards thermal conditions in the far-field. Thermal mapping studies, which consisted of a set of thermistor strings to map the temperature field of MHB, were conducted by Applied Science Associates (ASA) in 1997, 1998, and 1999. These data are valuable for studies of the far field plume; however, the ASA data set lacks the necessary velocity and density information within the near field required to quantify initial mixing processes. Mustard et al. [1999] analyzed the seasonal variability of surface temperatures in the Narragansett Bay region from thermal infrared satellite images, concluding that Mount Hope Bay exhibited average “sea skin” temperatures approximately 0.8°C warmer than other comparable regions of the greater Narragansett Bay region during autumn. These results are not necessarily reflective of Mount Hope Bay as a whole, due to the buoyancy associated with the thermal plume and the tendency for the heat to be trapped at the surface. Regardless, the resolution of the thermal imagery is insufficient to assess the near field region independently of the much larger expanse of the far field plume.

In order to close this gap in data with respect to the near field, and to further understanding of near field plumes
in general, a field experiment was conducted in September 2004 to assess the essential physics controlling the initial growth of the BPPS thermal plume, and the related dissipation of heat within the near field region. Essential data was collected to enable the calculation of key turbulence parameters through the use of volume and temperature budgets, and the development of a conceptual model of the near field plume structure and evolution.

2.2. Observations

[13] A field study to understand the structure and evolution of the near field of the thermal discharge plume in MHB was conducted on September 7, 2004. The data set included measurements of velocity, salinity and temperature from repeated passes at two transects across the thermal plume, located at 200m and 800m away from the discharge channel, respectively (Figure 2), as well as data provide by BPPS. Data was collected from shipboard instrumentation aboard the R/V Lucky Lady, the University of Massachussets Dartmouth research vessel.

[14] Mean velocity profiles were monitored with a vessel-mounted RDI 1.2 MHz broad-band acoustic Doppler current profiler (ADCP). Profiles were collected continuously using 25 cm bins, and averaged over 15 seconds. The vertical profiles of velocity extended from 1 m below the surface to approximately 85% of the water depth. During processing, velocities were extrapolated across the entire water column, assuming a quadratic profile, a no slip condition at the bottom, and a surface stress equal to zero. Direction and position information were obtained from the shipboard GPS.

[15] Profiles of pressure, temperature and conductivity were monitored in real time using an Ocean Sensors OS-200 conductivity/temperature/depth (CTD) profiler. The frequency of OS-200 sampling was approximately 7 Hz. The CTD unit was typically towed at four different depths, including the surface and approximately 0.5 m, 2 m, and 3 m, respectively, at both transects, in addition to vertical casts approximately every 100 m. The typical horizontal CTD towing speed was approximately 1.5 m/s, and the typical vertical CTD lowering speed was 0.3 m/s.

[16] The time required to complete a set of four passes was approximately 25 minutes for the 200 m transect, and 40 minutes for the 800 m transect. In both cases these time periods are assumed to be small compared to the tidal timescale, and all data from each set of passes was combined and resampled to a horizontal resolution of 40m. The vertical resolution of 0.25 m provided in the initial data set was maintained. Combining CTD data from the vertical casts and horizontal tows resulted in a 2-D grid of high

Figure 1. Locus of Mount Hope Bay, Massachusetts.
resolution salinity and temperature measurements, which was subsequently interpolated with a horizontally biased weighting scheme onto the 40 m by 0.25 m grid. The 200 m transect was sampled a total of 5 times and the 800 m transect was sampled 6 times, as shown in Figure 3. Note that the sampling period extended from low tide to high tide.

Conditions at the discharge outlet are taken from records provided by BPPS. The BPPS data set includes the average temperature of the water at intake and discharge, station flow rates, and weather data, such as wind speed and direction, relative humidity, air temperature at 10 m and 2 m above the sea surface, and solar radiation. Average values during the measurement period were 23.2°C and 28.4°C for intake and discharge temperatures, respectively, and 45.6 m³/s for the discharge flow rate. The air temperature increased from 25°C in the morning to 29.5°C by noon, and the average relative humidity was 78%. The wind direction was from the southeast with a speed of approximately 7 m/s during the sampling period. Previous CTD casts within the discharge channel, performed on June 17, 2004, show the water to be well mixed vertically.

3. Measurement Results

A local coordinate system was established with the X axis along the direction of the discharge channel, the Y axis across the plume from northwest (−) to southeast (+), and the Z axis positive upward.

Figure 4 depicts the mean velocity and temperature profiles for three of the five passes at the 200m transect, on September 7, 2004. The salinity profiles are not shown here because salinity in this region is relatively well mixed in the vertical, thus the density is dominated by temperature.

The core of the plume is identifiable as the area of warmer temperatures observed near x = 0. At low tide, the width of this region is approximately 150 m. It becomes wider as the flood progresses. By the last pass, which is closest to high tide, the width reaches more than 300 m. The reason for the growth of plume size with flooding may be explained by a decrease in the along plume pressure gradient as the tide transitions from ebb to flood, coincident with a reversal of velocities in the ambient waters.

Maximum velocities within the plume at the 200 m transect reach 0.45 ~ 0.60 m/s at low tide. The maximum velocity core also gets wider from low tide to maximum flooding, coincident with the temperature profile, but the maximum velocity component decreases at high tide, to approximately 0.30 m/s. Within the core, the volume flux and heat flux are much larger than other portions of the transect (Figure 6).

The issue of plume bottom attachment plays an important role in the dynamics at the 200 m transect. During pass 1, the plume is clearly bottom attached, as evidenced...
by data from a CTD cast located within the center of the plume. On subsequent passes, no vertical CTD casts were performed within the core of the plume due to shipboard limitations, and the vertical data from either side of the plume is insufficient to address the issue. However, the ADCP records can be used similarly to infer bottom attachment, and indicate clearly that the plume is bottom attached at the 200 m transect during passes 1, 2, and 3. At some point between pass 3 and 4, as the flooding tide decreases the along plume pressure gradient, resulting in a deceleration of the plume, the location of liftoff transitions to the discharge side of the 200 m transect.

[23] Temperature and velocity profiles similar to those shown in Figure 4, are shown in Figure 5 for the second of the six passes at the 800 m transect. In general, the maximum temperature in the core decreases by approximately 0.7°C between the 200 m and 800 m transects across the duration of the observation period. Note that a representative Froude number calculated at the 800 m transect is slightly greater than one, suggesting that the near field to far field transition lies to the seaward side of this transect, on the order of 1 km from the discharge channel.

[24] In order to delineate the heat budget, an estimate of heat flux across the water surface was estimated using weather data provided by the BPPS weather station and sea surface temperatures as recorded by the CTD. Mean heat flux across the surface for the first 200 m of the plume was found to be of order 10^9 W/m^2, which is two orders of magnitude smaller than the apparent heat loss suggested by Figure 6. Hence, we consider surface losses across this small region to be negligible.

[25] A key first order estimator of mixing intensity is volume dilution rate. The total volume dilution rate across the first 200 m and the first 800 m are estimated by comparing the integrated observed volume flux within the core of the plume at each transect with the known discharge volume, as shown in Figure 7. Within the first 200 m there is a volume dilution of roughly a factor of 4, indicating substantial mixing across this short initial distance. The total volume dilution between the discharge and the 800 m transect has a temporal mean through the sampling period of approximately 7, implying that the average dilution rate decreases by roughly an order of magnitude in the 200 m to 800 m region as compared to the 0 to 200 m region. Variability with time in the 800 m dilution values is likely due to the fact that a portion of the outflow appears to have been missed by the sampling transect for some passes. [26] The recovery of heat at each transect as compared to the known discharge of heat is also shown in Figure 7. The ratio of recovered heat to initial discharge is approximately equal to one at low tide for both transects. Values slightly less than one are likely due to the lateral loss of heat. However, these losses appear negligible. As the flooding tide progresses, more heat is observed at each transect than is being discharged, which may suggest a recirculation of “old” plume water into the core of the plume driven by the advancing flood waters.

4. Turbulence Analysis

4.1. Methods: The Control Volume Approach

[27] In order to evaluate the spatial evolution of the thermal plume, we need to connect point measurements to a more integral view of the plume dynamics and to estimate system-wide mixing rates and buoyancy fluxes. A control volume approach [MacDonald and Geyer, 2004] was used in the estimation of the turbulent transport quantities. This approach relies on conservation of two quantities: volume and heat. The two unknowns are the mean velocity across the bounding surface of the control volume (assuming the top surface of the control volume is taken coincident with the sea surface), \( u_n \), which is shown in Figure 8, and the turbulent transport. The mean velocity across the bounding surface of the control volume is highly dependent on the
inclination of the bounding surface, which serves as the reference frame for the normal directed velocities. Because the temperature at every point along the bounding surface is required for later calculations, isotherms are chosen as a natural reference frame for the bounding surface.

Volume conservation is used to estimate the mean velocity across the control volume bounding surface, which can be expressed as an integral across the control volume surface:

$$\iint_{C} \mathbf{U} \cdot d\mathbf{A} = \frac{\partial V}{\partial t}$$

where $\mathbf{U}$ represents the local velocity at the control volume surface, $A$ is the area of the bounding surface, and $V$ is the

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**Figure 4.** Velocity (cm/s) and temperature (°C) at the 200 m transect on September 7, 2004, for passes (a) 1, (b) 3, and (c) 5. Velocities directed out of the discharge channel have a positive sign, and are shown shaded in the above plots. The location of ADCP ensembles are shown by the dots at the top of each velocity panel. In the temperature panels, values greater than 23°C are shaded, and the path of the towed CTD unit is shown by the gray line in the background.
volume of sea water within the control volume. In our study, as there is no contribution for the volume flux from sea surface and bottom, the control volume continuity equation can be written in differential form as:

\[ u_n A = \int \int u_2 dydz - \int \int u_1 dydz + \frac{\partial}{\partial t} \int V dV \]  

(2)

where the subscripts “1” and “2” represent the end surfaces of the control volume, in our case oriented perpendicular to the direction of the outflow. These surfaces represent the majority of the volume flux entering or leaving the control volume, and they are connected by a bounding surface coincident with an isotherm as described above, across which the transport is represented in equation (2) by \( u_n A \), where \( u_n \) represents the mean velocity normal to the bounding surface and \( A \) is the area of the bounding surface.

A similar analysis of heat conservation gives the estimation of the total heat flux \( T u_n \) across the bounding surface, which is taken as a series of isotherms, except that we need to consider heat flux through the air-sea interface. The integral heat balance can be written as:

\[ \int \int T U \cdot dA = \frac{\partial}{\partial t} \int V dV \]  

This can be rewritten in differential form as:

\[ T u_n A = \int \int T u_2 dydz - \int \int T u_1 dydz + \frac{\partial}{\partial t} \int V T dV \]  

(3)

\[ + \text{Air} - \text{Sea Loss} \]

where \( T \) represents temperature.

Given the total heat flux through the isotherm surface, the turbulent part can be extracted through Reynolds decomposition:

\[ T u_n' = T u_n - \overline{T u_n} \]  

(5)

where \( T \) is the temperature of the isotherm coincident with the control volume bounding surface.

The turbulence calculations presented below focused on the region between the power station discharge and the 200 m transect. Turbulence calculations between the 200 m and 800 m transects could not provide reliable results, likely due to errors in estimating the time dependent term in equation (2), and unquantified lateral processes.

### 4.2. Turbulence Analysis

The mean velocities normal to isotherms, \( u_n \), as calculated from the control volume analyses are shown in Figure 9. All of the velocities through the isotherms with temperature higher than approximately 23°C are negative, indicating that water from the core of the plume is being transported across isotherms, directed away from the plume. Positive entrainment velocities at cooler temperatures suggest that cooler water is crossing isotherms, directed towards the plume, and resulting in the growth of an intermediate entrainment layer, where \( u_n = 0 \), which is in the vicinity of the isotherm 23.3°C for most of the passes, with the exception of pass 1. The growth of an entrainment layer is consistent with cooling of the plume through mixing processes, as it represents a growing mass of fluid resulting from the mixing of discharge waters with cooler ambient waters.

The total transport of temperature, \( u_n T \), across isotherms is shown in Figure 10. Figure 10 is similar to the velocity profiles of Figure 9, in that it suggests a growing entrainment layer between 23 and 23.5°C, and that there is a general trend towards decreasing transport from low tide to high tide.

The Reynolds temperature flux, \( u_n' T' \), was obtained using Reynolds decomposition after the total vertical transport \( u_n T \) was calculated using equation (5). The vertical Reynolds flux for temperature, \( w' T' \), was estimated by decomposing \( u_n' T' \) into vertical and horizontal components, based on the observed orientation of the isotherms at the 200 m transect. The observed angle of the isotherms with reference to the horizontal ranged from 0° to 15°, suggest-
ing that the majority of transport may be occurring in the vertical. Implicit in this transformation is the assumption that the isotherms are similarly oriented throughout the volume. Further discussion of this topic and the relative importance of horizontal vs. vertical mixing can be found in the next section.

Given the Reynolds temperature flux, $w^T$, buoyancy flux was estimated as:

$$B = g \alpha w^T$$

where the thermal expansion coefficient $\alpha$ is taken as a constant value of $2.5 \times 10^{-4}$ K$^{-1}$, corresponding to a temperature of 20°C and salinity of 35 practical salinity units (psu). Buoyancy flux represents the rate mechanical energy is converted to potential energy through mixing of higher density (cooler) fluid upwards in the water column. Profiles of $B$ are shown in Figure 11. The Reynolds temperature flux $w^T$ is on the order of $10^{-5}$°C m/s, with corresponding buoyancy flux values of the order $10^{-5}$ m$^2$/s$^3$.

The buoyancy flux decreases slightly from low tide to high

Figure 6. Distributions of volume and heat flux relative to cross plume location at both the 200 m (left) and 800 m (right) transects. Fluxes directed away from the discharge channel are positive. Passes 1, 3, and 5 at the 200 m transect, similar to Figure 4, and passes 2, 4, and 6 at the 800 m transect are shown. The time associated with each pass is shown in each panel. In each panel, the region considered to be the core of the plume is shaded.
tide, but remains generally of the same order throughout the duration of the sampling period.

5. Discussion

5.1. Mechanisms of Mixing

Mean values of buoyancy flux within the first 200 m of the plume were evaluated in the previous section. However, the analysis does not provide any indication as to what physical mechanisms serve as the source for the production of the turbulent kinetic energy that leads to mixing. Two significant mechanisms, which bear further evaluation, are the generation of instabilities along density interfaces due to velocity shear [e.g., Thorpe, 1971; Ivey and Imberger, 1991], and the generation of turbulence within a bottom boundary layer as a result of bottom friction [e.g., Schlichting, 1979].

The gradient Richardson number \( Ri_g = (- \frac{\partial \theta}{\partial z} / \frac{\partial u}{\partial z}) / ((\frac{\partial u}{\partial z})^2) \) is an effective tool for understanding the importance of stratification in a turbulent field [e.g., Thorpe, 1973; Koop and Browand, 1979]. Values of \( Ri_g \) were found to be on the order of \( 10^{-2} \) within the core of the plume for all passes, significantly lower than the critical value of \( 1/4 \) considered necessary for the development of instabilities in a stratified flow [Miles, 1961; Howard, 1961]. Such low values of \( Ri_g \) in the core of the plume suggest that the flow is strongly turbulent, with a lack of strong density gradients; thus the turbulence is unlikely to drive significant buoyancy fluxes. Although turbulence near the center of the plume may be intense, the observed values of \( Ri_g \) indicate that the turbulence is not limited by density stratification, and that it is likely only resulting in the mixing of already well mixed water, similar to the processes described by Nepf and Geyer [1996] for mixing within the bottom boundary layer of the Hudson River Estuary. In the present case the bottom boundary layer may be considered to extend across the entire depth.

Mixing against a density gradient is certainly occurring, however, as evidenced by observed buoyancy flux values on the order of \( 10^{-5} \text{ m}^2\text{s}^{-3} \). Extending the argument that the turbulence is driven by bottom friction, the observed buoyancy flux rates can be explained by using a scaling analysis for the bottom drag. Given the low values of \( Ri_g \) within the plume region, we assume that stratification does not significantly alter velocity profiles, and following Simpson et al. [1990], buoyancy flux from bottom drag can be estimated as \( B \sim Ri_f C_D \frac{\partial u}{\partial z} u^3 h \), where \( Ri_f \) is flux Richardson number, \( C_D \) is drag coefficient, \( u \) is the mean velocity, and \( h \) is the depth. Using values of 0.2 and \( 3 \times 10^{-5} \) for \( Ri_f \) and \( C_D \), respectively, and with representative values of \( u \) equal to 0.5 m/s and \( h \) equal to 5 m, yields a buoyancy flux on the order of \( 10^{-5} \text{ m}^2\text{s}^{-3} \), which is consistent with the order of the observed buoyancy flux.

Buoyancy flux rates due to shear instabilities, independent of bottom friction, can be assessed using the two independent variables \( \Delta\theta u \) and \( g' \) [Imberger and Ivey, 1991]. Measurements in the Fraser River estuary and elsewhere [MacDonald and Geyer, 2004] suggest that buoyancy flux in a stratified shear environment should scale as \( B = \)
is a mixing efficiency (typically taken as $\gamma \approx 0.2$), and $A$ is a constant found to be equal to approximately $2 \times 10^{-5}$. Based on the velocity and temperature observations shown in Figure 4, representative values of $\Delta u$ and $g^*$ for the first 200 m of the thermal plume are approximately $0.5 \text{ m/s}$, and $0.007 \text{ m/s}^2$. These values yield an estimate of buoyancy flux equal to approximately $1.4 \times 10^{-6} \text{ m}^2\text{s}^{-3}$. This value is one order of magnitude lower than the observed buoyancy flux, and the value scaled using a bottom drag law. This result, in combination with the low observed values of $R_i$, strongly suggests that shear stratified mechanisms are not responsible for the mixing observed in the first 200 m of the plume.

5.2. Horizontal Versus Vertical Mixing

[40] The analyses presented above support the conclusion that the turbulent energy responsible for mixing is driven by bottom friction, and suggest that vertical mixing plays the dominant role in transforming the plume water mass. These points are consistent with the conclusion reached by decomposing the Reynolds temperature flux into horizontal and vertical components based on the inclination of the isotherms.

[41] Earlier studies of thermal discharge plumes suggest that such plumes are strongly affected by the lateral intrusion of colder, ambient fluid at the base of the plume, resulting in a triangular liftoff footprint, and that the majority of mixing is horizontal in nature [Adams and Stolzenbach, 1977]. The present observations, which point to the importance of vertical mixing processes, are in direct disagreement with these earlier studies, and further exploration of this conclusion is warranted. An additional method of evaluating the relative importance of horizontal and vertical mixing is to compare the observations with what might be expected under a frictionless (or no mixing) scenario. Because horizontal mixing cannot redistribute heat in the vertical, we can address the importance of vertical vs. horizontal mixing by assessing the vertical distribution of heat, particularly with respect to the location of the liftoff point under the observed and hypothetical scenarios.

[42] For the frictionless plume, assuming there is no mixing or bottom friction, we can estimate the location of the point where the plume has fully lifted off from the bottom by utilizing theory related to the “lock exchange” problem [Schijf and Schonfeld, 1953], which describes the motion of two sharply separated liquids of different density, once the barrier between them is removed. In this case, both edges of the plume can be considered separately, with the passage of fluid across the endpoint of the discharge channel representing a sudden removal of the barrier between the discharging fluid and the cooler ambient waters. In this sense, the lock exchange occurs in the plane perpendicular to the direction of plume outflow, which only serves to advect the lock exchange process in the seaward direction. The lateral velocity of both the warmer discharging fluid, which begins to spread outward at the surface, and the cooler ambient water, which encroaches upon the plume structure at the bottom, can then be described by the internal wave speed associated with the density anomaly between the two fluids. According to the inviscid theory, each fluid should travel at a velocity equal to $\frac{1}{2}\sqrt{g'H}$. Using representative value for $g'$ of 0.02 m/s$^2$ consistent with the density difference between discharged and ambient fluids, and an $h$ value of 5 m, the resulting lateral velocity would be approximately 0.16 m/s. Given a discharge velocity, $u_{fr}$, on the order of 2 m/s, the trajectory of the bottom front can be assessed as it encroaches inward, ultimately tracing the triangular liftoff footprint described by Adams and Stolzenbach [1977]. With a channel half width, $L_{ch}$, of 5 m, the downstream distance of the lift off point, $x_{lo}$, can be estimated as $x_{lo} = 2L_{ch}u_{fr} (g'H)^{-1/2}$, yielding a value of approximately 60 m. This distance is more than a factor of 3 smaller than the distance to the 200 m transect, where the plume was consistently observed to be bottom attached. Because horizontal mixing processes cannot result in a redistribution of heat in...
the vertical, vertical mixing processes must be responsible for depressing heat in the water column and maintaining bottom attachment to the 200 m transect and beyond.

5.3. Conceptual View of Plume Structure and Evolution

[43] Taking an integrated view of the above discussions, we can describe the plume structure and its evolution over the first 200 m, as well as its further evolution towards a passive far field plume. The generic structure described here is likely to be consistent for a wide variety of plumes, from small scale industrial discharge plumes and tidal creeks to larger scale river plumes, particularly those that enter the ocean at supercritical speeds, such as the Mississippi [Wright and Coleman, 1971] or the Fraser [MacDonald and Geyer, 2004] Rivers. However, it is also likely that certain physical mechanisms may be more dominant at different scales, as discussed further in section 5.5.

[44] Plume water is initially discharged as a jet from the outlet channel as a well-mixed mass of fluid with a significant velocity and temperature anomaly. The initial reaction of the discharged plume water is to follow the inviscid model, and expand laterally at the top, while colder water encroaches upon the plume at depth. However, significant velocity shear in conjunction with a rough bottom results in strong bottom friction, and a boundary layer that may well extend across the entire water column. The resulting turbulence generates significant vertical mixing, although in the core of the plume, where the fluid remains well mixed vertically, the turbulence results in no net vertical transport of heat.

[45] The enhanced turbulence generated by bottom friction does play an important role, however, along the edges of the plume, where cold ambient water begins to encroach upon the plume near the bottom. In this region, the bottom driven vertical turbulence mixes heat down into the encroaching fluid, slowing its progress towards the center of the plume, and extending the length of bottom attachment as compared to the frictionless case. The cartoon in Figure 12 indicates the structure of the plume during this phase of its evolution, and the important regions with respect to vertical mixing.

[46] Eventually, the plume will completely lift off from the bottom. The present observations suggest that this happens somewhere between the 200 and 800 m transects. This will happen when either the lateral encroachment of the plume by colder ambient waters at depth overcomes vertical mixing processes and closes in across the entire plume, or when the local bathymetry triggers the formation of a bottom front. Lift-off in the latter case can be well described by classic one dimensional two-layer hydraulic theory [e.g., Armi and Farmer, 1986; Farmer and Armi, 1986], or more recent two-dimensional extensions of hydraulic theory [MacDonald and Geyer, 2005], which are beyond the scope of the present study. Atkinson [1993]

Figure 11. Buoyancy flux, B, derived from \( \bar{u}_g T' \) as described in text, for the 0 m to 200 m region.

Figure 12. Cartoon of plume structure prior to liftoff. The dynamics of buoyancy cause the plume to widen near the surface and become narrower at the bottom. Mixing caused by bottom friction redistributes heat in the vertical, particularly along the edges of the plume, increasing the along plume distance across which the plume remains bottom attached.
and others have predicted lift-off points for a uniformly sloping bottom. The practical irregularities in the bottom surface may change the nature of the lift-off process, which may ultimately be best described by hydraulic theory. Thus, discharges with higher initial Froude numbers, which is typical for smaller scale industrial discharges, may remain bottom attached for longer distances (all else being equal), than those with lower initial Froude numbers, as the flow must have sufficient time to decelerate towards a critical value of the Froude number. Based on the geometry of the discharge channel, the discharge Froude number at Brayton Point varies between approximately 3 and 7, for high tide and low tide, respectively.

47] Once the plume lifts off completely, it enters the outer portion of the near field, where shear stratified mixing processes control the turbulent environment. As discussed above, shear stratified processes are roughly one to two orders of magnitude less efficient at producing turbulent kinetic energy than bottom friction [e.g., MacDonald and Geyer, 2004], so mixing rates can be expected to decrease, particularly as the plume continues to lose momentum and velocity shears are reduced. This is confirmed by the volume dilution rates inferred from Figure 7, which suggest that mixing in the 200 to 800 m region is roughly an order of magnitude less as compared to the inner near field region (0–200 m).

48] At some point, the momentum of the plume will no longer play a key role in the dynamics, and the plume transitions to the far field, where buoyancy is the dominant mechanism for continued plume evolution, and mixing is dominated by wind stress and other environmental factors [e.g., Lentz, 2004]. This region is also where the role of heat transfer across the air-sea interface is expected to be greatest, due primarily to the large surface area of the far field plume [Fan and Brown, 2006; Yanagi et al., 2005].

5.4. Tidal Dynamics

49] The sampling period extended from low tide to high tide, covering most of the flooding tide. Thus, the changes of mixing dynamics associated with tidal variability across an entire tidal cycle cannot be directly characterized with the existing measurements. Tidal effects associated with many river plumes are often the result of increasing tidal stage resulting in significant storage of fresh water discharge within the estuary and in adjacent creeks and marshes. This stored fresh water is then released during the ebb, resulting in a pulsed outflow effect. In the case of the Brayton Point discharge, there is little additional storage capacity within the canal structure, and the discharge rate is sufficiently high that the flow adjusts quickly to changes in tidal height. Thus, the flow rate at the mouth of the discharge is effectively constant through the tidal cycle, and the only affect of the tides is to change the depth of water at the outlet, and in response, the discharge velocity. The parameterization of TKE production discussed in section 5.1 leads to a dependence on $h^{-4}$. Given a change in water depth of order 50% between low tide and high tide, a corresponding change in TKE production would be approximately a factor of 5, which is consistent with the difference between Pass 1 and Pass 5 as shown in Figure 11. Although the analysis of mixing mechanisms and plume evolution have been based on measurements obtained during the flooding tide, conditions during the ebb are not expected to be significantly different, and the structure of the thermal plume is not likely to be greatly affected by the phase of the tide.

5.5. Implications for Geophysical Plumes

50] There are many similarities between industrial plumes similar to the one described here, and larger scale geophysical plumes, most notably those formed by river discharge into the coastal ocean. Certainly, the scales of river plumes are typically large enough that the effect of the Earth’s rotation becomes important in the far field [e.g., Fong, 1998]. In the near field, most river plumes are narrow enough that Coriolis effects are negligible [e.g., Wright and Coleman, 1971; MacDonald, 2003], and in this region the most useful comparisons between industrial plumes and geophysical plumes can be drawn.

51] The most striking difference between industrial plumes and geophysical plumes is the aspect ratio (width/depth) of the flow as it discharges into the coastal ocean. The plume evaluated in the present study is approximately 10 m wide and 2 m deep at the discharge point, with an aspect ratio of approximately 5. In contrast, the Fraser River plume (British Columbia, Canada) is approximately 1000 m wide and 10 m deep, yielding an aspect ratio on the order of $10^2$ [MacDonald and Geyer, 2005], which is not atypical for larger flows. In all cases, it might be expected that local dynamics within the two types of plumes are similar, but the difference in aspect ratio suggests that different mechanisms may dominate the evolution of each plume in the seaward direction.

52] For instance, the mixing mechanism described here, arising from the interaction between bottom friction and encroaching dense fluid along the edges of the plume, should be expected to occur in both types of plumes. In the present case, the plume is narrow enough that this mechanism results in mixing a large percentage of the initial plume volume before liftoff occurs. In the geophysical case, the bulk of the plume lies within the interior, and is unaffected by mixing processes at the edges. Once the geophysical plume lifts off, shear stratified mixing processes, while presumably an order of magnitude lower than bottom friction driven mixing in local intensity, are imposed over an area that is several orders of magnitude larger than the edge regions initially exposed to boundary layer mixing. Hence, for the evolution of the plume as a whole, shear stratified mixing processes tend to dominate in geophysical plumes.

53] This distinction aside, the smaller scale of industrial plumes allows greater accessibility and for more detailed observations to be made with fewer resources than are required to study large river plumes such as the Mississippi or Fraser River outflows. Industrial plumes also offer the advantage of consistent outflow conditions that are less dependent on seasonal cycles and local weather conditions, and that adjust more quickly to the tides. For these reasons, there is the potential to learn a great deal about the nature of geophysical plumes by using industrial plumes in a laboratory-like capacity. For instance, despite the fact that industrial plumes might be greatly modified by boundary layer induced mixing prior to liftoff, the relationship between mixing and spreading in the shear stratified, near field
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