TURBULENCE IN LOW ENERGY LACUSTRINE ENVIRONMENTS AT THE SEDIMENT-WATER INTERFACE

A Dissertation
Presented to the Faculty of the Graduate School of Cornell University in Partial Fulfillment of the Requirements for the Degree of Doctor of Philosophy

by
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The sediment-water interface is an important region where biological, chemical, and physical processes occur. Fluid flow determines the ability of organisms to utilize the bottom sediments for food, shelter, and reproduction, the amount and rate of mass transfer between the sediment and bulk fluid, and the deposition, re-suspension and transport of sediment. Field experiments carried out in three medium sized basins, one a heavily polluted lake and the other two drinking water reservoirs, are used to describe typical turbulence levels during the mid to late stratified season (July through October) in a low energy lacustrine bottom boundary layer.

Using these measurements, a turbulence chamber was developed for use in turbulent scalar flux studies. The chamber was characterized using particle image velocimetry (PIV), characterizing the near bed turbulence and dissipation levels. The chamber successfully reproduces the range of turbulent energy levels observed in the field, allowing a more direct comparison of results obtained in field and laboratory studies of scalar fluxes.

Results from the two reservoirs were used in laboratory cohesive sediment erosion and resuspension tests of a mono-disperse kaolin clay and natural sediment cores obtained from various bottom and shore locations in one of the reservoirs. This testing showed the small bed stresses typical of the lacustrine bottom boundary layer were not sufficient to erode or resuspend significant quantities
of sediment. Higher stress levels caused erosion and resuspension, but it was heterogeneous in nature and appeared to be tied to flow structures associated with the facility used to carry out the experiments. Even at higher stress levels, very short settling times (only 1-2 hours) were needed for observable erosion to occur. A method for estimating erosion utilizing images of the sediment-water interface and tracking the interface as an intensity peak over time was developed. Initial results show this is a reliable means to gauge the sediment-water interface position when optical access is available.
BIOGRAPHICAL SKETCH

Peter J. Rusello spent 10 years in Ithaca attending Cornell University, the last four as a graduate student working on a PhD. More information can be found at rusello.net/bio.
This document is dedicated to my faithful steed *Schwarzbier* who has yet to go kaput and hopefully won’t before I’m truly done.
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CHAPTER 1

INTRODUCTION

The sediment-water interface is an important region in aquatic systems for biological, chemical, and physical processes. The sediment is home to large populations of organisms and serves as a source and sink for chemical and particulate inputs into the system including nutrients (phosphorus, nitrogen), contaminants (mercury), and sediments.

The interaction of flow with the benthic environment, particularly at the sediment-water interface, determines the ability of organisms to use habitats for feeding, shelter, and reproduction, fluxes of dissolved compounds (nutrients, contaminants, etc.) and the flux and retention of particles (Nowell and Jumars, 1984; Lorke et al., 2003b; Mackenthun and Stefan, 1998). Understanding the role of fluid dynamics, in particular the role of turbulent flow, in these processes is an important issue in lake and reservoir management, particularly where impacted waters are a concern (Effler, 1996; Auer et al., 1996; Owens et al., 2011; Gelda et al., 2009).

The interaction of flow with the sediment surface is influenced by the type of sediment (sandy or silty, fine or coarse) and the size, shape, and spacing of bedforms (Raupach and Antonia, 1991; Jimenez, 2004). Bed morphology can be influenced by the flow, resulting for instance in dune or ripple formation (Kennedy, 1969), by aquatic organisms such as fish and invertebrates (Statzner et al., 2003a,b). Jorgensen and Revsbech (1985) and Reidenbach et al. (2010) have shown the ability of boundary roughness to increase the flux of dissolved chemicals at the sediment-water interface.
From an ecological perspective, the flow environment can fundamentally structure ecological dynamics and is the primary determinant of physical habitat in streams (Bunn and Arthington, 2002). Small scale modification of flow patterns as well as physical changes to the bed may affect transport and retention of particulates at the bed. Feeding divots (or similar structures like nests and burrows) are regularly found in marine and estuarine environments (e.g., feeding depressions by rays; VanBlaricom (1982); Yager and Nowell (1993)), but are also common in streams with fine sediments and benthic feeding organisms. In marine environments, feeding depressions have been shown to alter boundary layer dynamics compared to a smooth bed, creating potential retention structures for coarse particulate organic matter and novel habitat structures for invertebrates (Yager and Nowell, 1993).

From a sediment transport perspective, the overlying flow determines the behavior of particles (mobile versus immobile, transport as bedload, saltation, or suspended load) (Shields, 1936; van Rijn, 1993). While granular materials are common, the behavior of cohesive sediments (small particles O(1-50 µm) in size) is important given the ubiquity of mud (a colloquial term for cohesive sediments) in natural environments (Black et al., 2002) and the potential problems caused by small particles such as turbidity (Peng et al., 2009).

1.0.1 Organization

This dissertation encompasses several different projects motivated primarily around understanding and characterizing the flow in benthic environments. Several field experiments (Chapter 3) were conducted to characterize the flow...
environment in three medium sized basins during the summer stratified season. These field measurements in turn guided laboratory experiments examining the erosion of cohesive sediments (Chapter 5) and the development of a turbulent chamber for studying scalar flux at the sediment water interface (Chapter 4).

Chapter 2 provides details on the basic methods used to analyze velocity measurements and an overview of instruments and measurement techniques. Measurement error and bias is also discussed, particularly important as many of the flows studied are characterized by low energy and slow flows. Instrument noise levels can have a significant impact on measurements in these situations.

Each chapter opens with a brief literature review (expanding on the introduction above) to motivate the measurement needs and goals of the various experiments.
CHAPTER 2
EXPERIMENTAL METHODS AND DATA ANALYSIS

Throughout this dissertation a variety of statistics and quantities are used to describe the flow and other measurements (such as wind speed). The following sections provide details of how each is calculated, basic assumptions made when sampling, and confidence interval calculation.

2.1 Sample Theory and Mean Values

Environmental flows are extremely complex, the result of a variety of driving forces (e.g. density and pressure gradients) occurring over a wide range of scales. Sampling typically occurs at a regular time or space interval, resulting in a sample frequency ($f_s$) or wavenumber ($k_s$) and a discretely sampled dataset with $N$ samples of a continuous process.

It is often convenient (and necessary) to regard the sampled dataset as an ergodic sample of a stationary process over the measurement period. Various statistics (such as the mean and variance) are then used to describe the dataset. Mean values are defined for discretely sampled data as

$$\bar{u} = \frac{1}{N} \sum_{i=0}^{N} u_i \quad (2.1)$$

where $u_i$ is the $i$th discrete sample. The mean value during a sample period describes a non-varying component of the $N$ samples while the instantaneous, discrete data will capture various forcing signals, allowing resolution of processes
with frequencies (or wavenumbers) up to the Nyquist sampling criterion of \( \frac{1}{2} f_s \) or \( \frac{1}{2} k_s \) (Shannon, 1949). When interpreting mean values, the sample period or region must be considered to understand what this value represents.

Statistical convergence (i.e. how many samples are needed to adequately determine a statistic such as the mean) can be examined by calculating the statistic of interest (e.g. the mean) for a subsample of size \( M \) from a size \( N \) dataset with a sample size from \( M = 2 \ldots N \). This explicit check on convergence provides information on a minimum number of samples needed to determine a statistic. In practice, this check becomes time consuming computationally when examining large quantities of data, and provides no information without already having collected the data. Rather than perform this check explicitly for all datasets, it is assumed a properly sampled dataset is sufficiently large to represent the process through converged statistics and sampled over a short enough sample period it is minimally affected by low frequency components.

Based on prior experience in the laboratory and field (utilizing the above explicit convergence check) and estimates of important time scales in the field, standard sampling criteria are developed. Velocity datasets are typically 10 minutes in length with sample rates greater than 1 Hz, including at minimum several hundred samples and more typically thousands. Sample rates are selected based on various constraints of the instrument or technique and represent a balance between sample independence, memory/storage limitations, power consumption (field only), and processing capabilities of the instrument or technique (quantitative imaging techniques).
2.1.1 Reynolds Decomposition and Turbulence Statistics

Velocity signals may be decomposed into the sum of a mean component \( \bar{u} \) subject to the considerations above and a fluctuating (turbulent) component, \( u'(t) \)

\[
u(t) = \bar{u} + u'(t)
\]  

(2.2)

This decomposition is referred to as the Reynolds Decomposition (Davidson, 2004) and is a basic assumption needed to statistically describe turbulence.

The most basic turbulent statistic is the intensity, also referred to as a root-mean-square (RMS) value, which is equivalent to the standard deviation of a signal (normalized by \( N \) instead of \( N - 1 \))

\[
\sqrt{\frac{u^2}{N}}
\]  

(2.3)

When expressed as a variance (\( \overline{u'^2} \)) this quantity will is often referred to as a turbulent normal stress (technically \( \rho \bar{u}^2 \), where \( \rho \) is the fluid density, to achieve units of stress).

The turbulent kinetic energy (TKE) is the weighted sum of the turbulent normal stresses

\[
TKE = \frac{1}{2}(\bar{u}^2 + \bar{v}^2 + \bar{w}^2)
\]  

(2.4)

Where \( u, v, \) and \( w \) are designated as the velocities in the \( x \) (streamwise), \( y \) (cross-
stream), and $z$ (vertical) directions. Occasionally it is convenient to designate a coordinate direction using a numeric notation related to Einstein notation, but with modification summation is not implied by a repeated index. Directions are designated as 1 (streamwise), 2 (cross-stream), and 3 (vertical).

Reynolds shear stresses (as opposed to the previously defined normal stresses) are defined as

$$-ho u_i' u_j'$$

(2.5)

Where $\rho$ is the fluid density and $i$ and $j$ take on values of 1 to 3, representing the three coordinate directions ($x$, $y$ and $z$). The Reynolds stresses will often be represented using the $u, v, w$ symbols and expressed as $u'u', w'u', v'w'$. The above three statistics are commonly used to describe the strength of turbulence. They are simple to calculate and can be robustly measured in most flows with commercially available instruments.

### 2.1.2 Velocity Spectra, Structure Functions and Dissipation

The turbulent dissipation rate, $\varepsilon = \omega$, is the ultimate sink for turbulent energy in a flow where viscosity physically dissipates energy by turning it into heat. Researchers are interested in measuring $\varepsilon$ accurately because it is needed to close energy budgets. In combination with the fluid’s kinematic viscosity $\nu$ it characterizes the smallest scales of turbulence. Finally, it is a fundamental characteristic defining turbulence (Tennekes and Lumley, 1972; Davidson, 2004).
Researchers also use $\epsilon$ as a means of characterizing the strength of turbulence in specific regions of a flow or basin (Wuest and Lorke, 2003). Mathematically defined, dissipation is

$$\epsilon \equiv 2\nu s_{ij} s_{ij}$$

(2.6)

where $s_{ij} = \frac{1}{2}(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i})$ is the fluctuating rate of strain (Tennekes and Lumley, 1972). In the above equation, $i$ and $j$ take on values of 1–3 representing the three coordinate directions, and $\nu$ is the kinematic viscosity.

Measuring $\epsilon$ directly requires resolving spatial velocity gradients at the viscous or dissipative length scale $\eta \equiv (\nu^3 \epsilon)^{\frac{1}{4}}$. While possible with quantitative imaging techniques (Cowen and Monismith, 1997; Doron et al., 2001), assumptions are still needed since most imaging techniques resolve only two components of velocity at the scales needed for this direct calculation. An assumption of local isotropy for one of the horizontal gradients, typically in the lateral (2) direction, is usually made (Doron et al., 2001) to allow direct calculation of $\epsilon$.

Rather than rely on the above direct calculation and requisite difficulties in experimentally measuring the small scale gradients, several methods which rely on isotropic turbulence theory and utilize measurements at larger scales are used to estimate dissipation.

Kolmogorov’s Theory of Isotropic Turbulence (Davidson, 2004) succinctly states turbulent flows take on universal forms (at least statistically) uniquely determined by $\epsilon, r$ (a separation vector) and $\nu$. Importantly, this universal equilibrium range defined by Kolmogorov exists at larger length scales than $\eta$ which are easier to measure. More specifically, Kolmogorov’s second similarity hy-
hypothesis predicts the velocity spectrum and velocity structure function (both defined mathematically below) take on universal forms at scales intermediate between the start of the universal equilibrium range and $\eta$. These intermediate length scales are termed the inertial sub-range.

Using predictions from Kolmogorov’s second similarity hypothesis a velocity spectrum or structure function is scaled appropriately to produce an estimate of $\epsilon$ from values in the inertial subrange. While many researchers have shown local isotropy is not truly established in most flows, even at the dissipative scales (Doron et al., 2001; Saddoughi and Veeravalli, 1994), for practical work the theory still produces reasonable estimates for $\epsilon$.

The one dimensional velocity spectrum is defined

$$
\int_0^\infty E_{ii}(k_1)dk_1 = \overline{u_i'^2}
$$

(2.7)

Where $E_{ii}$ is the normalized one dimensional velocity spectrum of component $i$, $k_1$ is the wavenumber in the streamwise direction, and $\overline{u_i'^2}$ is the variance. No summation over $i$ is implied in this definition. This integral is sometimes defined to equal $\frac{1}{2}\overline{u_i'^2}$, which will alter the value of the constants $C_1$ and $C'_1$ defined below.

The one dimensional velocity spectrum is calculated using the Fourier Transform as

$$
E_{ii} = \mathcal{F}[u_i'\mathcal{F}[u_i']]^*
$$

(2.8)

Where $\mathcal{F}$ is the Fourier transform of signal $u_i'$ and a complex conjugate is de-
noted by \(^\ast\). Normalization so the integral equals the variance for temporal signals is accomplished by multiplying the above unnormalized spectrum by \(\frac{1}{f_s N}\) and for spatial signals by \(\frac{1}{N^2}\), where \(N\) denotes the length of signals used to calculate the spectrum. It is very common to divide a dataset into subrecords, compute a spectrum for each and average these sub-spectra to reduce noise.

A co-spectrum is defined similar to Eqn. 2.8, replacing the second term with \(F[u'_j]^\ast\), where \(u'_j\) is a second signal sampled at the same times as \(u'_i\) with the same number of samples. In this instance, the normalization will not change, but the created equality will be to \(\sqrt{u_i^2} \sqrt{u_j^2}\).

Finally, a power spectral density (PSD) is often utilized on non-velocity signals to examine the energy distribution as a function of frequency. In this instance, no normalization is typically performed since relative levels are of interest instead of an absolute level, although Equation 2.8 is still used for calculation of the PSD.

Kolmogorovs Second Similarity Hypothesis predicts the velocity spectrum takes on a universal form in the inertial subrange, dependent only on \(\varepsilon\) and \(k_1\) (Davidson, 2004). One of the results of this hypothesis is a prediction commonly referred to Kolmogorov’s 5/3 Law

\[
E_{11}(k_1) = C_1 \varepsilon^{2/3} k_1^{-5/3}
\]  

(2.9)

\(C_1\) is a constant equal to \(\frac{18}{55} C\) and \(C\) is Kolmogorovs constant with a value 1.5 +/- 0.1 (Saddoughi and Veeravalli, 1994). The scales at which this relationship holds are known as the inertial subrange. The 5/3 Law can be used to estimate \(\varepsilon\) in flows of sufficiently high Re number when a velocity spectrum shows the char-
acteristic $-\frac{5}{3}$ slope expected in the inertial subrange. For velocity components 2 and 3, Eqn. 2.9 utilizes a different constant, denoted $C'_1 = \frac{4}{3}C_1$. If the normalization of Equation 2.8 is set equal to $\frac{1}{2}u'^2$, the constant $C_1 = \frac{9}{55}C$ (Tennekes and Lumley, 1972).

The velocity structure function provides the same information as the velocity spectrum (the two can be related to one another, see Davidson (2004) for details). The $n$th order velocity structure function, adopting the nomenclature used in Davidson (2004) is defined

$$\langle [\Delta v(r)]^n \rangle = \langle [u_i(x + r, t) - u_i(x, t)]^n \rangle$$

(2.10)

Where $\Delta v(r)$ is a spatially separated velocity difference in the streamwise direction, $n$ takes on a power such 2, 3, etc., $r$ is the separation distance in the streamwise direction (corresponding to $k_1$), and $u_i(x, t)$ is a velocity measurement at some instant in time $t$ at position $x$.

The second order structure function will follow Kolmogorov’s 2/3 Law in the inertial subrange

$$\langle [\Delta v(r)]^2 \rangle = C_2 \varepsilon^{2/3}r^{2/3}$$

(2.11)

With $C_2$ taking on a value of 2.0 +/- 0.1 Saddoughi and Veeravalli (1994). As for the 5/3 Law, coordinate directions 2 and 3 receive a different constant, $C'_2 = \frac{4}{3}C_2$. The inertial subrange, identified in a velocity spectrum by a $-\frac{5}{3}$ slope on a log-log plot, is identified by a $\frac{2}{3}$ slope in the structure function when the radial distance is plotted on a log scale for the second order structure function.
The constants $C_1$ and $C_2$ appearing in Equations 2.9 and 2.11 are not known exactly and have been experimentally determined (e.g. see Monin and Yaglom (1975)). This adds to the uncertainty in estimates of $\epsilon$ from these two relationships.

The third order structure function will follow Kolmogorov’s 4/5 Law in the inertial subrange

$$\langle [\Delta v(r)]^3 \rangle = -\frac{4}{5} \epsilon r$$  \hspace{1cm} (2.12)

Where is constant $\frac{4}{5}$ is obtained during the derivation of the 4/5 Law from the Kármán-Howarth equation (Davidson, 2004; Antonia et al., 1997). Equation 2.12 is derived for the streamwise velocity component only and is not typically applied to the cross-stream and vertical components (e.g. see Antonia et al. (1997)). Experimental measurements show, however, the two orthogonal components produce similar estimates of $\epsilon$ using Equation 2.12 ($\S$2.1.4). This is without a correction similar to the $\frac{4}{3}$ factor seen in Equations 2.9 and 2.11 for the cross-stream and vertical components. A derivation of the equivalent of Equation 2.12 for these two components is needed to understand what corrections are needed to correctly utilize this relationship with these velocity components.

The above three quantities all rely on spatial measurements, while most velocity measurement systems are temporal. The Taylor Frozen Turbulence Hypothesis, alternatively called the Flying Hotwire Analysis, allows conversion of temporal measurements to spatial measurements (Taylor, 1938). Specifically, the radial distance $r$ is simply equal to $ut$, where $u$ is the velocity component in the $r$ direction and $t$ is a time vector beginning with $t = 0$ at the start of velocity
measurements, with $u$ typically chosen as the mean velocity. The wavenumber used in velocity spectra is estimated as $k = \frac{2\pi f}{u}$, where $f$ are the discrete frequencies where the spectrum is calculated. Taylor’s hypothesis is generally applicable when turbulence, $u'$, is much less than the mean flow $\bar{u}$. The commonly accepted limit for this ratio is $\approx 0.1$, although there are problems in very slow flows, oscillatory flows, and other situations (e.g. see Piomelli et al. (1989) for discussion of its application in wall bounded flows). For spatial measurements, no such assumption is needed and wavenumber spectra and structure functions are calculated directly from the data.

### 2.1.3 Noise and Noise Bias

As mentioned in the introductory chapter, many of the systems studied are low energy and characterized by slow flows. For the acoustic instruments used in the field and laboratory to measure velocity, this means the variance due to noise can be the same order magnitude or larger than the variance due to turbulence. The noise in these instruments has numerous sources, including electronic noise, Doppler noise due to random motions within the sample volume, velocity shear, and several other sources. For a more complete discussion of noise sources, the reader is referred to Lhermitte and Serafin (1984); Voulgaris and Trowbridge (1998) and references there.

In order to estimate the bias due to noise in velocity measurements, certain assumptions about the structure and nature of the noise are needed. For Doppler instruments discussed in §2.2.1 the common assumptions are (Lohrmann et al., 2002; Hurther and Lemmin, 2001; Blanckaert and Lemmin,
• It has a flat spectral response at all frequencies (white noise).
• It is unbiased (zero mean) but has a finite variance ($\sigma^2$).
• Its skew is zero.
• It is statistically independent of the velocity.
• It is statistically independent between independent components.

The above assumptions describe what is commonly referred to as Gaussian white noise. For particle image velocimetry or other quantitative imaging techniques, the above assumptions also apply (Cowen and Monismith, 1997).

The fourth and fifth assumptions are important when examining the influence of noise on flow statistics. The fourth assumption can be expressed mathematically as

$$\overline{(u'_i + \sigma_i)^2} = \overline{u'^2_i} + 2\overline{u'_i\sigma_i} + \overline{\sigma^2_i} = \overline{u'^2_i} + \overline{\sigma^2_i}$$  \hspace{1cm} (2.13)$$

Where the middle term of the expansion averages to zero because of the definition of $u'$ and $\sigma$ as having zero mean and being uncorrelated and $u_i$ and $\sigma_i$ are the measured velocity component and associated noise. A similar relationship using two velocity components, each with their own independent noise component yields

$$\overline{(u'_i + \sigma_i)(u'_j + \sigma_j)} = \overline{u'_i u'_j} + u'_i \overline{\sigma_j} + u'_j \overline{\sigma_i} + \overline{\sigma_i \sigma_j} = \overline{u'_i u'_j}$$  \hspace{1cm} (2.14)$$
where the last three terms in the expansion all average to zero. The above two basic relationships are important as they allow an estimate of the bias due to noise in various flow statistics. Hurter and Lemmin (2001) provides an overview of the bias inherent in common flow statistics as well as a method to estimate the value of $\sigma_i^2$. This method is essentially an Optimal (Weiner) Filter, using velocity spectra to estimate the shape of the noise spectrum and directly integrating it to obtain $\overline{\sigma_i^2}$. The results of Hurter and Lemmin (2001) show noise bias is present in variance based quantities (turbulence intensity, velocity spectra, second order structure function), absent in co-variances (Reynolds shear stresses), and absent in tri-covariances (third order structure function). Because the mean of the noise is assumed zero, mean statistics are unbiased by noise.

The above assumptions are developed for an ideal system, which in reality does not exist. For instance, Voulgaris and Trowbridge (1998) showed for a three receiver acoustic Doppler velocimeter (discussed in §2.2.1) that the Reynolds stress sees a noise bias of 2% due to alignment errors in constructing the instrument. In practice, dependent on system quality, the above assumptions are quite good at eliminating noise bias in turbulence statistics.

### 2.1.4 Self Consistency of Dissipation Estimates

Given the potentially low energy of the study systems and the need to correct dissipation estimates for bias, a small laboratory experiment was conducted in a turbulent open channel flow to validate the three dissipation calculation schemes above.
Bias due to noise in dissipation estimates is fairly straightforward to estimate. Hurther and Lemmin (2001) showed the velocity spectrum is biased by a term proportional to $\sigma^2$. Using the noise properties discussed in §2.1.3, it can be shown the second order structure function is biased by a term equal to $2\sigma^2$.

In an ideal system the third order structure function, involving cubed and uncorrelated squared noise terms, will be unbiased. It is difficult to estimate the actual bias in the third order structure function because the assumptions regarding the noise terms difficult to verify. This would require measurements capable of estimating the value of terms such as $\sigma_i^3$ or $\sigma_i^2\sigma_j$, which are unavailable. The noise spectrum generally has a frequency dependent shape (Figure 2.1) indicating it is not truly white noise (at least as estimated), suggesting these terms will have non-zero values, even if very small.

Mean flow was approximately 0.40 m s$^{-1}$ and measurements were made 0.08 m above the bottom boundary, expected to be well outside the boundary layer and largely free of wall effects which might affect measurements.

A Nortek Vectrino acoustic Doppler Velocimeter (see §2.2.1 for details) was used to measure velocities with $f_s = 200$ Hz. The Nortek Vectrino produces two, redundant but independent, estimates of the vertical velocity. These two vertical velocity estimates allow estimation of the noise spectrum using the method of Hurther and Lemmin (2001). This method utilizes the two vertical velocity estimates, which are presumed to have independent noise components, to calculate a co-spectrum. The difference between this co-spectrum and the spectra from each vertical velocity estimate is the noise spectrum. The co-spectrum, co-structure functions and the resultant dissipation estimates are presumed to be unbiased by noise, while the noise spectrum allows correction to spectra and
structure functions biased by noise.

The noise variance ($\sigma^2$) is estimated by integrating a noise spectrum obtained using the method of Hurter and Lemmin (2001). The y-intercept of the second order structure function, expected to be equal to $2\sigma^2$, was calculated by a linear least squares fit to the first five $r$ values. The ratio of these two estimates of the noise variance was 0.49, corroborating the expected bias in the velocity spectrum and second order structure functions. This check consistently produced similar values when employed in routine data analysis.
Table 2.1 provides a summary of the dissipation estimates obtained during these experiments with each method and for each component. Examples of the compensated, normalized one dimensional velocity spectrum, second and third order structure functions are shown in Figures 2.2 through 2.4. A compensated spectrum or structure function refers to solving Equation 2.9, 2.11, or 2.12 for dissipation throughout the frequency, wavenumber or separation distance \( r \) domain. Normalization refers to scaling the wavenumber or separation distance \( r \) by \( \eta \) to place the compensated spectrum or structure function in a universal form.

Table 2.1: Dissipation estimates in \( m^2 s^{-3} \) from Rusello and Cowen (2011).

<table>
<thead>
<tr>
<th></th>
<th>( u )</th>
<th>( v )</th>
<th>( w_1 )</th>
<th>( w_2 )</th>
<th>( w_1w_2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>5/3 Law</td>
<td>( 4.9\times10^{-5} )</td>
<td>( 7.28\times10^{-5} )</td>
<td>( 3.30\times10^{-5} )</td>
<td>( 3.24\times10^{-5} )</td>
<td>( 3.18\times10^{-5} )</td>
</tr>
<tr>
<td>2/3 Law</td>
<td>( 3.52\times10^{-5} )</td>
<td>( 3.63\times10^{-5} )</td>
<td>( 1.61\times10^{-5} )</td>
<td>( 1.75\times10^{-5} )</td>
<td>( 1.74\times10^{-5} )</td>
</tr>
<tr>
<td>4/5 Law</td>
<td>( 3.13\times10^{-5} )</td>
<td>( 3.19\times10^{-5} )</td>
<td>( 3.08\times10^{-5} )</td>
<td>( 3.56\times10^{-5} )</td>
<td>( 2.99\times10^{-5} )</td>
</tr>
</tbody>
</table>

The third order structure functions estimates are uniformly consistent across all velocity components (recall Equation 2.12 is not explicitly defined for the cross-stream and vertical components). This value, because of its consistency and the expected properties of the third order structure function is used as an expected value for the the spectral and second order structure function estimates. There is good agreement in the \( u \) estimate among the methods, however the \( v \) spectral estimate is significantly higher, likely contaminated by vibrations of the instrument detected in this velocity component. The second order structure function estimates for the \( w \) velocities are consistently lower than the other two methods. Both the spectral and second order structure function estimates
Figure 2.2: Compensated, normalized wavenumber spectra for each velocity component and the cospectrum of $z_1z_2$. Streamwise ($\bullet$), cross-stream ($\times$), $z_1$ (+), $z_2$ ($\Delta$), $z_1z_2$ ($\ast$). The solid line has a value of $C_1 = 0.491$ or $\frac{4}{3}C_1$ depending on the component being plotted.

yield higher $\epsilon$ values for the streamwise and cross-stream components than the vertical by a factor of 2. Because of the noise correction employed, this difference is not expected to be due to the noise characteristics of the instrument.

Correcting for noise bias will be important when measuring in low energy systems where the $\sigma^2$ term can be of the same order magnitude as the actual turbulence. While ideal agreement was not obtained between the three methods, given the simplicity of calculating each method and their reliance on the
Figure 2.3: Compensated, normalized second order structure functions for each velocity component and $z_1 z_2$. Streamwise ($\bullet$), cross-stream ($\times$), $z_1$ ($+$), $z_2$ ($\triangle$), $z_1 z_2$ ($\star$). The dashed line represents $C_2$ while the solid line is $\frac{1}{3} C_2$.

same basic data, there is little reason not to utilize each method as a simple consistency check for a dataset.
Figure 2.4: Compensated, normalized third order structure functions for
each velocity component and $z_1 z_2$. Streamwise (•), cross-stream
(x), $z_1$ (+), $z_2$ (∆), $z_1 z_2$ (∗).

2.2 Velocity Measurements

2.2.1 Acoustic Doppler Velocity Measurements

A variety of commercially available acoustic Doppler velocity measurement
systems were utilized for both field and laboratory experiments. These fall into
two main categories based on the transmit–receive path. The first type of system
is referred to as monostatic and utilizes the same transducer for transmitting
and receiving. The second type of system utilizes one transducer for transmitting and another transducer for receiving the return echo and is referred to as a bistatic system.

Many of the acoustic Doppler instruments used in this dissertation utilize a processing technique known as pulse coherent processing Lhermitte and Serafin (1984). Pulse coherent processing utilizes pulse pairs to measure a Doppler effect induced phase shift. Advantages over incoherent processing are lower noise, higher spatial resolution and higher sample rates. Pulse coherent processing does have limitations, primarily with aliasing of large velocities due to phase ambiguity, pulse interference when located near boundaries, and limited spatial range.

Monostatic Systems Used

The monostatic instruments used are commonly referred to as acoustic Doppler current profilers (ADCP) or simply current profilers. Specific commercial systems used are the Teledyne-RDI Workhorse Monitor in both 600 KHz and 1200 KHz operating frequencies and the Nortek Aquadopp High Resolution (HR) Profiler operating at 2 MHz. Both instruments feature an array of monostatic sub-systems which measure a profile of along beam velocities in individual range cells, with the acoustic beams oriented at a small angle (typically 15°-25°) from vertical. All three instruments used feature a diverging beam pattern. The Teledyne-RDI ADCPs use four beams, oriented in two orthogonal planes in what is called a Janus configuration, while the Nortek HR Profiler uses three beams in the configuration shown in Figure 2.5.
Figure 2.5: (top) Teledyne-RDI Workhorse Monitor. (bottom) Nortek Aquadopp Profiler.
An orthogonal set of basis vectors (i.e. $x,y,z$ components) formed from the beam velocities is aligned with an arbitrary axis of the instrument and referred to as XYZ or instrument coordinates. For the Teledyne-RDI ADCPs the positive $x$-axis points in the direction of beam 3. For the Nortek HR Profiler the $x$-axis is indicated by a marking on the head and aligned with beam 1. Rotating this instrument coordinate system into a fixed Earth based reference frame (East, North, Up or ENU) (Lohrmann and Hackett, 1990) simplifies averaging if the instrument is not stationary during measurements and provides a convenient reference frame for interpreting velocities.

While not primarily intended for measuring turbulence, current profilers have been shown to resolve turbulence in a variety of flows, particularly when using pulse coherent processing for velocity estimation (Lorke, 2007; Wiles et al., 2006). When measuring turbulence with a current profiler it is important to work with beam velocity data as the spatial averaging occurring in the transform to XYZ or ENU coordinates significantly reduces the variance (and thus the turbulence) in the signal. Specific techniques have been developed for estimating Reynolds stresses (Stacey et al., 1999) using a current profiler but are only applicable to systems with opposing beams like the Teledyne-RDI.

Two basic deployment configurations were used for measurements in the bottom boundary layer. A bottom frame constructed from PVC pipe was used in several deployments with the current profiler mounted facing upwards. Due to the offset to the first cell (blanking distance) and the height of the instrument itself, most of the boundary layer structure is missed in this deployment configuration.

In the second configuration the current profilers were deployed looking
downwards at the bed from a height of 1-2 m. This configuration allows near boundary velocities to be measured, but there is a region very near the bed where side-lobe interference and boundary echos bias measurements. Data obtained with either deployment method can experience problems when the flow is not horizontally homogeneous, however, this situation is more likely near the bed where local bathymetry can influence the flow. These inhomogeneities are easy to identify by examining beam velocities, which should be equal in magnitude but opposite in sign and reflect similar trends in the velocity.

**Bistatic Systems Used**

The instruments used in the bistatic class are typically referred to as acoustic Doppler velocimeters (ADV, a trademark of Sontek-YSI, Inc.). Acoustic Doppler velocimeters measure the velocity in a single sample volume of water located a fixed distance from the instrument (nominally 5 and 15 cm from the central transmitter for the instruments used).

This distance is set by the probe geometry and system timing. Sample volume diameter is related to the size of the ceramic transducer used in the probe head and divergence along the beam path (negligible for the systems used because of focused transducers and short beam paths). Sample volume height is controlled by the receive time and transmit length. While convenient to think of the sample volume as a cylinder as a first order approximation, the actual travel time surfaces form a saddle shape significantly more complex. For a detailed discussion of the sample volume including modeling of the acoustic characteristics (e.g. signal strength) the reader is referred to Zedel (2008).
Bistatic systems are well suited to turbulent velocity measurements because of the high sample rates (typically maximum rates of 50–200 Hz) and the small volume of water sampled. Numerous studies have validated their use in a variety of turbulent flows as well as providing extensive characterization of the noise and uncertainty in their measurements (Lohrmann et al., 2002; Voulgaris and Trowbridge, 1998; Hurther and Lemmin, 2001; Blanckaert and Lemmin, 2006).

The specific commercial systems used are all manufactured by Nortek AS of Norway or Nortek Scientific of Halifax, Nova Scotia, Canada. For field experiments, the Nortek Vector, a ruggedized three receiver system was used (Figure 2.6). For laboratory experiments, two versions of the Vectrino, a four receiver system with a streamlined head (Figure 2.6) were used. The first is the standard Vectrino with optional plus (+) firmware enabling high sample rates measured velocities at a single point. The second version was the Vectrino II, a profiling system based on the Vectrino probe head with new electronics and signal processing. The Vectrino II is discussed more fully in Craig et al. (2011).

**Standard Screening for Doppler Velocity Measurements**

Most acoustic instrument will report at least one of two common data quality metrics, signal-to-noise ratio (SNR) and correlation. Pulse coherent profilers (e.g. the HR Profiler) typically only report correlation while bistatic systems like the Vector will report both. All instruments will typically report a return signal strength (amplitude), however SNR is generally preferred for judging data quality.
SNR will vary depending on an instrument or manufacturer, with dependence on instrument electronic noise, transmitted signal strength, transmitted signal length, received signal strength and sample volume height. It will also further depend on whether the noise level is subtracted from the received signal during calculation of the ratio. It is reported in decibels (dB) on a logarithmic scale.

Correlation is reported as a percent value based on the phase determination calculation in pulse coherent processing.

In both cases higher values are better. Standard screening identifies measurements with low SNR (typically 5-10 dB) and low correlation (below 40%) and eliminates them from a dataset. Statistical outliers are then identified using an iterative, adaptive Gaussian filter (Cowen and Monismith, 1997), identify-
ing as outliers measurements 3–4 standard deviations (typical thresholds) away from the center of the distribution (the median on the first pass, mean on subsequent passes). Outliers are removed from the dataset with stopping criteria a maximum number of iterations, reduction of identified outliers to zero, or a constant number of outliers identified for multiple passes (e.g. one outlier on three consecutive passes). For current profilers, each range cell is treated as a time series for screening purposes.

Velocity statistics are then calculated on the screened time series. Velocity spectra require a continuous dataset for efficient calculation with the fast fourier transform, so linear interpolation is used to replace missing values or the unscreened data (if of sufficient quality prior to screening) is utilized. Comparisons of screened and unscreened statistics and histograms are used to determine whether screened or unscreened data is used for spectral calculations.

### 2.3 Quantitative Imaging

For many laboratory experiments particle image velocimetry (PIV) is utilized. A typical PIV setup consists of a camera and lens, a computer for recording images directly to hard discs, a laser and associated optics, and a timing computer controlling image exposure and illumination. Either an Argon-Ion continuous wave laser or a pulsed pair of Nd:Yag lasers is used to illuminate the image, with a thin light sheet formed by scanning the laser beam across the field of view or by spreading the beam with a cylindrical lens. All PIV applications in this dissertation use a light sheet oriented normal to the bottom boundary to capture a wall parallel and wall normal velocity component (typically the
streamwise and vertical components). A minimum of 1000 image pairs were typically collected at 1 Hz to ensure statistical convergence. An image of a ruler was taken after velocity data collection for calibration of the images from pixels to physical units.

The specific implementation of PIV used here is based on the cross-correlation of single exposed image pairs in a 2D plane (Cowen and Monismith, 1997). A variety of custom code bases are used to process the image pairs including the code developed by Cowen and Monismith (1997), a MATLAB MEX file based version of this algorithm developed by Liao and Cowen (2005) and a rewrite of the basic algorithm used in both to take advantage of parallel processing (pPIV, discussed below). A typical processing sequence consists of three passes through a dataset. The first pass is supplied with an initial uniform displacement field, while the second pass receives the mean first pass displacement field (an average over all image paris at each interrogation point) and the third pass receives the second pass instantaneous image pair velocity field.

A local median filter (Westerweel and Scarano, 2005) is used to screen each image pair displacement field. When examining mean statistics (such as creating the mean displacement field for the first pass) the adaptive Gaussian filter described in §2.2.1 is also utilized, applied at each interrogation point over all image pairs after local median filtering. For spatial spectra 2D linear interpolation in MATLAB is used to replace missing values.
Parallelized Image Processing

One of the main drawbacks of quantitative image processing techniques is the time needed to post process images for displacements. The algorithm used by Cowen and Monismith (1997) and Liao and Cowen (2005) was re-written in C to utilize parallel processing capabilities of Apple’s Mac OS X. For clarity, this implementation of the cross-correlation image analysis will be called pPIV.

The Mac OS X digital signal processing library (vDSP) is used for Fourier transforms and other numerical operations. Parallel computation is handled by the OS X Application Programming Interface Grand Central Dispatch (GCD), which is Apple’s OS X optimized implementation of the libdispatch library.

pPIV includes universal local median outlier detection Westerweel and Scarano (2005) and Shepard (1968) inverse distance weighted for handling data interpolation. Subpixel fit is handled with the three point Gaussian curve fit (Westerweel, 1997). Control of various processing parameters is handled via simple text files.

For the GCD implementation, processing an image pair from start to finish (including local median filtering and interpolation) is designated as a code block. The Grand Central Dispatch API handles a block associated with each image pair concurrently (i.e. as resources are available) using the Global Concurrent Dispatch Queue. This vastly simplifies the parallel implementation details as the OS is responsible for managing resources and not the programmer and application. Because specific resources are not allocated the number of tasks executed by the Global Concurrent Dispatch Queue and the total processing time will vary depending on the number of available cores, the amount of work
being done by other processes, and the number and priority of tasks in other serial dispatch queues.

Performance of the parallelized and serial pPIV was benchmarked by processing the Cowen and Monismith (1997) dataset on an Apple MacPro 3,1 with eight 2.8 GHz Intel Xeon processors and 16 GB of memory. The Cowen and Monismith (1997) code was compiled using the GNU FORTRAN compiler (http://gcc.gnu.org/fortran/) with no optimization flags. It should be noted the -O3 compiler flag should significantly reduce the processing time for the Cowen and Monismith (1997) FORTRAN code. The pPIV code was compiled with the default XCode 3 gcc flags for code debugging (-O0).

Images were processed with 32 x 32 pixel subwindows at 75% overlap, resulting in a 53 x 123 point interrogation grid with three total passes through the dataset. The first pass used an initial uniform shift of five pixels, with the second and third passes using a mean velocity field from the previous pass result. No refinement of the interrogation grid between passes was performed.

Code execution was timed with the UNIX utility time. Results from each processing code are presented in Table 2.2. Note that results from the Cowen and Monismith (1997) code are for one pass through the dataset and include time to process the data for the next pass which is implemented as a separate routine called pivpst. The results of each of the three passes are presented as Runs 1, 2 and 3. Results for pPIV are for all three passes and include processing time needed to prepare data for the next pass. A total processing time for the FORTRAN code would be the summation of the times from Runs 1, 2 and 3 or three times the average processing time. For comparison, a prior processing run of the data using the Liao and Cowen (2005) code is included, representing a 3
pass processing scheme but with successive refinement of the processing grid from 25% to 50% to 75% overlap.

pPIV takes the least time to process the dataset. Absolute time comparisons are problematic given the different compilers, languages and interfaces utilized for each code, however. Parallel processing shows clear benefits for any of the code bases. The reduction in processing time between serial and parallel processing for pPIV is just less than a factor of eight. This is approximately the number of available processors, but is slightly lower due to the overhead needed to direct each processor as well as resources needed by the operating system itself.

The Time Profiler tool available within Apple’s XCode development environment provides a breakdown by function of the time used during code execution. Not surprisingly, over 50% of the processing time is used by the FFT operations. The next most significant process is the image subsampling code at approximately 15%. The image subsampling time can be reduced in the future by utilizing vector based functions to subsample the image data.

Table 2.2: Time results from processing the Cowen and Monismith (1997) dataset with different processing codes.

<table>
<thead>
<tr>
<th></th>
<th>Time (mm:ss)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>pPIV (parallel)</td>
</tr>
<tr>
<td>Run 1</td>
<td>11:36</td>
</tr>
<tr>
<td>Run 2</td>
<td>11:44</td>
</tr>
<tr>
<td>Run 3</td>
<td>11:38</td>
</tr>
<tr>
<td>Average</td>
<td>11:40</td>
</tr>
</tbody>
</table>
Comparison of pixel displacements and fluctuations are presented in Figure 2.7. Differences obtained by subtracting the Cowen and Monismith (1997) result from the Liao and Cowen (2005) result and pPIV results are shown in Figure 2.8. The streamwise mean profiles show very good agreement with the mean difference from the Cowen and Monismith (1997) result, averaged vertically over the profile, of 0.038 and 0.044 pixels for the Liao and Cowen (2005) and pPIV results respectively.

This slight difference between the three code bases is due to the different quality control procedures utilized in each code and in the case of the Liao and Cowen (2005) code a different subpixel estimation method. pPIV suffers somewhat compared to the other two codes because it has no processing refinements to handle subwindow edges (discussed fully in Liao and Cowen (2005)), while both the Liao and Cowen (2005) and Cowen and Monismith (1997) codes feature various subwindow weighting functions to reduce the influence of particles at the edge of a subwindow.

Vertical displacements are similar in shape and magnitude, with the largest differences observed in the middle portion of the profile. Vertically averaged differences between the Cowen and Monismith (1997) and Liao and Cowen (2005) and pPIV results are 0.032 and 0.024 pixels respectively.

The fluctuation profiles show the largest variation, with pPIV overestimating the streamwise RMS fluctuation. While some of the difference in the three results is due to the subpixel fit utilized, Cowen and Monismith (1997) and pPIV use a three point Gaussian while Liao and Cowen (2005) utilizes a spectral subpixel fit, the quality control again will play an important part in the results here. Mean vertically averaged differences in the streamwise and vertical directions
Figure 2.7: Streamwise (left), vertical (middle) and fluctuating (right) pixel displacements produced by the three PIV codes. Cowen and Monismith (1997) $\triangle$, Liao and Cowen (2005) $\circ$, pPIV $\bullet$.

for the Liao and Cowen (2005) code are 0.112 and 0.056 pixels, for pPIV these values are 0.222 and 0.043 pixels. For both of these codes, the difference increases in the bottom 200 pixels nearest the wall. This suggests special consideration and quality control are needed in this region, regardless of the code base used.

Overall, the differences between the three codes are fairly subtle. While there are obvious benefits to pPIV in terms of processing time, the more well developed and validated Cowen and Monismith (1997) and Liao and Cowen (2005)
Figure 2.8: Streamwise (left), vertical (middle) and fluctuating (right) pixel difference between the Cowen and Monismith (1997) result and Liao and Cowen (2005) and pPIV.

code bases were used extensively during experimental work both before and after the development of pPIV.
2.4 Other Measurements

2.4.1 Temperature measurement

A variety of temperature measurement devices are used. Some are part of a larger instrument package such as a conductivity, temperature, depth (CTD) probe while others are self contained logging units for temperature or temperature and pressure only. Temperature in these devices is typically measured by a thermistor (a thermal resistor), a small resistor with a response strongly dependent on temperature. Most sensors have a response time of seconds to minutes, with exceptionally low power consumption and enough non-volatile storage for longterm deployments lasting days, months, or years.

2.4.2 Pressure measurement

Pressure sensors are found on a variety of instruments and are typically treated as an absolute measure of instrument vertical location in the water column (small offsets due to changing atmospheric pressure are neglected). Most pressure sensors utilize a silicone strain gage to measure the deflection of a small membrane (or the sensor themselves) which is calibrated to produce pressure in engineering units such as decibars. While depth will ultimately depend on the Equation of State for water and local atmospheric pressure, there is a very close 1:1 ratio between depth in meters and pressure in decibars, such that pressure measurements are interpreted directly as depths.
2.4.3 Wind Speed and Direction

Wind speed is measured by mechanical wind speed sensors, a calibrated propeller whose rotation rate is related to the wind speed. Wind direction is measured by a wind vane equipped with an electronic compass, with direction reported as the direction wind is blowing from. Direction is measured as a magnetic heading. If needed, it is corrected for local magnetic declination at the time of measurement using data obtained from the National Geodetic Data Center, a part of the National Oceanic and Atmospheric Administration of the United States. All wind measurements are reported at 15 minute intervals and represent a vector average of wind speeds during the interval.

2.4.4 Optical Back Scatter and Beam Attenuation Coefficient

Optical back scatter (OBS) was used during various experiments to characterize turbidity and suspended sediment concentration. OBS relies on an infrared light source projected into the water and receptors (photodiodes) oriented at an angle to this beam path (Downing et al., 1981). The OBS outputs an analog voltage proportional to either turbidity or suspended sediment concentration depending on how it was calibrated based on the intensity detected by the photodiodes.

Infrared radiation is strongly attenuated in clear water (more than 98% within 0.2 m at 850 nm) (Instruments, 1989), with the exact sample volume location dependent on the amount of this attenuation. The OBS gain is adjusted to match the output voltage to the data acquisition system measurement capabilities. A self logging version was utilized for field measurements, while the lab-
oratory version was sampled using a USB based analog data acquisition system sampled at 100 Hz. Gain also affects the maximum turbidity or concentration value. A too low gain will result in poor signal to noise ratio while a too high gain will saturate the sensor output.

Beam attenuation coefficient (BAC) was measured by Upstate Freshwater Institute and relies on similar optical principles as OBS. Rather than measuring backscatter strength, attenuation of a light source at 660 nm is measured. BAC has been shown to be linearly related to turbidity, with its use discussed fully in Effler et al. (2006b).

2.5 Error Analysis

All measurements presented are subject to both random errors (such as the noise discussed in §2.1.3) and potential bias in measurements. Biases discussed here are in the basic measurements and potentially inherent to the technique used. When available, manufacturer accuracy estimates are provided. When unavailable, reasonable estimates of accuracy are developed based on known properties of the technique or system.

Confidence interval widths at a 95% level are estimated using the statistical bootstrap (Efron and Tibshirani, 1993). Confidence interval widths will be denoted as \([\text{lower limit}, \text{upper limit}]\) for individual measurements. An average confidence interval width for similar data (e.g. burst data from the HR Profiler) may be cited for brevity rather than specifying exact confidence intervals for all measurements. For some statistics, such as on the second and third order structure functions, calculating a confidence interval is computationally expensive.
for the entire domain. Regions of interest, such as the inertial subrange used to
determine estimates of dissipation, will be analyzed instead.

2.5.1 Velocity Measurements

Doppler velocity systems

Manufacturer accuracy for the three versions of acoustic velocimeters used in
this dissertation, the Nortek Vector, Vectrino, and Vectrino II, are listed as 0.5%
of the measured value +/- 1 mm/s. Unless the speed of sound is specified incor-
rectly, these systems should have no bias associated with their measurements.
Bias due to noise in calculated quantities is discussed in §2.1.3.

The Nortek Aquadopp High Resoultion Profiler has a manufacturer provided
accuracy of 1% of the measured value +/- 0.5 cm/s. Again, unless there is an
error in the speed of sound, there should be no bias in its measurements. Bias
due to noise will behave the same as for the acoustic velocimeters in the previ-
ous paragraph. The Teledyne-RDI Workhorse ADCPs (both 600 KHz and 1200
KHz transmit frequencies) have manufacturer provided accuracies of 0.3% of
the measured value +/-0.3 cm/s. They are subject to the same potential bias
and noise considerations as the HR Profiler.

Particle Image Velocimetery

For extensive discussion of errors in PIV, the reader is referred to both Cowen
vides a review of the potentially largest source of error in PIV, the tendency for
displacements to be biased towards integer displacements (peak locking). Both a three point Gaussian sub-pixel fit and the continuously shifting sub-window technique of Liao and Cowen (2005) are used to reduce this potential bias. Based on results shown in Liao and Cowen (2005), a maximum total error of approximately 0.1 pixel is expected in determining particle displacements for the typical $32 \times 32$ pixel subwindows used in analysis.

A second potential significant source of error is in the calibration of images to convert pixels to physical units (e.g. millimeters). An image of a ruler is used for this step, analyzed in Matlab to determine the mean number of pixels per millimeter in an image. The rulers used in this step are marked in millimeter increments with typically 30-60 millimeters visible in an image. A minimum of 11 points in the image, coinciding with millimeter markings, are selected. These provide a minimum of 10 estimates of the number of millimeters per pixel which are then averaged for the subsequent calibration value. A typical steel ruler will have an accuracy of 0.5 mm for a 1 m length, translating to an error of $5 \times 10^{-4}$ mm in the marker intervals (the same point on each mark is selected, e.g. the top edge). Averaging 10 estimates of the number of pixels per mm determined this way, the standard deviation of the measurement will be reduced by a factor of $\frac{1}{\sqrt{10}}$, or to approximately $1.6 \times 10^{-4}$ mm. Subpixel fit to the edge location would improve the accuracy of this estimate further. For a typical calibration value of 30 pixels per mm, this corresponds to an error of 0.0005%. Typical timing errors are on the order of $\mu$s for the systems used to control image illumination, representing a minimal contribution to overall error.
2.5.2 Other Measurements

Temperature and Pressure

For the SBE-39 temperature loggers, manufacturer specified accuracy is 0.002°C. This is expected to increase over time, and given these sensors are several years old, accuracy is expected to be worse than specified, increasing by 0.0002°C per month. Temperature measurements from on board the Doppler velocity instruments will have significantly worse accuracy, typically $\mathcal{O}(0.2°C)$.

Pressure measurements will have a typical accuracy of 0.1% of the full scale range, or approximately 0.02 m for the typical 20 m range of most sensors used here.

Wind Speed

Manufacturer stated accuracy is $\pm 0.3 \text{ m s}^{-1}$ or 1% of measured value with a minimal detection value of 1.1 m s$^{-1}$.

Optical Back Scatter

When calibrated for turbidity, the optical back scatter sensors will have an accuracy of 2% of the measured value or 0.5 NTU. For suspended sediment concentration of muds accuracy is 2% or 1 mg L$^{-1}$. 
CHAPTER 3
FIELD MEASUREMENTS IN THE BOTTOM BOUNDARY LAYER

As mentioned in Chapter 1, the sediment-water interface is an important region in aquatic systems from chemical, biological and physical perspectives. The interaction of the bottom boundary layer (BBL), defined by Wuest and Lorke (2003) as the region above the sediment where the boundary has an effect on velocity, with the benthic environment has important consequences for habitat use for feeding, shelter, and reproduction (Nowell and Jumars, 1984). Turbulence in the BBL will govern mass transport of compounds such as dissolved oxygen (Lorke et al., 2003a), further influencing benthic ecology by increasing the sediment oxygen demand and creating anoxic conditions unsuitable for many organisms (Effler, 1996). The transport and retention of particles at the bed is important from a management, health, and legal perspective for surface waters utilized for drinking water supply (Owens et al., 2011; Gelda et al., 2009) where these particles influence water quality, affecting both ecology and suitability for use as drinking water sources. High turbidity caused by large sediment mass fluxes associated with elevated discharge can also influence ecology by limiting light penetration, promoting the growth of one species at the expense of another (Pick, 1991; Middelboe and Markager, 1997; Strayer et al., 1999).

Common to each of the above problems is a need to understand the fluid dynamics, in particular turbulent flow, in these systems to address transport rates and the fate of compounds within a basin. Because bottom sediments are the ultimate repository for a significant portion of mass entering a basin, the fluid mechanics in the BBL are particularly important as the flow here will govern where compounds are concentrated and their eventual fate. The flow above
the boundary is expected to be turbulent in most environments (Tennekes and Lumley, 1972), driving transport rates of compounds such as dissolved oxygen. Characterizing the flow as either hydraulically smooth or rough is also important, as diffusive transport above a rough boundary can be significantly higher than over a hydraulically smooth surface (Raupach and Antonia, 1991).

The transport and mobility of bottom sediments, in particular the small, turbidity causing particles (Peng et al., 2009) needs examination from two perspectives. One is the initial transport of particles into the basin and the subsequent fate of these particles once they enter the basin. This transport mechanism will be governed by discharge events in basin tributaries and the mean flows within the basin (Gelda et al., 2009). For particles which encounter the bed, mobility will be governed by the various tractive forces maintaining the particle at rest and the ability of the flow to counteract those forces with drag (Shields, 1936). Complicating the mobility of fine sediments are the inter-particle forces which govern small particle behavior and create strongly cohesive beds with potentially significant resistance to mobilization and erosion (Winterwerp, 2004).

Field measurements were conducted in three basins, two affected by severe turbidity problems and one heavily impacted by pollution. Measurements of the bottom boundary layer flow, including mean and turbulent quantities, were made in all three to provide information used in the development of a turbulence chamber for scalar fluxes (Chapter 4) and to examine the mobility and erosion of cohesive sediments in the laboratory (Chapter 5). More general full water column measurements, along with boat based surveys described in Chapter 6, are used to examine the signature of large discharge events and the initial transport into a basin and the fate of particles in the water column.
3.1 Small Scale Hydrodynamics in Lakes

All of the measurements were made during the stratified season lasting from early summer through late fall. The vertical structure of the water column is set by density stratification controlled by temperature (salinity will play only a minor role) arising from a net positive surface heat flux into the basin. The surface mixed layer is the region from the water’s free surface to the thermocline. The thermocline is a region where temperature (and thus density) changes sharply between the warmer surface waters and cooler, deeper waters (the hypolimnion). The surface mixed layer is characterized by strong turbulent mixing as a result of wind, waves, and strong thermal convection due to surface heat fluxes. An extremely dynamic area it will often be well mixed and homogenous in physical and chemical properties. The hypolimnion is generally calm given its isolation from outside forces such as wind and waves, while the BBL is a region of enhanced mixing due to the presence of a physical boundary generating a mean shear and turbulence (Wuest and Lorke, 2003). This stratification structure is typically approximated using a simple two layer model with two densities. For deeper basins, a third layer can form, called the metalimnion, with a different density than the surface mixed layer or hypolimnion.

Wuest and Lorke (2003) suggest the surface mixed layer and the BBL are regions of enhanced mixing relative to the quiet interior regions and can be characterized (and potentially physically identified) by their turbulence levels. They utilize the turbulent dissipation rate ($\epsilon$) to characterize these regions from a hydrodynamic perspective, with typical values in the interior of $O(10^{-9} \text{ m}^2 \text{ s}^{-3})$ and in the BBL of $O(10^{-9} - 10^{-8} \text{ m}^2 \text{ s}^{-3})$. 
Most mechanical energy in the system is contained within basin scale seiches (internal waves) with periods of several hours (Wuest and Lorke, 2003). These internal waves travel along the theromcline (i.e. they can be more properly called interfacial internal waves) and are set up by a surface wind stress moving water to one end of a basin, depressing the thermocline and building up potential energy at the other. When the system releases this energy, either due to a reduction in surface winds or because the wind has blown long enough, the thermocline responds by oscillating, generating standing internal waves. Most energy winds up concentrated in natural standing waves with an integer number of nodes (i.e. 1, 2, 3, etc.) in the horizontal and vertical. The period of these natural oscillations is a function of the basin length, layer depths, and layer densities. Most energy winds up concentrated in the lowest modal response with one node in the horizontal and vertical (Wiegand and Chamberlain, 1987). A simplistic model used to predict seiche frequencies assumes a simple rectangular basin with two layers (Fischer, 1979). Higher mode waves have also been observed (Wiegand and Chamberlain, 1987), largely as a result of continuous stratification (i.e. density is now a function of depth rather than assumed constant for a layer) or from the formation of three distinct density layers, permitting a second (and higher) vertical mode to develop.

The BBL flow, because of the large amount of energy contained in internal seiches, will show a strong correlation with the these waves, oscillating at the dominant seiche period. The BBL is characterized by fairly weak mean flows, periods of zero velocity corresponding to the maximum thermocline displacement, and direction changes as the thermocline oscillates between being elevated and depressed from a mean level. The several hours long period of typical basin scale waves is expected to permit a steady flow to develop, allowing
analysis using the canonical turbulent flat plate boundary layer.

Lorke et al. (2002), showed for the 24 hour long seiche period in Lake Alpnach, the BBL flow was modified by the oscillatory nature of the forcing. Mean velocity profiles were fit to a logarithmic velocity profile (specifically Equation 3.3 discussed in §3.4.4), but the results were found to be inconsistent in producing reasonable estimates of the friction velocity \( u_* = \sqrt{\frac{\tau_{bed}}{\rho}} \) and a characteristic roughness length for the sediment surface. Using a \( k - \epsilon \) turbulence model incorporating an oscillatory forcing (Stokes Second Problem), they were able to reproduce their observations of velocity and turbulence.

However, the inconsistent results they obtained using a unidirectional flow model (the Law of the Wall, discussed further when presenting mean velocity profiles in §3.4.4) were located well away from the boundary (greater than 0.5 m) and thus well outside the region where this model is expected to be valid. Their results (see their Figure 14) showed no difference between the unidirectional and oscillatory models in the near wall region (i.e. less than 0.5 m). Because the near wall region is of primary interest in the processes under study, a unidirectional model is used to interpret results and expected to be valid.

Lorke (2007) reports observations of turbulence in Lake Constance, a fairly large basin in central Europe. A high resolution ADCP and acoustic velocimeter were used to measure mean currents \( O(0.05-0.10 \text{ m s}^{-1}) \) and estimate \( \epsilon \) as \( O(10^{-8}-10^{-7} \text{ m}^2 \text{ s}^{-3}) \) using Taylor’s Frozen Turbulence Hypothesis and Equation 2.9 on velocity time series data. Lorke (2007) showed the magnitude of \( \epsilon \) is directly tied to basin scale internal waves, in turn tying boundary mixing to the interaction of internal waves and the sloping boundary.
Numerous other studies of BBL dynamics in various sized lakes are available (Marti and Imberger, 2006; Lorke et al., 2003a; Bryant et al., 2010), but all generally feature similar conclusions. Baroclinic forcings (with or without rotational effects) contain a significant portion of the overall energy in the basin. These baroclinic motions exert significant influence over the mean flows, which in turn control the production and dissipation of turbulent energy.

Field experiments utilizing ADCPs, ADVs, and temperature loggers, all deployed with the assistance of Upstate Freshwater Institute (UFI) in Syracuse, NY were conducted in the 2004 and 2007 stratified seasons in three basins. These three basins represent three different systems, a medium sized dimictic lake (Onondaga Lake), a small volume relative to its watershed and thus highly dynamic reservoir (Schoharie Reservoir), and a larger, less dynamic reservoir (Ashokan Reservoir). Measurements were made with the express goal of characterizing the BBL through measurements of temperature to determine seiche periods and velocity measurements in the near wall region to characterize mean flow and turbulence.

3.2 Description of Field Sites

3.2.1 Onondaga Lake

Onondaga Lake is a medium sized urban lake located northwest of downtown Syracuse, NY (Effler, 1996). The basin consists of two lobes joined by a slightly shallower saddle region. It is approximately 10 km long and 2 km wide. Each of the lobes has a relatively flat expanse with a mean depth of approximately 20
m, while the saddle region has a mean depth of approximately 17 m. The lake is termed dimictic, with a summer stratified season, a fall overturn, and a well mixed, unstratified water column in winter. The lake is potentially ice covered in winter.

Onondaga Lake has been referred to as the most polluted lake in North America (Effler and Hennigan, 1996). The production of soda ash in Solvay on the south shore of the lake resulted in large amounts of ionic pollution entering the lake. The waste water effluent from Syracuse and surrounding communities also resulted in high nutrient loads which have recently been reduced. The pollutant of primary concern to the current work, however, is mercury.

Starting in 1946 and continuing through 1972, approximately 70 metric tons of mercury were dumped into the lake as a byproduct of chlorine gas production (Effler and Hennigan, 1996). This has resulted in the entire lake bed being declared a Superfund site. Because of high nutrient loading and stratification, anoxic conditions persist at the sediment-water interface for much of the stratified season. While the low dissolved oxygen (DO) levels are ecologically undesirable, oxidation-reduction reactions involving sulfur which occur in the low DO conditions also provide a preferential pathway for mercury compounds to move out of the sediment and enter the water column. Recent remediation work has focused on addressing the low DO levels through electron acceptor augmentation (Auer et al., 2010).
3.2.2 Schoharie Reservoir

Schoharie Reservoir is the most upstream reservoir in the Catskill watershed for New York City. The reservoir was completed in 1924 by damming Schoharie Creek with the Gilboa Dam. It is located partially in Schoharie, Delaware, and Greene counties in western New York. The nearest town is Prattsville and it is approximately 36 miles southwest of Albany. An aerial photo of the reservoir is shown in Figure 3.1.

At full capacity of 17.6 billion gallons of water, the reservoir basin is 8 km long and 1 km wide. The mean depth is 17 m, with a maximum depth of 42 m (Owens et al., 2011). Bathymetry begins fairly shallow at the southern end and deepens to the north, with the greatest depths just behind the Gilboa Dam. Schoharie Creek enters the reservoir at the southern end and is the main tributary, draining 75% of the watershed. Several smaller streams (or kills) enter at various points around the basin.

The Schoharie Creek thalweg can be seen in Figure 3.2 near the creek mouth as it enters the main reservoir body. Water from the reservoir is discharged into the Shandaken Tunnel. This water eventually enters Esopus Creek and flows into Ashokan Reservoir. The tunnel intake structure is located in front of the building visible bottom center in Figure 3.2. Schoharie Reservoir is subject to periods of extreme drawdown, exacerbating the elevated turbidity levels motivating measurements described here. A more thorough treatment of the turbidity problems in Schoharie Reservoir is available in Owens et al. (2011) and references therein.
Figure 3.1: An aerial view Schoharie Reservoir. The Gilboa Dam is visible at the top of the image. Schoharie Creek is visible at the bottom of the image.
Figure 3.2: An aerial photo of the southern end (looking south) of Schoharie Reservoir during extreme drawdown conditions. Schoharie Creek enters at top right and follows the reservoir thalweg west then north. The basin width is approximately 500 m. Distance from the Schoharie Creek dam to the abrupt turn north is approximately 800 m.
3.2.3 Ashokan Reservoir

Ashokan Reservoir, located in Ulster County, NY, approximately 13 miles west of Kingston, NY, is a New York City Water Supply System reservoir. It was completed in 1915 and is one of the largest reservoirs in the New York City water supply system. The reservoir was formed by damming Esopus Creek and flooding a valley where several small villages were located. It is comprised of two basins, the East and West basin. The West basin receives water from Esopus Creek and discharges through an underflow dam to the East basin. The East basin receives water from the West basin and discharges into Esopus Creek and the Catskill Aqueduct leading to the Kensico Reservoir.

Figure 3.3 shows the bathymetry of the basin, as collected by Upstate Freshwater Institute and GZA GeoEnvironmental, Inc. Bathymetry is based on depth below an elevation of 178.3 m representing a 100% full reservoir. The Western Basin is approximately 7 km long and 2 km wide with an average depth of 14 m and a maximum depth of 52 m. Its principal axis runs northwest to southeast. The Eastern Basin is approximately 11 km long and 2 km wide with an average depth of 15 m and a maximum depth of 27 m (dependent on reservoir level). Its principal axis runs east-northeast to west-southwest.

Along with Schoharie Reservoir, the West basin experiences periods of elevated turbidity which routinely exceed the EPA limits for surface waters utilized for drinking water supply. Details on the West basin turbidity problems are available in Gelda et al. (2009); Effler et al. (1998, 2002). Measurements were made only in the West basin where turbidity problems primarily occur.
Figure 3.3: The West basin of Ashokan reservoir with 5 m depth contours. Deployment locations are marked by ●.

3.3 Methods

Two fundamental physical processes are of interest for the above systems, sediment resuspension and transport (Ashokan and Schoharie Reservoirs) and benthic mixing rates (Onondaga Lake). Turbulence plays an important role in both of these processes, influencing particle behavior, bed stress and mass transfer rates. Turbulence at the sediment water interface is set primarily by the overlying flow.
Mean shear in the velocity profile generates turbulence in the boundary layer (Davidson, 2004), which is organized into a variety of structures (Robinson, 1991). These structures and their potential influence on sediment transport will be discussed further in Chapter 5. While these structures are suspected of playing an important (if not primary) role in turbulent mass transfer (O'Connor and Hondzo, 2008), resolving them in field measurements is extremely difficult. An understanding of the dynamics in the turbulent boundary layer can be gained, however, by measuring vertical profiles of velocity and calculating various other quantities (primarily related to turbulence). Comparisons of these measured and calculated profiles to both non-dimensional scalings (Ligrani, 1988) and direct numerical simulation results (Spalart, 1988) provides insight into how well a boundary layer model will capture the flow physics.

Surface wind speed and direction were typically measured on each basin, however equipment limitations occasionally did not permit this. Temperature profiles from point sensors or a vertical profiler were used to examine stratification, determine the thermocline location, and examine the response of the system to wind. Velocity measurements generally included a profile from a high resolution current profiler and at least one near bed point measurement for turbulence. Velocities were examined for oscillatory motion, direction, and if applicable a boundary layer analysis. Turbulence was characterized statistically, compared to the mean flow magnitude, and used to estimate \( \epsilon \). Reynolds stresses were calculated when available, as an indicator of mixing strength and bed stress.

Field experiments were conducted from mid-summer (July) through the fall (October), encompassing a range of environmental conditions and partially cap-
turing a significant discharge event in Schoharie Reservoir. In Schoharie and Onondaga Lake a strong cooling of the surface mixed layer during the second half of October 2007 was captured. While each basin will have a specific forced response to the wind forcing at the surface primarily dependent on stratification and bathymetry, general characteristics of a medium sized basin subject to wind forcing can be made from these observations.

3.4 Results

3.4.1 Onondaga Lake

Measurements were carried out over a one month period during October 2007 at two locations, with approximately one week of data obtained at each location. The first location sampled from September 27 until October 11 was located on a saddle between the two main lobes of Onondaga Lake and will be referred to as the Saddle deployment. Local water depth was 17 m. The second location was sampled from October 11–31. It was located in the South Basin on a broad plain of approximately 20 m local water depth and will be referred to as the South Deep deployment.

Meteorological data for both deployments was obtained from a Remote Underwater Sampling Station (RUSS) approximately 50 m from the second deployment location. This platform was maintained by UFI. A string of Seabird SBE39 temperature and temperature/pressure loggers was deployed near the RUSS to supplement the water column profiles measured every six hours. Deployment depths were approximately 2, 4, 6, 8, 9, 10, 11, 12, 14, 16, 18, and 19 meters. A
map of Onondaga Lake showing bathymetric contours and the sample locations is shown in Figure 3.4.

Figure 3.4: Satellite image of Onondaga Lake near Syracuse, NY. Depth contours in meters are overlaid on the image. Deployment locations are marked by .

For the first deployment, three Nortek Vector acoustic Doppler velocimeters, approximately 0.20, 0.30 and 0.80 m from the sediment-water interface, were mounted on an aluminum frame. These instruments recorded data in hourly bursts lasting 10 minutes. For the highest elevation Vector, the sample rate was 4 Hz for the Saddle deployment and 8 Hz for the South Deep deployment. Each burst contained 2560 or 5120 samples. The lower two Vectors sampled at 16 Hz with 10,240 samples per burst. An operator error during data download resulted in the loss of the lowest Vector data for the Saddle deployment.
A 2 MHz Nortek Aquadopp High Resolution (HR) Profiler was added to the frame for the second deployment. The down looking HR Profiler was mounted approximately 1 m above the bed, resulting in a 0.90 m velocity profile, resolved into fifty, 0.02 m measurement cells. Data was collected in hourly bursts at 2 Hz over 12.8 minutes, 1536 profiles per burst.

The SBE39 temperature loggers sampled every 15 seconds, while the temperature/pressure loggers sampled every 30 seconds. Meteorological data from the RUSS was reported at 15 minute intervals. Wind speeds were measured with an R. M. Young mechanical wind monitor approximately 2 m above the water surface.

**Meteorological Conditions and Stratification**

Moderate winds were present during the month long deployment. Maximum wind speeds were approximately 11 m s\(^{-1}\), but the mean wind speed for the deployment was 3.3 m s\(^{-1}\) (Figure 3.5). The strongest winds persisted for only a few hours, however direction remained relatively constant during the growth and decay of these events. While wind direction varied throughout the deployment (Figure 3.6), consistent winds came from the west (fronts moving down from Canada and across the Great Lakes) and the southeast (fronts moving inland from the Atlantic seaboard). The strongest winds, however, came from the south or north-west. A diurnal variation in wind speed occurs throughout the deployment period.

Starting on Day 12 and lasting for three days a wind event with mean wind speed 4.8 m s\(^{-1}\) occurs. Wind direction during this period is consistently out
Figure 3.5: Wind speed (top) and direction (bottom) measured at the Onondaga Lake RUSS.

of the west, closely aligned with Onondaga Lake’s long axis. As will be seen in the velocity measurement results, this sustained wind event is as important for understanding the physics of Onondaga Lake as the stronger (but shorter) events on Days 3 and 19.

The SBE39 temperature data was interpolated onto a 1 m vertical resolution grid using a linear, nearest neighbor interpolation scheme based on Delauney triangulation and resampled to match the meteorological data time stamps. The resulting temperature profiles are shown in Figure 3.7. The thermocline depth,
defined as the location of the maximum gradient in temperature \( \left( \frac{dT}{dz} \right) \), is determined by numerically differentiating using central differences, the temperature profiles. It was approximately 12 m for a majority of the deployment period (days 1–28). This is much greater than half the mean water depth (16 m) and close to the BBL measurement location depths (approximately 4 and 7 m deeper respectively). The potential for stratification effects to influence the structure of the BBL is therefore a possibility.

Internal waves on the thermocline can be seen in Figure 3.7 with amplitudes of a few meters. Given the shallow water depth and the proximity of the ther-
mocline to the sediment-water interface, the internal seiche is expected to play a strong role in BBL physics.

Figure 3.7: Temperature profiles during October 2007 in Onondaga Lake. The surface temperature loggers stopped recording earlier, resulting in a shorter time series at the surface.

The surface mixed layer gradually cools during the deployment, starting near 20°C, before quickly cooling from days 12–15 to 17°C. It then cools more slowly to 15°C (Figure 3.8). The rapid cooling from days 12–15 is coincides with the sustained wind event over this same time period. Deeper effects of this wind event are seen in Figure 3.7 as a loss of sharpness (i.e. mixing) in the thermocline after Day 15 as the temperature gradient becomes weaker.
The hypolimnion temperature remains relatively constant around 10°C throughout the deployment. Because of the surface mixed layer cooling and change in density, the lake’s response to wind will be slightly different from days 1–10 and from days 15–30. Using a simple two layer model, in a rectangular basin Fischer (1979) predicts the baroclinic seiche period \( T \) will be

\[
T = 2 \times \frac{L}{c_i}
\]  

(3.1)

Where \( L \) is the surface fetch and \( c_i \) is the internal wave speed, defined as
\sqrt{g' \frac{h_1 h_3}{h_1 + h_2}}. In the definition of \( c \), \( h_1 \) and \( h_2 \) are the upper and lower layer depths (divided at the thermocline) and \( g' \) is reduced gravity equal to \( g \frac{\rho_2 - \rho_1}{\rho_2} \) with \( \rho_1 \) and \( \rho_2 \) the upper and lower layer water densities. Density is calculated from temperature measurements using the UNESCO Equation of State assuming zero salinity (Fofonoff and Millard Jr., 1983).

From days 1–12.5, using \( T_1 = 19.0^\circ \text{C}, T_2 = 10.4^\circ \text{C}, h_1 = 12 \text{ m}, \) and \( h_2 = 4 \text{ m} \) (set by the previous definition of thermocline location as the maximum value of \( \frac{\partial T}{\partial z} \)) the internal seiche period is 21.5 hours. There is no consistent definition for the thermocline location in the literature, and an alternate definition using the inflection point in the temperature gradient \( \frac{\partial T}{\partial z} \), results in \( h_1 = 10 \text{ m} \) and \( h_2 = 6 \text{ m} \). Using a basin length of \( L = 7.5 \text{ km} \) (Effler et al., 2004) results in an internal seiche period of 19.3 hours. This two hour difference, approximately a 10% change, defines a range of seiche periods which may be observed in the field, in addition to the uncertainty from using the simple two layer model.

For days 15–28, using \( T_1 = 16.7^\circ \text{C}, T_2 = 9.9^\circ \text{C}, h_1 = 12 \text{ m}, \) and \( h_2 = 4 \text{ m} \) the internal seiche period is 26 hours. Using the alternate definition for thermocline location, \( h_1 = 10 \text{ m}, \) and \( h_2 = 6 \text{ m}, \) the internal seiche period is 23 hours.

Effler et al. (2004) found good agreement between observed seiche periods and the two layer model for early September conditions in Onondaga Lake. Based on the above calculations, internal seiching in Onondaga Lake during the deployment period is expected to occur with a period between 20–26 hours.

Temperature spectra from SBE39s located in the surface mixed layer (9 m depth) and at the thermocline (11 m depth) (Figure 3.9) support the estimates for baroclinic seiche periods above. For the first half of the deployment, there is
a broad peak near a period of 20 hours. For the second half of the deployment the peak shifts to a slightly lower frequency as expected. During the second half of the deployment, the seiche signal disappears from the 9 m thermistor as the thermocline moves deeper and erodes. This absence of seiche signal is confirmed by examining a time series of the its data (Figure 3.10).

Figure 3.9: Temperature spectra from 9 m (thin line) and 11 m (thick line) for Days 1-12.5 (top) and Days 15-28 (bottom). The red box represents the range of expected internal seiche periods based on the two layer model and varying layer depth definitions.

The four strong wind events (approximately days 2, 19, 23, and 28) generate a strong, almost immediate response in the thermocline displacement. Examining the cross-correlation function between the wind speed and 11 m tempera-
ture time series, there is a lag of approximately 5 hours between the two signals (Figure 3.11). Using the internal wave speed estimate developed for the two layer model, $c_i \approx 0.2\ \text{m s}^{-1}$, this observed lag is very close to the amount of time a wave formed at the center of the lake would take to propagate to either end of the basin along the long axis, confirming the predictions of the two layer model and its applicability to this system. A negative correlation peak at approximately 15 hours is likely a harmonic of this initial peak.

The observed wind events generate thermocline displacements of 2-3 m. The interface slope (and associated shear) during these displacement could generate
instabilities, which subsequently collapse and generate vertical mixing. This mixing would enhance thermocline erosion, in turn leading to more mixing as stratification weakens (i.e. the temperature difference between the surface mixed layer and the hypolimnion decreases) and the lake becomes less stably stratified. Depending on the proximity of the thermocline to the bottom boundary, this could be an additional source of turbulence and mixing in the BBL.
3.4.2 Hourly average velocity data - Saddle

The instrument frame was deployed September 27 at the Saddle measurement location in approximately 16 meters local water depth. The instrument frame as recovered at the end of the second deployment is shown in Figure 3.12

![Figure 3.12: The Onondaga Lake bottom frame. The three Nortek Vectors were deployed at both locations, while the HR Profiler (visible on the top horizontal bar) was only utilized during the second deployment.](image)

Storms and personnel availability delayed the deployment of the SBE strings until October 1, so detailed temperature data is unavailable for the first three days of this deployment. Day numbers are referenced to October 1, resulting in negative day numbers at the start of the velocity data time series, with Day -2 corresponding to Sept. 27. Because of low power, the middle elevation Vector began to cycle bursts more quickly than the programed one hour interval, beginning around day zero, which also affected data quality as fewer samples
were available for averaging. The lowest elevation Vector data was lost due to operator error, also resulting in the loss of heading information to reference the velocity data to an Earth reference frame.

Figure 3.13 presents burst averaged magnitude and direction data for the middle and highest elevation Vectors during the Saddle deployment. The magnitude has been assigned a sign by utilizing a four quadrant arctangent (i.e. the function \texttt{atan2} in most computer languages, which interprets angles as clockwise or counterclockwise based on the sign of components) and the measured horizontal velocity vectors.

Mean flows at the Saddle are 0.035 m s$^{-1}$ (mean confidence interval width of 4 x 10$^{-4}$ m s$^{-1}$ for all valid bursts), with maximum values over 0.10 m s$^{-1}$. Magnitude falls to zero during changes in direction lasting 1–2 hours. Direction is presumed to be along the main axis of the lake with a 180$^\circ$ change in direction during flow reversals. The 180$^\circ$ change in flow direction is due to the strong influence of the baroclinic seiche on the BBL.

Correlations between the wind, burst averaged velocity at the highest Vector and temperature at 11 m are shown in Figure 3.14. Temperature and wind records were resampled by pulling the sample closest to the burst time from the records shown in Figures 3.5 and 3.7.

There is a very weak correlation between the wind magnitude and water velocity at the Saddle. A much stronger correlation exists between temperature and velocity magnitude. This strong correlation is expected based on the two layer model and conservation of mass, as water in the hypolimnion must move to the opposite end of the basin as the surface water moves to the other (pushed
Figure 3.13: Burst averaged magnitude and direction from the high Vector (●) and middle Vector (△) for the Saddle deployment.

In addition to the strong correlation at zero lag between the temperature and water velocity signals, there is a second peak at a lag of 8–9 hours. This peak is due to the horizontal distance between the Saddle velocity measurements and the temperature measurement location and the time it takes information to propagate between the two locations (i.e. the internal wave speed). This lag is approximately twice the lag observed between the wind and temperature, accounting for a return journey of the wave to the center of the lake.
Figure 3.14: Normalized cross-correlation between the wind speed (●) or 11 m temperature (△) and $U$ at the high Vector.

Spectra of the burst averaged magnitude and direction are presented in Figure 3.15. There is a broad peak in the same frequency range as seen in the temperature spectra shown in Figure 3.9. This again fits with the physics of the two layer seiche model.
Figure 3.15: Burst magnitude (●) and direction (△) from the high Vector. The rectangle shows the expected seiche frequency range.

### 3.4.3 Hourly average velocity data - South Deep

The instrument frame was redeployed in the South Deep location on day 11. This deployment also includes data from a Nortek Aquadopp HR Profiler. Presented here are results from the three Nortek Vectors for comparison to the Saddle dataset.

Figure 3.16 presents burst averaged magnitude and direction data for the three Vectors during the South Deep deployment. By utilizing the heading data
from the HR Profiler, the direction has been referenced to North at 0°, with measured directions corrected for local declination of -12.6°. The lowest elevation Vector data quality was significantly poorer than the middle and highest elevation instruments and is generally only suitable for mean velocity estimates, and entirely eliminated beyond Day 18.

![Graph showing velocity magnitude and direction](image.png)

**Figure 3.16:** Burst averaged magnitude and direction from the high Vector (●) and middle Vector (△), and low Vector (*) for the South Deep deployment.

Mean flows are much lower than at the Saddle, rarely exceeding 0.03 m s⁻¹ and a mean value at the highest elevation on only 0.015 m s⁻¹. Mean confidence interval width is 4.0 x 10⁻⁴ m s⁻¹ for all valid bursts. Maximum velocities
are only 0.05 m s⁻¹. Near zero magnitudes occur frequently during changes in direction of approximately 180° degrees. Flow is oriented nominally along the long axis of the lake which runs northwest-southeast. Similar to the Saddle data, there is clear evidence of the baroclinic seiche driving the mean flow.

Because of the slightly greater water depth and larger cross sectional area, magnitudes are lower than at the Saddle. The location of the measurements away from the expected Mode 1 internal wave node location near the middle of the lake (close to the Saddle deployment position) also results in a decrease in the mean flow strength.

### 3.4.4 Hourly average velocity profiles - South Deep

With the addition of the Nortek HR Profiler to the instrument frame for the South Deep deployment, a powerful dataset was added for characterizing the BBL. The HR Profiler is specifically suited to measurements in the bottom boundary layer and to conditions encountered at the South Deep location such as very slow flows and a low scattering environment.

The instrument frame was oriented almost due east with a heading of 90.6°, corrected for a local declination of 12.6° West of North. This aligned the X and Y components of the HR Profiler with East and North respectively. Similar to the Vector data, the horizontal X and Y components are used to calculate a burst average magnitude in each range cell, creating a mean velocity profile. Measured flow is predominantly along the lake’s main axis running northwest-southeast. Psuedo-color plots of velocity magnitude and direction are shown in Figure 3.17.
Figure 3.17: Burst average magnitude (top) and direction (bottom) measured by the HR Profiler during the South Deep deployment.

Magnitudes rarely exceed 0.01-0.02 m s$^{-1}$ while mean confidence interval widths are $4 \times 10^{-4}$ m s$^{-1}$. There is again a strong periodicity in the data, established as the baroclinic seiche from an examination of the Vector burst average data. While difficult to pick out in Figure 3.17, a mean shear exists in many profiles away from the boundary, outside the region where the no slip condition is expected to generate shear. Shear away from the boundary is due to other forcing, for instance a lag in response during direction changes. Direction is generally uniform throughout the profile. On those occasions when there is directional shear in the profile, such as Day 15.5, it does not persist for long. More
data is needed to fully characterize the source of this shear given the numerous potential sources for it.

Near the bottom boundary, a turbulent boundary layer flow is expected. Outside of the viscous region very near the wall, the mean velocity is expected to follow the Law of the Wall

\[
\frac{\bar{U}}{u_*} = \kappa^{-1} \log \frac{z}{\kappa u_*} + \beta
\]  

(3.2)

with \( \kappa = 0.41 \) (von Karman’s constant) and \( \beta \approx 5.0 - 5.5 \). Here \( \beta \) is assigned a value of 5.0. The LHS of Equation 3.2 and the argument to the natural log are often replaced with the following symbols, \( u^+ = \frac{U}{u_*} \) and \( z^+ = \frac{z}{\kappa u_*} \). This normalization is used throughout the boundary layer analysis, with velocities normalized by \( u_* \) and lengths by the viscous length scale \( \frac{\nu}{u_*} \).

Modifications to Equation 3.2 are often encountered to account for wake effects in the outer flow region (Raupach and Antonia, 1991). Equation 3.2 is modified to allow for roughness effects by the inclusion of an additional term, \( \frac{\Delta U}{u_*} \), subtracted from the RHS. This term is called the roughness function and represents the offset from a smooth wall velocity profile. Raupach and Antonia (1991) provides extensive discussion of \( \frac{\Delta U}{u_*} \) and its relationship to the physical roughness height and \( u_* \). As roughness effects increase, parametrized through the roughness Reynolds number \( h^+ = \frac{h u_*}{\nu} \) where \( h \) is the roughness height, the roughness function should increase with a log dependence. Typical laboratory values of \( \frac{\Delta U}{u_*} \) are O(1-10) with fully rough atmospheric flows generated values O(20). There are numerous other forms of Equation 3.2 which account for roughness effects (Raupach and Antonia, 1991). Given each form is intended
to capture the same physics, relationships between the various forms do exist. Equation 3.2 with $\frac{AU}{u_*}$ is utilized here because of its simplicity and the straightforward collapse to the smooth wall version.

For $h^+ < 5$ the flow will behave as a hydraulically smooth flow, with $\frac{AU}{u_*}$ going to zero Raupach and Antonia (1991). Fully rough flow will be found for $h^+ > 70$ and maximum observed values of $\frac{AU}{u_*} \approx 20 – 30$. Transitional flow is found in between these two limits. Using a mean (0.02 m s$^{-1}$) and maximum (0.05 m s$^{-1}$) magnitude estimate from the middle and highest elevation Vector records to approximate $u_* = 0.05U$ and $\nu$ at 10°C, the $h$ value needed for $h^+ > 5$ is in the range of 2–6 mm.

The bottom of Onondaga Lake is predominantly a thick viscous mud, heavily contaminated with various industrial compounds (Auer et al., 1996). Due to anoxic conditions in the hypolimnion it is also largely devoid of life capable of perturbing the sediment surface (Effler, 1996). The South Deep deployment region is described by Auer et al. (1996) as a broad flat plane, with surface observations using a boat mounted depth finder during deployment confirming this.

While flow in the BBL of Onondaga Lake is expected to be either hydraulically smooth or at the very low end of the transitional regime, the rough wall formulation above is used to allow for roughness effects, particularly since roughness will affect turbulence and turbulent mass flux (Reidenbach et al., 2010).

In the atmospheric community an alternative formulation of Equation 3.2 is used, discarding the viscous length scale in favor of normalization by a charac-
teristic roughness length $z_0$. Using this normalization, the Law of the Wall takes the form

$$\frac{\bar{U}}{u_*} = \kappa^{-1} \log \frac{z}{z_0}$$

(3.3)

This is simply an alternative, but equivalent, form of Equation 3.2. The relationship between $z_0$ and $\frac{\Delta U}{u_*}$ is given by Raupach and Antonia (1991) as

$$\frac{\Delta U}{u_*} = \beta + \kappa^{-1} \log \frac{z_0 u_*}{\nu}$$

(3.4)

Thus, $z_0$ and $\frac{\Delta U}{u_*}$ are entirely equivalent and interchangeable measures of roughness. Note Equation 3.3 is generally used in the atmospheric boundary layer where fully rough conditions are expected to exist. It is still valid for smooth walls and in the limit as $h^+ \to 0$, $z_0$ approaches the value $\frac{r}{u_*} \exp \kappa \beta$ (Raupach and Antonia, 1991).

One final modification incorporated to Law of the Wall is an adjustment of the measurement elevations. This is done by replacing $z$ with $Z = z + z_{\text{offset}}$. Ideally, the value of $z_{\text{offset}}$ is constrained between 0 and $h$ and represents the zero plane offset for the flow, the point at which the flow suggests the wall is located. Allowing for negative values permits adjustment for measurement uncertainty in the bed location. Many researchers have developed methods of estimating $z_{\text{offset}}$ independent of the standard velocity profile fit. Analysis by Thom (1971) and a later analytical treatment by Jackson (1981) showed $z_{\text{offset}}$ is the level at which the mean drag force appears to act on a rough surface, relating it to the geometry of the individual roughness elements. While intriguing, the detailed geometric knowledge needed for this calculation is not available for Onondaga
Lake nor is it generally relevant given the expected surface roughness characteristics (i.e. a smooth or very low wavenumber roughness).

Following the example of various researchers (Bandyopadhyay, 1987; Castro, 2007), the value of $z_{offset}$ is optimized in the least squares fitting of the velocity profile. This is expected to be inaccurate (Raupach and Antonia, 1991), but as will be shown in the results, does not affect the velocity profile fits in a significant manner. While there are methods to estimate $z_{offset}$ and check it’s validity independent of the velocity profile fit (Thom, 1971), they require an accurate, independent measure of $u_*$ and several measurement in the viscous or roughness sublayer (i.e. $z^+ < 4$), neither of which are routinely available for this dataset.

Typical velocity profiles are shown in physical coordinates in Figure 3.18. These are three typical profile shapes observed during the deployment. The center and left profiles shows evidence of a log region in the lower half of the profile. The left profile shows a decrease in velocity at measurements above 500 mm while the center profile continues the increase through the entire profile, possibly with the outer region wake function affecting the profile. The right profile shows a large bump in velocity in the near bed region before decreasing to a relatively flat profile above this. This right most profile will be discussed further below.

These same three profiles are plotted in plus coordinates using the least squares fit parameters and compared to Equation 3.2 and the DNS results of Spalart (1988) in Figure 3.19. Agreement with Equation 3.2 does not occur throughout the profile, but in all cases there is a region which appears to behave as the log region of a turbulent boundary layer. Agreement such as this is
Figure 3.18: Example velocity profiles from Days 12.6, 15.9, and 18.2 (left to right).

evident in approximately 85% of the profiles considered (141 out of 166). Those not showing agreement typically have only a few velocity points defining a log region or very little change in velocity with elevation. These profiles are also characterized by a profile average velocity below 0.01 m s\(^{-1}\) indicating fairly weak flows.

The right profile in Figure 3.18 shows a shape characteristic of an oscillatory (or Stokes) boundary layer. As previously mentioned, the main forcing leading to velocities in the BBL of Onondaga Lake is the baroclinic seiche, with regular
changes in mean velocity direction (i.e. the definition of oscillatory). However, one of the regions where this profile shape occurs frequently is days 17-18. During this time there is very little wind forcing, velocities are among the weakest observed, and there is no change in direction of the flow which is predominantly northwest. This indicates this is not an oscillatory flow (this mean direction persists at least as long as the seiche period) but rather some other forcing occurring in the BBL.

One possibility is a density driven current entering at the side of the lake. In
the late stratified period there is potentially a strong density difference between inflows and the lake water, as well as the potential for significant loss of heat at the surface. Onondaga Creek has been observed to enter as a plunging inflow due to its higher salinity (Effler et al., 2009), while the waste water treatment plant effluent enters the southern basin and very likely has different salinity and temperature than the lake water allowing for density driven flow. Another possibility is rotation. The Rossby number ($Ro = \frac{U}{fL}$, where $L$ is an appropriate basin length scale of 8 km length or 2 km width) and $f = 0.994 \times 10^{-4}$ is the Coriolis parameter at Onondaga Lake’s latitude) is $O(0.01-0.1)$ for Onondaga Lake using mean flows of 0.01-0.02 m s$^{-1}$. This indicates rotation could or very likely is influencing the flow.

Backscatter profiles from this period show occasional structure coinciding with the velocity bump lending some support to a different water mass in the near bed region. There is also a directional shear possibly indicating an Ekman spiral.

The best fit $u_*$ values from each model (Equations 3.2 and 3.3) are compared against one another in Figure 3.20. The agreement is generally quite good, although slightly higher values are typically obtained from the atmospheric model. When compared between fits performed including $z_{offset}$ and those not, there is essentially no difference in $u_*$ estimates (Figure 3.21). While the values obtained for $z_{offset}$ are not expected to be accurate, the inclusion of this parameter in the fit routine is reasonable as there is uncertainty in the location of the boundary. The boundary position is estimated from the backscatter profile with an accuracy +/- 0.01 m, half the range cell size.

Quantitatively there is little difference between the various estimates of $u_*$. 
Figure 3.20: Comparison of $u_*$ estimated from Equation 3.3 versus Equation 3.2 with roughness function incorporated. Both models were fit allowing for elevation adjustment.

and the mean flow estimate used to seed the least squares fit (Figure 3.22). The $u_*$ values used in subsequent boundary layer scalings are arbitrarily chosen as those obtained with $z_{off, set}$ included as a fit parameter and from Equation 3.2.

**Rough wall parameters**

The Onondaga lake BBL is expected to behave primarily as a smooth wall turbulent boundary layer based on scaling estimates. Occasionally, conditions might
lead to a transitional rough wall flow (e.g. during the highest velocity periods). Roughness effects were included when fitting \( u_* \) by including the \( \frac{\Delta U}{u_*} \) when fitting Equation 3.2 and by fitting Equation 3.3 and estimating \( z_0 \).

Using Equation 3.4, the best fit values for \( \frac{\Delta U}{u_*} \) were converted to the \( z_0 \) value, while \( z_0 \) estimates were converted to \( \frac{\Delta U}{u_*} \). Ideally, when \( \frac{\Delta U}{u_*} \) is converted to \( z_0 \), it should closely match the estimate of \( z_0 \) obtained by fitting Equation 3.3 (and vice versa).
Figure 3.22: $u_\ast$ estimated from Equation 3.2 with elevation adjustment and $\frac{\Delta U}{\mu_\ast}$ and the scaling estimate used to seed the least squares fit $(-)$.

A comparison between each best fit estimate and the equivalent value obtained by converting $\frac{\Delta U}{\mu_\ast}$ or $z_0$ is shown in Figure 3.23. Agreement between the two parameters is generally good as shown by the 1:1 line in each plot. The atmospheric model consistently produces higher values of both parameters however.

Histograms of the best fit $z_0$ values, the equivalent $z_0$ from $\frac{\Delta U}{\mu_\ast}$, and the smooth wall value $\frac{\Delta}{\mu_\ast} \exp -\kappa \beta$ are shown in Figure 3.24. The best fit and equivalent $z_0$ values have similarly shaped distributions and are typically within 0.002
Figure 3.23: (top) $\frac{\Delta U}{u_\ast}$ calculated from $z_0$ versus best fit $\frac{\Delta U}{u_\ast}$ values. (bottom) $z_0$ calculated from $\frac{\Delta U}{u_\ast}$ versus best fit $z_0$ values. The solid line indicates 1 to 1 correspondence.

$m$ of the expected smooth wall value, indicating roughness effects are minimal.

$z_0$ is directly related to the physical length scale of roughness. Raupach and Antonia (1991) provides the approximate relationship $z_0 \approx \frac{h}{30}$ for fully rough flow, while H (1936) defines the equivalent sand grain roughness as $h_s = 32.6z_0$. For crop and forest canopies, Raupach and Antonia (1991) cites values as low as $\frac{1}{17} - \frac{1}{10}$. There is thus a different scaling between $z_0$ and $h$ depending on the nature of roughness (granular, vegetation, etc.). Auer et al. (1996) describe the area around the South Deep deployment location as a broad, flat plain with an
average sediment particle size of $4 \times 10^{-5}$ m. Given the common use of equivalent sand grain roughness $h_s$ in engineering work, the $\frac{1}{36}$ scaling factor provided by Raupach and Antonia (1991) is used to estimate $h$.

Values of $h^+ = \frac{h u_c}{v}$ are calculated for the two $h$ estimates obtained from the best fit $z_0$ and $\frac{\Delta U}{u_c}$ equivalent $z_0$ converted to an estimate of $h$. $h_0^+$ will indicate the flow type (hydraulically smooth, transitional, or fully rough) with $h^+ < 5$ indicating smooth flow, $5 < h^+ < 70$ transitional flow, and $h^+ > 70$ fully rough flow. Histograms of the $h^+$ estimates are shown in Figure 3.25. There is a large per-
centage (≈33%) of values near the transitional value of $h^+ \approx 5$. The remainder are sporadically distributed over a wide range of values, with many values greater than 70. Given the slow mean flows at the South Deep site, it is extremely unlikely these are valid estimates of $h^+$. Given the slow measured velocities and the known characteristics of the bottom, these extremely large $h^+$ values are not believable. They are likely a result of increased uncertainty in the estimation of $u_*$ from a limited number of points in the velocity profile.

![Figure 3.25: Histograms of $h^+$ estimated using values of $z_0$ from Equation 3.3 (–•–) and calculated from $\frac{\Delta U}{u_*}$ (–○–).](image)

These roughness parameter results and $h^+$ scalings point to the Onondaga Lake boundary layer behaving as primarily as hydraulically smooth or weakly...
transitional. While very large values of $\frac{AU}{u_\ast}$ or $z_0$ are generated during the fit, there is little physically about the flow or bottom to lend support to these being anything other than errant results from the automated fitting routine. The small average sediment particle size at the South Deep site and their characterization as a "malodorous muck" (Auer et al., 1996) also suggest a smooth surface. For comparison, the highest values of $\frac{AU}{u_\ast}$ reported by Raupach and Antonia (1991) are from atmospheric measurements and are in the range 20-30 with most laboratory measurements less than 10.

Raupach and Antonia (1991) discusses modifications to turbulence above a rough wall in detail. The region important to turbulent mass transfer at the sediment-water interface is the roughness sublayer (or the analogous viscous sublayer in a hydraulically smooth flow). The roughness sublayer is defined by Raupach and Antonia (1991) as lying between $h < z < z_w$, where $z_w$ is dependent on the roughness geometry and lies between $2h$ and $5h$ (based primarily on measurements in the laboratory and atmospheric boundary layer). It is a region of very high turbulence intensity with the ratio $\frac{\sqrt{u'^2}}{\overline{u'}}$ typically between 0.5 and 5. Further, the diffusivities of momentum and scalars (heat and water vapor being the most common atmospheric scalar measurements) are enhanced by the presence of roughness above smooth wall values. With the expectation turbulence plays a dominant role in scalar fluxes at the sediment-water interface (discussed more fully in Chapter 4), understanding the boundary conditions expected at a field site where scalar flux is to be modeled or studied becomes exceedingly important.

The results here, while supporting a general characterization of the boundary as hydraulically smooth, have enough uncertainty transitional flow could
be encountered at times, resulting in enhanced scalar fluxes.

3.4.5 Hourly average turbulence statistics - Saddle

The fluctuating velocity component is calculated for each Vector burst using Equation 2.2, using as the mean component the burst average velocities. Turbulence is characterized by the intensity and Reynolds shear stress components as well as turbulent dissipation rates (discussed in §3.4.8).

As mentioned in §2.1.3, variance terms (i.e. the turbulence intensity) will be biased by noise. This is important to consider when examining turbulence in Onondaga Lake because the low energy of the flow results in turbulence and noise of comparable magnitudes.

As a first order test for turbulence, velocity spectra (Equation 2.8) were calculated for each burst averaging using multiple 512 point sub-spectra. For full length bursts this results in 5 sub-records at the highest elevation and 20 at the middle elevation, with respective frequency bin widths of $df = 0.0078\,\text{Hz}$ and $df = 0.0312\,\text{Hz}$. The inertial subrange of turbulent motion as described by Kolmogorov for isotropic turbulence Davidson (2004) should be seen as a $-5/3$ slope in the velocity spectra. At the Saddle, this characteristic slope is easily identified in both horizontal and vertical components.

For each burst and velocity component an estimate of the noise level is made by averaging values over the frequency range showing a flat response characteristic of noise. The average noise level is assumed the same at all frequencies (i.e. white noise) following the assumptions discussed in §2.1.3. The noise spec-
trum is then numerically integrated to obtain the variance due to noise in each component for correction of turbulence statistics. The frequency range for estimating the noise was set at > 1.0 Hz for the highest elevation Vector and > 3.0 Hz for the middle elevation Vector, with an upper limit set by the Nyquist sampling criteria (Shannon, 1949).

The resulting noise levels, averaged over all bursts, for the middle elevation Vector are $3.6 \times 10^{-3}$ m s$^{-1}$ in the horizontal components and $6.3 \times 10^{-4}$ m s$^{-1}$ in the vertical, a ratio of $\sigma_h/\sigma_v \approx 30$ as expected (Voulgaris and Trowbridge, 1998). For the highest elevation Vector, these values are $2.2 \times 10^{-3}$ m s$^{-1}$ in the horizontal components and $4.3 \times 10^{-4}$ m s$^{-1}$ in the vertical. The highest elevation Vector has slightly lower noise because it sampled at 4 Hz and thus averaged a larger number of internal pings, with the standard deviation of the data proportional to $1/\sqrt{N}$, where $N$ is the number of data points.

An example velocity spectrum and the estimated noise spectrum from the highest elevation Vector is shown in Figure 3.26. The estimated noise level in m$^2$ s$^{-1}$ from each burst for each component are shown in Figure 3.27. The difference in noise between the two horizontal components is due to either mean shear in the sample volume (Lhermitte and Lemmin, 1994) or the summation of noise terms from three receivers for the $u$ velocity component while the $v$ velocity component utilizes only two receivers. The vertical component, owing to the probe geometry, is significantly lower in noise as expected.

Figure 3.28 presents the burst average turbulence intensity values, corrected for noise bias. Correction was performed using Equation 2.1.3 prior to taking the square root in the RMS sequence of calculating turbulence intensity. Turbulence intensities are typically 0.005-0.01 m s$^{-1}$ in the horizontal and 1-4 $\times 10^{-3}$
m s$^{-1}$ in the vertical. Mean confidence interval widths are 5 and $1 \times 10^{-4}$ m s$^{-1}$ averaged over all valid bursts for the horizontal and vertical components respectively. Intensities are similar at each elevation, and follow similar trends as the mean flow as it increases and decreases during the seiche period. A scatter plot of each intensity versus the mean flow magnitude is shown in Figure 3.29, showing the approximately liner relationship between the two.

Dimensional Reynolds shear stresses are shown in Figure 3.30. Because of the geometry and noise characteristics of the Vector, the Reynolds stress estimates are essentially unaffected by noise (see discussion in §2.1.3). Stresses are normally extremely small with average values near zero. Estimating $u_*$ as 5% of the mean flow (mean velocity profiles are unavailable to estimate $u_*$ from a least squares fit as was done at South Deep), shown in §3.4.4 to be a reasonable estimate, the $\overline{u'w'}$ stress is scaled by $u_*^2$ and plotted alongside $u_*^2$ in Figure 3.31.
Figure 3.27: Spectral noise level for the stream wise (●), cross-stream (△), and vertical (*) velocity components.

Approximately every 24 hours, coinciding with increases in turbulence intensity and mean velocity, there are increases in the Reynolds shear stresses. During periods of higher mean flow, indicated by higher values of $u_*$ in the bottom axis of Figure 3.31, $u_2^*$ is larger than the corresponding $u'w'$ term. During periods of slower mean flow, the magnitudes change and $u_2^*$ becomes fairly small. This accounts for the high values of dimensionless stress seen in the top axis of Figure 3.31.
The turbulence data available from the South Deep Vector datasets is limited by the combination of generally weak flows and a suboptimal instrument setup. As mentioned in §3.4.3 this affected the lowest elevation Vector from Day 18 on, but played a role in the measurement quality at all three elevations. The lowest elevation Vector was the most affected and excluded from turbulence analysis.

Turbulence intensities corrected for noise using the same method as for the Saddle are shown in Figure 3.32. There are numerous times when turbulence in-
Figure 3.29: Turbulence intensities for the high (●) and middle (△) Vector at the Saddle plotted against the corresponding burst mean flow magnitude.

tensities are less than the noise variance, resulting in negative values for the $\overline{u'^2}$ term and imaginary values for the turbulence intensity. At the highest elevation, the horizontal turbulence is almost always below the noise floor of the instrument, expressed as a variance due to noise of $6 \times 10^{-5} \text{ m}^2 \text{ s}^{-2}$. Vertical turbulence intensities are more frequently detectable, but still falling below the noise floor ($2 \times 10^{-6} \text{ m}^2 \text{ s}^{-2}$) a majority of the time. At the middle Vector, horizontal turbulence intensities are generally higher than the noise floor ($4 \times 10^{-5} \text{ m}^2 \text{ s}^{-2}$). The vertical component noise floor is $10^{-6} \text{ m}^2 \text{ s}^{-2}$.
Reynolds stresses (shown in Figure 3.33) are extremely weak at this site. While there are a few increases in Reynolds shear stress seen in the time series, they are extremely intermittent occurring every few days on average, far less frequently than observed at the Saddle. Stress magnitudes are significantly smaller than the Saddle, particularly during turbulent events.
Figure 3.31: (top) $|\overline{u'w'}|$ at the high (●) and middle (△) Vector at the Saddle scaled by an estimate of $u_*$ as 5% of the mean magnitude at each sensor. (bottom) Dimensional $|\overline{u'w'}|$ compared with $u_*^2$ from the high elevation (–) and from the middle elevation (– –) mean magnitude.

3.4.7 Hourly Vertical Turbulence Profiles - South Deep

Using beam velocities as a surrogate for vertical velocity (the beam angle is 25°, meaning some horizontal velocity will be measured), estimates of the vertical turbulence intensity are obtained from the HR Profiler data set. Individual beam intensities are shown in Figure 3.34 and the average across all three beams is shown in Figure 3.35.
Figure 3.32: Turbulence intensities for the high (●) and middle (△) Vector at South Deep.

All intensities have been corrected for noise bias following the same method used for the Vector data. Each bin and beam is treated as a separate time series, with noise expected to be a function of cell range from the transducer (Gordon et al., 1999). Individual beam velocity spectra were calculated with a noise estimate estimated from the mean spectrum value from $0.30 < f < 1$ Hz, where 1 Hz is the Nyquist sampling frequency. An example velocity spectrum from range cell 10 showing a characteristic $\frac{-5}{3}$ slope indicative of the inertial subrange is shown in Figure 3.36.
Intensities compare favorably with Vector measurements for order of magnitude. But, there is significant scatter between the two estimates as shown in Figure 3.37. The HR Profiler has greater spatial averaging (range cells are roughly 0.02 m high with a diameter of 0.03 m, the Vector sample volume has an approximate height and diameter of 0.015 m) but uses fewer internal samples in a velocity measurement due to the increased processing time needed to measure and process a profile. The resulting transfer function is thus different and likely accounts for some of this scatter. The different physical sampling locations also contribute to scatter. The velocities measured by both instruments show clear
Figure 3.34: $w'_{\text{rms}}$ for each beam for the HR Profiler.

Inertial subrange indicating there is turbulent energy at scales measurable by each instrument, while the comparison of noise corrected turbulence intensities suggest this is not merely noise tracking.

Intermittent bursts of turbulence, approximately twice the magnitude of background levels, occur in the upper half of the profile during the first few days. Comparing with Figure 3.17, periods of higher turbulence intensity coincide with periods of high mean flow, although surprisingly the most energetic events (e.g. Day 12.5) do not coincide with the strongest mean flows (e.g. Day 13.5). After Day 16 when there is a near bed peak in velocity, there is a coincid-
Profiles of turbulence intensity for the three profiles shown in Figure 3.18 are plotted in Figure 3.38. The left and center profiles show some variation near the bed, increasing and decreasing respectively in the last few bins of the profile. The center profile shows an isolated burst of turbulence in the upper portion of the profile, approximately 0.10-0.15 m in size with 2.5 times higher intensity. There is no detectable change in backscatter associated with this event, ruling out biological activity such as a fish within the beam being responsible,
Figure 3.36: Beam velocity spectra (beam 1: –, beam 2: –, beam 3: – –) from range cell 25 for the burst occurring on Day 12.6294. A $\frac{-5}{3}$ slope is indicated by – – while the rectangle indicates the region averaged to determine noise level.

an unlikely scenario given the anoxic conditions in the hypolimnion. Given the proximity of this event in the upper profile, close to the frame top cross beam, and it’s occurrence within only one beam (specifically Beam 3, oriented across the long axis of the frame) it seems this is instead a potential wake effect of the frame.

When normalized and plotted in wall coordinates, using the $u_*$ estimates developed earlier in §hrSouthDeepMean by fitting the mean velocity profile,
turbulence intensities are of an appropriate magnitude but show significant deviations from the expected structure for a smooth wall boundary layer. Profiles from the same bursts shown in Figures 3.18 and 3.38 are shown in Figure 3.39 compared to the DNS results of Spalart (1988). The center profile shows the best agreement with the expected structure in magnitude and shape, coinciding with the good agreement of the mean profile and Equation 3.2 in Figure 3.18. While $u_\tau$ is expected to be an appropriate velocity scale for boundary layers, in the absence of classic boundary layer mean profile structure it is not ideal for
producing consistent scaling of the turbulence such that a measurement of the mean profile will yield reasonable estimates of the turbulence.

3.4.8 Turbulent Dissipation - Saddle

Turbulent dissipation ($\epsilon$) is estimated for each velocity component for the high and middle Vectors using three methods (discussed in §2.1.2), a velocity spectrum, a second order structure function, and a third order structure function. Corrections to the spectrum and second order structure function for noise bias
Figure 3.39: Profiles of $w'_{\text{rms}}$ for Days 12.6, 15.9, and 18.2.

are applied. The third order structure function is expected to be unbiased by noise. Taylor’s Frozen Turbulence Hypothesis, using the mean magnitude for a burst ($U$), is used to estimate the separation distance between measurements ($r = \bar{U}t$) and wavenumber ($k = \frac{2\pi f}{U}$), where $t$ time in seconds and $f$ is frequency in Hz for a temporal velocity spectrum.

Time series of dissipation from all three methods for the middle elevation Vector are shown in Figure 3.40. Dissipation follows similar trends as the turbulence intensities and Reynolds stresses, changing by several orders of magnitude as mean velocity rises to a peak and then decays towards zero during
a direction change. During the second half of this dataset, dissipation estimate quality decreased as the number of samples within a burst decreased due to falling battery power. This resulted in much higher scatter of dissipation estimates and poorer agreement between the three methods. Throughout the deployment, the vertical velocity fluctuations provide the best estimates and discussion will focus primarily on them. The anisotropy observed in turbulence intensities carries over to $\epsilon$, particularly during the zero velocity periods.

Figure 3.40: Turbulent dissipation at the Saddle from the middle Vector. Top to bottom represent estimates from the horizontal and vertical velocity components. Velocity spectrum estimate ($\bullet$), second order structure function ($\triangle$), third order structure function ($\ast$).
The velocity spectrum estimates of dissipation are typically the lowest values, with the second order structure function generally slightly higher, and the third order structure function reporting the highest estimates. The differences between the three estimates are primarily due to noise and the imperfect noise correction applied, suggesting noise is a factor in the third order structure function for low energy systems. The spectral noise estimate is assumed constant at all frequencies, however using the redundant vertical velocity measurements a frequency dependent structure occurs (see Figure 2.1 and Rusello and Cowen (2011); Hurther and Lemmin (2001); Blanckaert and Lemmin (2006)). This is likely due to the instrument transfer function.

Understanding bias in the third order structure function requires knowing the magnitude of noise terms such as $\sigma_i^2\sigma_j^2$ which are presently assumed to be zero. Because of the low energy of this system, these terms might be of comparable magnitude to the cubed velocity terms, creating a bias in the third order structure function. Because of the significantly lower noise in the vertical velocity component, discussion will be limited to the vertical dissipation estimates.

During the flow reversal periods, i.e. when the mean velocity drops to zero, dissipation drops to values $O(10^{-10} \text{ m}^2 \text{ s}^{-3})$. This is approximately one to two orders of magnitude lower than values cited by Wuest and Lorke (2003) for the BBL. Given the preceding discussion of noise, care must be taken to verify values this low. By compensating and normalizing the velocity spectrum and second order structure function and comparing to expected universal forms such as the Pao spectrum (Davidson, 2004), the reasonableness of the dissipation estimates can be checked. The structure function, when compensated and normalized should be equal to the constant $C_2$ in the inertial subrange.
Both the velocity spectrum and second order structure function show the expected structure when normalized (Figure 3.41). There is some discrepancy with the expected forms suggesting slightly too low values of $\epsilon$ and $\eta = (\epsilon^3)^{1/4}$, but structurally, the velocity spectrum and second order structure function confirm the presence of turbulence even during these very low energy periods.

Figure 3.41: Normalized vertical velocity spectrum compared to the Pao theoretical spectrum Davidson (2004) (top). Compensated second order structure function compared to the expected value of the constant $\frac{4}{3}C_2$. 

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3.4.9 Turbulent Dissipation - South Deep

In addition to the estimates obtained from the middle and highest elevation Vectors, the HR Profiler velocity data, with each bin treated as an individual time series, can be used to estimate dissipation. Additionally, using the $u_*$ estimates obtained from the mean velocity profiles, dissipation can be estimated using the following scaling estimate Tennekes and Lumley (1972)

$$\epsilon = u_*^3 (kz) \frac{1}{1}$$

(3.5)

Time series of $\epsilon$ from the middle elevation Vector are shown in Figure 3.42. There is slightly more scatter among the three estimates with the structure function estimates performing slightly worse (based on scatter and generally higher values) than the spectral estimate. This poorer performance is primarily due to the simplistic procedure used to determine $\epsilon$ from the structure functions, where the maximum value of the compensated structure function in the range $11 < \frac{\xi}{\eta} < 100$ is selected as the estimate for dissipation. A refined peak identification algorithm to exclude outliers (i.e. spikes in the structure function curve) will lead to more reasonable estimates.

The spectral estimate uses a fixed wavenumber range which consistently contains the inertial subrange. The influence of outlying spectral values in this wave number range is reduced by taking an average over the range to estimate $\epsilon$. When employing an average over an assumed range of $\frac{\xi}{\eta}$ representing the inertial subrange for the structure functions, $\epsilon$ is underestimated because the expected peak coinciding with the inertial subrange is not consistently located at the same $\frac{\xi}{\eta}$ value. This results in values below the peak (i.e. outside the
Figure 3.42: Turbulent dissipation at the South Deep location from the middle Vector. Top to bottom represent estimates from the horizontal and vertical velocity components. Velocity spectrum estimate (●), second order structure function (△), third order structure function (*).

(inertia subrange) included in the average, biasing estimates of $\epsilon$ low.

Comparison between the spectral estimate from the middle elevation Vector, the spectral estimate from the nearest bin from the HR Profiler, and the scaling estimate from Equation 3.5 are shown in Figure 3.43. The value of $z^+$ here is approximately 400 but varies with $u_*$ and thus the mean flow. This approximate $z^+$ value is at the outer edge of the log velocity region, suggesting a transition
to outer layer scaling (DeGraaff and Eaton, 2000) is likely needed. However, the outer layer scaling requires an estimate of the boundary layer thickness and $Re_{\theta}$, unfortunately both of which are subject to large uncertainty in this flow given the often complex structure of the flow. Despite this, the inner layer scaling (i.e. $z^+$) and Equation 3.5 should provide a reasonable, although certainly not ideal, expectation for the magnitude of $\epsilon$. Spectral estimates from the HR Profiler and middle elevation Vector are typically less than the boundary layer scaling estimate, sometimes by several orders of magnitude, while the HR Profiler estimates are consistently lower than those from the Vector. Uncertainty in $u_*$ contributes directly to the uncertainty in the scaling estimates.

Compared to the Saddle location, dissipation is slightly lower on average at the South Deep location owing to the lower mean flow and turbulence levels. Drops to $O(10^{-10} \text{ m}^2 \text{ s}^{-3})$ during flow reversal are routine, and depending on which instrument is being examined values below this might occur. It seems unlikely there is turbulent production occurring during these flow reversals given the low values of $u'w'$ measured by the Vectors (with turbulent production equal to $u'w' \frac{du}{dz}$), suggesting this is primarily decaying turbulence generated during the high velocity periods.

Average profiles of $\epsilon$ across all three beams for each method from the HR Profiler are compared to Equation 3.5 in Figure 3.44. Profiles coincide with the three mean velocity profiles shown in Figure 3.18. The best comparison to the scaling result is achieved in the center profile, which also showed the best comparison to the expected mean velocity and turbulence intensity profiles. The left and center profiles both show decreases in dissipation near the bed, while the right hand profile shows an increase near the bed. The scaling estimate derived
Figure 3.43: Dissipation estimates from the middle Vector spectral method (●), the HR Profiler spectral method (△), and the boundary layer scaling estimate (−).

from Equation 3.5 routinely overestimates $\epsilon$ in this flow.

### 3.4.10 Onondaga Conclusions

The bottom boundary layer mean flow and turbulence in Onondaga Lake is strongly tied to baroclinic seiching in the basin. Seiche periods are predicted reasonably well by the simple two layer model introduced in §3.4.1, with temperature, velocity and velocity direction power spectral densities confirming
periods in the 20 hour range.

Measured mean flows were fairly weak throughout the basin, slightly higher at the Saddle than at the South Deep location. Mean velocities rarely exceeded 0.05 m s$^{-1}$ while flow direction was oriented along the main axis of the lake (northwest-southeast) at both sites. Portions of the mean velocity profile showed agreement with Equation 3.2 approximately 85% of the time, typically the bottom 0.50 m of a profile following the expected logarithmic velocity pro-
file (Equation 3.2). Some evidence of an underflow during the last few days of the South Deep deployment was observed in the bottom 0.30 m of the profile.

Corrections for noise bias improved estimates or turbulence and indicated turbulence intensities below detectable levels for the instrument setup at the South Deep site. All three turbulence components followed the same trends as the mean flow, increasing with increasing mean flow magnitude. Vertical turbulence profiles measured by the HR Profiler showed fairly constant intensity with elevation and when scaled by $u_*$ estimates obtained from fitting the mean velocity profile were somewhat weaker than expected.

Reynolds stresses were weak at both sites, with near zero values except during the strongest mean velocity periods. They quickly decayed to near zero once the mean velocity decreased. Spikes in stress coincided with the thermocline displacement, positive stresses (i.e. a negative $\overline{u'w'}$) occurring when the thermocline moved away from the boundary and negative stresses when it moved towards the boundary. Compared to vertical turbulence intensities, $\overline{u'w'}$ was of comparable magnitude, suggesting a rough estimate for the magnitude of the $\overline{u'w'}$ by measurement of the vertical turbulence intensity.

Dissipation was extremely low throughout the basin, with typical levels $O(10^{-8} \, \text{m}^2 \, \text{s}^{-3})$. Compared to scaling estimates, measured dissipation rates were below expected values for a turbulent boundary layer despite the agreement of mean velocity profiles with Equation 3.2. Uncertainty in $u_*$ will affect the scaling estimate, however because dissipation is proportional to $u_*^3$, this is not expected to translate into an order of magnitude change in dissipation values. Dissipation again followed the trends seen in the mean velocity, increasing during periods of higher velocity and decaying to $O(10^{-10} \, \text{m}^2 \, \text{s}^{-3})$ during flow reversals.
when mean velocities were near zero. Velocity spectra and second order structure functions showed good agreement with universal forms even during low periods of weak dissipation.

The mean velocity below 0.50 m behaved like a turbulent boundary layer in a hydraulically smooth or weakly transitional state based on values of the roughness function, roughness height $h$ and roughness Reynolds number $h^+$. Turbulence measurements suggest $u_*$ is a reasonable velocity scale for the flow. Turbulence levels were less than scaling estimates based on prior smooth wall boundary layer results.

While no direct measurements of turbulent mass flux at the sediment-water interface were made during this deployment, these measurements provide a range of expected turbulence levels a laboratory facility will need to produce to be used in a study of turbulent mass transfer (e.g. Chapter 4). They also represent the first direct measurements of velocities in the bottom boundary layer of Onondaga Lake and will undoubtedly prove useful in ongoing efforts to remediate this highly polluted lake.

3.5 Ashokan

Measurements were carried out during July 2007 in the West Basin during two different deployments. The first deployment was made in Esopus Creek before it enters the West Basin. The second deployment was made in the main basin, south of the East basin connection, on a sloping section of the bottom. The instrument frame was deployed at the approximate thermocline depth.
On lake wind was recorded by an Upstate Freshwater Institute (UFI) Remote Underwater Sampling Station (RUSS) located in the northwestern half of the West Basin, approximately 2 km from the Esopus Creek mouth. This RUSS (referred to as the west RUSS) and one deployed near the East basin connection (east RUSS) recorded temperature profiles every six hours. Wind speed and direction were reported as 15 minute vector average.

The instrument frame (Figure 3.45) had four legs and was approximately 2 m high, permitting a down looking Teledyne-RDI 1200 kHz ADCP to be mounted at the apex of the four legs. The four acoustic beams of the ADCP were oriented between the legs of the frame and utilized high-resolution pulse coherent processing (Mode 11 as referred to by the manufacturer) to measure water velocities. Based on backscatter and pressure measurements, the ADCP was approximately 1.5 m from the bed in both deployments. A Nortek Vector was mounted sideways at the bottom on one leg (≈0.20 m elevation) to measure near bed turbulence. While the XYZ Vector coordinate system was expected to be aligned with the vertical, the mounting sagged during the deployment resulting in an unknown roll angle in the data. During the main basin deployment, an Optical Backscatter Sensor (D&A Instrument OBS-3), calibrated to measure turbidity, was mounted to one of the legs. It sampled approximately the same location and elevation as the Nortek Vector.

Mounted on one leg were SBE39 temperature and SBE39 temperature/pressure loggers. During the Esopus Creek deployment, all four SBE39s sampled pressure and temperature, during the main basin deployment only the lowest also sampled pressure. The SBE39s were mounted at 1.55, 1.18, 0.78, and 0.38 m above the bed. During the Esopus Creek deployment, all SBE39s sam-
Figure 3.45: Instrument frame used during the Summer and Fall 2007 deployments in Ashokan and Schoharie Reservoirs.

Sampled at 3 second intervals. This was increased to 6 seconds for the main basin deployment. Because the 3 second interval does not allow the SBE39s to sleep, only two days of data is available for the Esopus Creek deployment.

Local water depth during each deployment was 3.0 m (Esopus Creek) and 7.6 m (main basin) and remained relatively constant throughout the deployment period as the reservoir was near full capacity.

3.5.1 Meteorological Conditions and Stratification

Wind during the period were weak to moderate as seen in Figure 3.46, except for a few short duration wind events. Winds were predominantly out of the northwest (Figure 3.47), closely aligned with the basin’s main axis.
Figure 3.46: Wind speed and direction during July 2007, measured in the West basin of Ashokan reservoir.

Figure 3.48 shows temperature profiles during July 2007 measured at the two RUSS stations in the West basin. Internal waves can be seen in the thermocline, especially in the East RUSS temperature profiles, but they are of fairly small amplitude. Because this is a managed drinking water reservoir, the thermocline position will also influenced by controlled inputs from the upstream Schoharie Reservoir and withdrawals transferring water from the West to East basins.
The two layer model described in §3.4.1 predicts a baroclinic seiche period of 11.5 hours. This estimate is based on a mean thermocline position of 8 m with $h_1 = 8$ m and $h_2 = 10$ m, using an average depth in the West basin of 18 m. Average temperatures in the surface mixed layer (top 8 m) and hypolimnion (9 m and below) are 21 °C and 10°C, respectively. The basin length used in these calculations was $L = 5.6$ km.

The two strongest wind events, on Days 13 and 21, have little effect on the thermocline position but do appear to mix out the small gradient that formed in the surface mixed layer preceding each event. Blowing from the northwest, these events appear to sharpen the thermocline temperature gradient. Surface mixing lowers temperatures slightly during these events, mixing out the temperature gradient in the surface mixed layer, but also depresses the thermocline, moving it deeper. This has the effect of reducing the wind energy available to
Figure 3.48: Temperature profiles measure at the West (left) and East (right) RUSS stations in the West Basin.

for mixing at the thermocline itself.

3.5.2 Esopus Creek

Esopus Creek Discharge

The Esopus Creek discharge is shown in Figure 3.49 for Days 10 to 24, starting 1.5 days before the instrumentation was deployed. Discharge was measured by the USGS at the Coldbrook, NY gage, number 01362500. Discharge was fairly low, but from Day 13.5 to 15.5 there is an increase in flow with characteristics of a release from the Shendaken Tunnel into Esopus Creek and the West basin (i.e. a flat hydrograph). The discharge increase just prior to Day 24 occurred after instrumentation had been retrieved. The other small increases showing more typical hydrograph behavior, such as an exponential decay, are attributed to precipitation events. While no rainfall gages were maintained as part of the meteorological monitoring, it was the middle of summer and thunderstorms are
frequent occurrences in the Catskills during this season.

Figure 3.49: Esopus Creek discharge during July 2007.

Temperature Time Series and Spectra at the Instrument Frame

Figure 3.50 shows temperatures from the USGS Coldbrook gage (USGS gage number 01362500), the ADCP and ADV thermistors. The SBE-39 T/P loggers all died within 2 days of deployment are not shown. Temperature shows a strong diurnal variation at the streamflow gage and at the ADCP. Temperatures at the streamflow gage and the instrument frame warm and cool at the same rate, with slightly cooler temperatures seen throughout the deployment at the streamflow
gage. This apparent bias is approximately 1°C, is probably a result of drift in the temperature sensors onboard the ADCP and Nortek Vector. Both instruments were several years old at the time of deployment and had not had ancillary sensors such as temperature calibrated in that time. With the exception of Day 20 when a discharge increase occurs, there is no variation in the temperature from its typical diurnal cycle.

Figure 3.50: Temperatures measured at the Coldbrook streamflow gage (−) and at the instrument frame by the ADCP (- - -) and the Vector (−).
Velocity Profiles and Boundary Layer Parameters

A 1200 kHz ADCP was set to record a burst of 1024 pings at 2.5 Hz (approximately 6 minutes and 50 seconds of data) every hour. The blanking distance was 0.5 m and depth cell size 2 cm, resulting in a 1 m profile after processing. Processing follows similar methods discussed for Onondaga Lake velocity profiles, with a burst average magnitude and direction calculated and turbulence statistics performed on the individual beam velocity data. The Vector recorded a 2400 sample burst at 8 Hz (5 minutes of data) hourly. Because there was no capability of reprogramming start times in the field on this deployment, weather delays pushed deployment past the programmed start times resulting in ADCP and ADV bursts offset by 20 minutes, with the Vector sampling prior to the ADCP.

The ADCP data shows a biased region throughout the deployment starting near 0.5 m elevation and moving slowly up as the time advances. This is a region of pulse interference, an unfortunate consequence of utilizing pulse coherent processing near boundaries, where a prior acoustic pulse reflecting off of the boundary interferes with the present pulse. The changing position of this interference band is atypical of pulse interference. Based on backscatter measurements, the boundary position appears to be stable. Speed of sound effects would have been significantly smaller than the observed change in position. Teledyne-RDI was contacted regarding this problem but did not come up with a cause. One potential cause of such behavior, which unfortunately is not verifiable with the data produced by the ADCP, is a changing pulse distance throughout the deployment. While this behavior is supported, it requires an additional operating mode (bottom tracking) which was not installed on this instrument.
during the deployment. This region is not considered when discussing results.

Figure 3.51 presents burst averaged magnitude and direction from the ADCP. The period between Days 14 and 16 has been removed from the dataset because of strongly inhomogeneous flow in the four beams, leading to biased velocity estimates. When used to measure velocities in an orthogonal coordinate system (XYZ or ENU), there is an assumption of flow homogeneity over the horizontal region the diverging beam pattern covers. In this case, strong inhomogeneities were observed in the upstream and downstream beams resulting in severely biased velocity estimates when compared to the Vector, which does not suffer from this potential problem. Because a correction for these regions was not possible, they are not considered when examining flow magnitude but are considered when examining some turbulence statistics which utilizes the beam velocities directly. The unequal response of the up and downstream beams is either due to a real physical flow effect (supported by the turbulence intensity data discussed later) or potentially by ambiguity problems in resolving the measured velocity.

This region corresponds to the presumed Shandaken tunnel discharge event. The higher flow could have moved debris into the measurement region causing a flow obstruction visible in only one beam. While unlikely because of a fairly large ambiguity velocity, one of the beams could have been subject to velocity aliasing. However, this would have affected both the up and down stream beams since they measure the same velocity magnitude with opposite signs. Unfortunately, without direct observation of the site, it is difficult to identify the cause of such a problem from the signatures contained in the ADCP data.

Magnitudes are O(0.05 m s\(^{-1}\)) with nearly constant flow along the stream
channel to the east-southeast. There are four times during the deployment when flow is considerably slowed (approximately Days 13, 21, 22, and 23), with the first and last two stoppages resulting in a flow reversal for several hours. Each of these flow stoppages occurs with an increase in wind from the northwest. Temperature is unfortunately an ineffective tracer to see if this is main basin water moving upstream as both it and the Esopus Creek water are the same temperature. Backscatter, discussed more fully later increases during each of these events, potentially serving as an effective tracer for the water mass origin.

Figure 3.51: Magnitude and direction measured by the ADCP during the Esopus Creek deployment.

The Vector magnitude and vertical velocity is shown in Figure 3.52. The
Vector was unaffected by the flow inhomogeneities affecting the ADCP and provides valid data during Days 14-16. Near bed velocities average 0.10 m s\(^{-1}\), but during the Shendaken Tunnel release almost double to approximately 0.20 m s\(^{-1}\). This was a fairly small discharge increase from 350 cfs to 750 cfs, but because it is sustained at this level for approximately 2 days the effect on mean velocities will be different than for a precipitation induced discharge event, which will peak then decay exponentially. Regardless, the Esopus Creek channel velocity, especially near the bed, is affected by discharge.

![Graph](image)

**Figure 3.52:** (top) Magnitude (–) and vertical velocity (–) and (bottom) direction measured by the Nortek Vector in Esopus Creek.

Velocity profiles are fit to Equations 3.2 and 3.3, allowing for roughness ef-
fects and an elevation offset. A time series of the best fit value from Equation 3.2 and the mean velocity scaling estimate is shown in Figure 3.53. Agreement of the best fit values between Equations 3.2 and 3.3 is shown in Figure 3.54, with the atmospheric model producing consistently higher best fit estimates, with an RMS difference between the two estimates of $4 \times 10^{-4}$ m s$^{-1}$. The estimates from Equation 3.2 are used for scaling flow statistics.

Figure 3.53: Best fit values of $u_*$ (●) and the mean velocity estimate (−) for the Esopus Creek deployment.

Best fit values of $\frac{\Delta U}{u_*}$ and $z_0$ and their equivalents calculated using Equation 3.4 are compared in Figure 3.55. The atmospheric model produces higher estimates of each roughness parameter, but both forms of the equation result in a
wide range of roughness parameter values. Histograms of the best $z_0$ and equivalent $z_0$ from $\frac{M_U}{\bar{u}_r}$ along with the expected smooth wall value are shown in Figure 3.56. While highlighting the wide range of $z_0$ values obtained in the fitting procedure, there is an appreciable mass of $z_0$ estimates greater than the smooth wall value. Histograms of $h^+$, with $h$ estimate as $30z_0$, are shown in Figure 3.57, with most values lying above the transitional flow threshold of $h^+ \approx 5$.

Using the $\frac{1}{30}$ scaling to estimate the roughness height $h$ for Esopus Creek produces a mean estimate of 0.07 m and median estimate of 0.03 m. While
Figure 3.55: Best fit values of (left) $\frac{\Delta U}{u_\ast}$ and (right) $z_0$ and their respective equivalent values for the Esopus Creek deployment. The solid lines indicate a 1:1 correspondence.

not directly observed, the streambed in the deployment region was granular and potentially sandy based on the feel of bed when deploying the instrument frame. Upstream of the deployment location there were a variety of rocks, boulders and other debris. Roughness height estimates of a few centimeters appear reasonable based on the observational evidence available.
Figure 3.56: Histograms of $z_0$ from Equation 3.3 (●), the equivalent $z_0$ from $\frac{\Delta u}{u_p}$ (△), and the smooth wall value $\frac{u_p}{u_*} \exp(-\kappa \beta)$ (– –) during the Esopus Creek deployment. The x-axis is on a log scale.

**Turbulence**

Turbulent intensities from the ADCP are shown in Figure 3.58 and from the Vector in the top plot of Figure 3.59. Because each ADCP beam is utilized separately, flow inhomogeneities do not affect turbulence intensity calculations while velocity ambiguities will typically reduce variance because of the phase wrapping effect constraining values to a narrow range. Using velocity spectra to estimate a noise variance, turbulence intensities have been corrected for noise bias using
Equation 2.1.3. For each ADCP beam (1-4), the average noise variance across all range cells and bursts are 0.019, 0.017, 0.007, and 0.030 m s\(^{-1}\). For the Vector, these noise variances are \(1.4 \times 10^{-5}\), \(2.0 \times 10^{-5}\), and \(4.0 \times 10^{-6}\) for the \(x\), \(y\) and \(z\) components in the instrument coordinate system, with the \(z\) component oriented approximately cross-stream while the \(y\) component was oriented vertical due to the sideways head mounting.

Typical turbulence levels are 0.01-0.02 m s\(^{-1}\) in all three velocity components, with a large increase in turbulence intensity during the Shandaken Tunnel dis-
charge from Days 14-16. Good agreement in turbulence intensity magnitude is observed between the ADCP and Vector data. During the four flow stoppages discussed earlier, turbulence decreases significantly, often falling to near zero intensities.

Figure 3.58: Beam turbulence intensities measured by the ADCP during the Esopus Creek deployment.

Reynolds shear stresses from the Vector are shown in the middle plot of Figure 3.59. Stress levels are typically non-zero and average 0.1 Pa. During the Shandaken Tunnel discharge event from Days 14-16, all three stress components increase significantly to O(1 Pa). Stress levels during the flow stoppages decay below background levels, mirroring the decreases seen in turbulence intensity.
Figure 3.59: From the Vector during Esopus Creek deployment:

(Top) Turbulence intensities, $\sqrt{u'^2}$ (●), $\sqrt{v'^2}$ (△), $\sqrt{w'^2}$ (⋆).

(Middle) Reynolds stresses, $-\rho u'u''$ (●), $-\rho v'v''$ (△), $-\rho w'w''$ (⋆).

(Bottom) Dissipation estimated from the vertical velocity spectrum (●), second order structure function (△) and third order structure function (⋆). The y-axes for Reynolds stresses and dissipation are on a log scale.
Reynolds stresses calculated from the ADCP data (Stacey et al., 1999) show similar trends and magnitudes as the Vector estimates (Figure 3.60. During Days 14-16, the flow inhomogeneity period mentioned previously, causes problems with the method used to estimate the Reynolds shear stresses. This method requires homogeneity in the mean and variance of the beam data, which is not met during this period (mean beam velocities as well as variances are different between beams) and results in biased stress estimates for the $\overline{u'w'}$ components which utilize the upstream and downstream beams. The cross-stream beams were not affected and produce valid stress estimates throughout the deployment. Vertical structure exists in the $\overline{v'w'}$ component, with stress increasing above 0.4 m elevation. The $\overline{u'w'}$ component is more uniform throughout the water column. The deployment location was near a bend in the channel which could account for the vertical structure observed in the cross-stream stress component.

Dissipation estimates from the Vector are shown in the bottom plot of Figure 3.59, while the ADCP dissipation estimates obtained from velocity spectra are shown in Figure 3.61. Following similar trends as the other turbulence quantities, dissipation increases from background levels during the Day 14-16 discharge event, reaching values greater than $O(10^{-5} \text{ m}^2 \text{ s}^{-3})$, while background levels tend to be an order of magnitude less. The four flow stoppages occurring during the deployment all result in dissipation falling to $O(10^{-8} \text{ m}^2 \text{ s}^{-3})$.

There is good agreement between the velocity spectra and structure function estimates of dissipation during this deployment for the Vector data, with all three estimates at similar magnitudes rather than spread over an order of magnitude as seen in the Onondaga Lake data. In addition to higher turbulence
levels, the Vector data quality was better owing to higher scatterer content in the water column. Dissipation estimates from the ADCP agree with the estimates from the Vector, showing similar trends and orders of magnitude. Occasional patches of increased dissipation occur in the ADCP profile, most evident from Day 20 on, but there is little vertical structure observed.

The overall higher turbulence levels in the Esopus Creek channel compared to expected conditions in the main basin are partially due to the higher velocities encountered in the stream channel. However, the effects of boundary roughness
Figure 3.61: Dissipation estimates for the ADCP during the Esopus Creek deployment obtained using velocity spectra.

are a significant contribution to the amount of turbulent energy available (see discussion in §3.4.4) and will have a significant effect on mass transfer.

**Backscatter**

Backscatter, i.e. an attenuation corrected value of the return signal strength produced by the ADCP, from the ADCP is shown in Figure 3.62 and SNR from the Vector in Figure 3.63. Backscatter has been corrected using estimated values for attenuation due to acoustic spreading and range dependent water absorption.
Vector SNR reflects changes in the return signal strength, but is not corrected since the acoustic path is short and thus any attenuation is minimal. Predicted peak sensitivity particle diameters are $50 \, \mu m$ for the Vector (at 10 MHz) and $400 \, \mu m$ for the ADCP (at 1200 kHz) (Lohrmann, 2001).

Both show substantial increases during Days 14-16, with a peak occurring for a few hours before decaying quickly to a fairly constant level for the remainder of the discharge. The initial higher peak is similar to the signal expected from a precipitation based runoff event as loose, available sediment is washed downstream with the increased discharge. Because this discharge event involves water from the Shandaken Tunnel (and Schoharie Reservoir), backscatter does not decay, reflecting conditions in that reservoir and the tunnel. Several smaller increases associated with increased discharge occur in each record (e.g. Days 18 and 20) and with the flow stoppages on Days 13, 21, 22, and 23.

The backscatter increase on Day 12.5 is associated with a flow stoppage and reversal (see Figure 3.51). Prior to this flow stoppage, there was an increase in discharge and backscatter peaking on Day 12. The backscatter increase associated with the flow reversal is expected to be the same water mass moved upstream by the wind induced baroclinic forcing. Similarly, just before and after Day 22 there are flow stoppages with associated increases in backscatter representing advection of downstream water back upstream to the measurement location, with the initial increased backscatter potentially due to a discharge event peaking on Day 20.

While there was increased discharge on day 20, neither the Vector nor ADCP records an increase in SNR or backscatter during this period. During the Day 21 and 22 events, the Vector SNR drops sharply to values around 15 dB while
the ADCP backscatter increases. The opposite behavior of the two acoustic signals suggests a different particle size distribution than what occurred in the discharged water.

Larger particles are expected to settle faster than smaller particles, resulting in a smaller mean suspended particle size as time increases. The reverse of what is observed with regards to the acoustic signals is then expected, since the Vector is considerably more sensitive to smaller particles than the ADCP. One possible explanation for this behavior is the strong winds just prior to Day 22 moved the
Day 20 water mass out of the Esopus Creek mouth, allowing main basin water to move in front of the mouth. This main basin water could have different physical characteristics than the discharge water (including zooplankton populations which would be large enough for the ADCP to detect). Supporting this possibility is the lower temperatures measured at the instrument during this period, suggesting primarily main basin water from around the thermocline was present. Without grab samples to examine particle distributions however, there is no direct evidence available to explain this signal.

Figure 3.63: SNR for the Vector during the Esopus Creek deployment.
Conclusions

While Esopus Creek behaves like a creek, it is still subject to baroclinic (and potentially barotropic) forcing. This forcing is generally not strong enough to stop or reverse the flow in the creek, but evidence in the turbulence data shows it weakens turbulence when acting against streamflow (suggesting it might enhance turbulence when working with streamflow). During north and northwest winds flow stoppages and reversals in the stream flow are observed, consistent with expected transport of water by the wind. This physical transport of water can generate a return flow moving water upstream. Location in the water column is expected to be controlled by density relative to the creek water and is typically observed at the bottom.

Velocity profile fits support boundary roughness playing an important role in the creek flow. Values of roughness parameters are widely distributed, but histograms of $z_0$ and $h^+$ show values are consistently above expected smooth wall limits and into the transitional flow regime. Estimates of roughness size are consistent with observational evidence from instrument deployment.

Turbulent stresses are generally 0.1-0.2 Pa, but can increase significantly during discharge events. No appreciable change in the mean flow or stress was observed during the numerous small discharge events occurring during this deployment, suggesting discharge needs to increase to $\approx 750$ cfs or mean flow needs to increase beyond 0.15 m s$^{-1}$ for an increase in stress to occur as observed on Days 14-16.

Estimates of roughness height $h$ from the boundary layer fit are $O(10^{-2}$ m). Bed features (i.e. dunes) are expected to contribute significantly to rough-
ness effects, so using $h$ as an estimate of particle size is not recommended. It likely represents the average size of bed features. Significantly, it is significantly larger than the small, turbidity causing particles with sizes from 1-10 $\mu$m citep-Peng2009 of concern in the reservoirs.

Stress levels are in the range of critical stress levels (0.08-0.20 Pa) used by Owens et al. (2011) to model resuspension in Schoharie Reservoir for particle size classes with Stokes equivalent sizes of 1.0, 3.1, and 8.1 $\mu$m. The Stokes equivalent size is based on observed settling velocities and represents the diameter of a sphere with the equivalent settling velocity. This suggests these small particles will not settle out in the streambed under typical conditions and will instead be transported into the main basin.

### 3.5.3 Main Basin

**Temperature Time Series and Spectra at the Instrument Frame**

The SBE39s mounted on the instrument frame sampled significantly faster than the two RUSS stations. Temperature time series are shown in Figure 3.64, with every 10th sample plotted after applying a 10 minute moving average filter. Fairly obvious seiche activity can be seen in the time series closely matching the predicted 11.5 hour period. There is also stratification over the 1.5 m measurement profile, with occasionally 4°C separating the top and bottom thermistors. This stratification tends to persist over the deployment period, maintaining a 2°C average difference between the top and bottom of the profile. The persistence of this temperature gradient points to minimal mixing despite the apparent motion of the thermocline, the presence of a solid boundary to generate
shear, and the potential for vertical mixing by the wind.

Figure 3.64: Temperatures measured by the four SBE39s in the West basin. Locations are 1.55 m (–), 1.18 m (- -), 0.78 m (...), and 0.38 m (–).

A temperature spectrum for each SBE39, and the ADCP magnitude and direction (discussed in the next section) is shown in Figure 3.65. A peak in the temperature spectra occurs around 12.5 hours, with a corresponding peak in the ADCP data around 13.3 hours. These closely match the expected period of 11.5 hours from the two layer model indicating seiching is a driving force for flow in the West basin.
Figure 3.65: Spectra from the main basin deployment showing a peak near the expected internal seiche period of 11.5 hours. The box encompasses periods from 10-14 hours. ADCP magnitude (–) and direction (– –). SBEs at 1.55 m (–), 1.18 m ( - -), 0.78 m (...), and 0.38 m (–•–)

**Velocity and Backscatter Profiles**

The ADCP recorded a burst every twenty minutes containing 1024 samples at 2 Hz (8.5 minutes of data). Sampling began on the hour so that bursts were at 12:00, 12:20, and 12:40 for example. Blanking distance was 0.5 m and the depth cell size was 0.05 m, resulting in a profile of approximately 1 m after processing. The Vector recorded a burst of 220 samples at 8 Hz (approximately
4 minutes of data) every 40 minutes, with the first burst starting on the hour and synchronized with the ADCP start time.

Burst averaged velocity profiles from the ADCP are examined to understand the influence of the baroclinic seiche on flow in the bottom boundary layer and the possibility of mixing and resuspension of sediment from the bed. Bins within 0.06 m of the bottom were removed from analysis due to bias created by boundary echoes and side lobe interference. This estimate was developed using the estimate provided by Teledyne-RDI of 10% of the profile length or equivalently the cosine of the beam angle, 20°, times the distance to the boundary being contaminated. It is slightly smaller than the predicted 0.09 m contamination region, having been adjusted down based on observed measurement quality. Bottom position was determined from backscatter peaks, averaged across each of the four beams.

Figure 3.66 presents the burst average magnitude and direction during the main basin deployment. Magnitudes are fairly weak, occasionally exceeding 0.05 m s\(^{-1}\) but on average less than 0.02 m s\(^{-1}\). Local bathymetry slopes down approximately to the northwest and curves to follow the basin shoreline (see Figure 3.3). It and the the connection to the East basin, a deep channel to the north, influence the flow direction, generating predominantly north-south flow rather than the expected northwest-southeast direction from a seiche traveling along the basin’s main axis.

The highest velocities are seen during flows to the north, in the direction of the East basin connection and are generally found in the upper half of the profile, occasionally approaching the bed but usually appearing no closer than 0.10-0.20 m (Days 24, 27, 30, and 32). Throughout the deployment there no
identifiable structure in the mean velocity profile characteristic of a boundary layer (i.e. a logarithmic velocity profile), and no boundary layer analysis was performed.

Figure 3.66: Burst average magnitude (top) and direction (bottom) during the main basin deployment.

Acoustic backscatter is used as a surrogate to examine particulate concentration along with the OBS-3 turbidity measurements. Profiles of backscatter averaged across all four beams and the OBS-3 turbidity time series are shown in Figure 3.67. Based on the turbidity signal, there are two fairly large increases in turbidity, Day 24 and Day 29. The backscatter signal shows similar increases at these two points in the upper half of the profile, with several additional events
not occurring in the turbidity signal (e.g. Days 25 and 31). The two turbidity events directly coincide with significant decreases in backscatter in the lower half of the water column.

![Backscatter and Turbidity Graphs](image)

Figure 3.67: (top) Backscatter recorded by the ADCP during the main basin deployment. (bottom) Turbidity recorded by the OBS-3 during the main basin deployment.

The Day 24 event is shown in detail in Figure 3.68. Coinciding with the increase in turbidity is a sharp decrease in temperature at the bottom two sensors, near zero velocities and directional scatter due to the near zero velocities. There is a small, brief increase in the $\overline{u'v'}$ Reynolds stress, but nothing which would indicate a resuspension event and the resultant increase in turbidity 2.5 hours
later. Based on temperature and backscatter this is hypolimnetic water moving into the sample region.

![Temperature vs. Time Graph]

Figure 3.68: Turbidity event detail for Day 24. Symbols are the same as in Figure 3.64 for the top plot.

On Day 24.4 there is low backscatter and colder water, but no resultant increase in turbidity. The temperature is slightly warmer than on Day 24 by approximately 2°C, suggesting this is a different water mass than what passed the instrument frame before. While this second water mass bears signatures of hypolimnion water, the lack of a turbidity increase and the warmer temperature suggests it is from higher in the water column. This suggests turbidity is strongly dependent on position in the water column.
Figure 3.69 shows the time series at the instrument frame from Days 27-30 during the second turbidity event. Coinciding with the two main turbidity increases on Days 27.5 and 28.5 are temperature decreases at the lowest sensor. Backscatter during these two turbidity increases drops at the bottom of the profile, with a much larger mass of low backscatter water during Day 28.5. Flow is to the south during the Day 28.5 increase in turbidity, while the Day 29.5 turbidity increase is oscillating flow between the north and south. These directional oscillations have a approximately two hour period, moving a mass of higher turbidity water back and forth across the sampled region that gradually decreases in peak magnitude. The short period and direction of these oscillations suggests they were an effect of basin management and the connection to the East basin was involved.

Turbidity and backscatter are inversely correlated when a turbidity increase is observed. Peng et al. (2009) showed turbidity causing particulates in Ashokan are typically fine clays in the O(1-2 |\mu|m) size range. While the OBS is sensitive to particle size, acoustic backscatter is significantly less sensitive to particle sizes below the frequency dependent peak sensitivity diameter. The maximum sensitivity for the 1200 kHz ADCP occurs for a particle of O(400 |\mu|m) diameter, while sensitive for a 1-2 |\mu|m particle is \( \frac{1}{10^8} \) of this (Lohrmann, 2001). The OBS sensitivity increases as particle size decreases and should reach maximum sensitivity for the turbidity causing particles (Downing, 2006). Thus, the hypolimnion water is potentially very low in acoustic backscatter for a 1200 kHz system, but high in optical backscatter.
Conclusions

Mean velocities during this deployment are very weak while the velocity profile has no identifiable boundary layer structure. The connection to the East basin has a fairly strong local effect on the flow at the main basin measurement location. Turbulence is extremely weak at this site. Reynolds stresses show periodic increases tied to higher velocities, but at very small levels of 0.05 Pa maximum.

Temperature and acoustic backscatter records suggest stratification largely
eliminates vertical transport and mixing, effectively separating the surface mixed layer and hypolimnion waters. Backscatter also supports minimal mixing, with strong delineation between the two water masses apparent throughout the deployment. Turbidity appears to be confined largely to the hypolimnion waters based on OBS and temperature records.

Observations at this site were for a limited period and captured no strong wind events or large increases in discharge. Increases in either of these forcings could result in appreciable increases in mixing and transport of particulates.

3.6 Schoharie Reservoir

Measurements were made in Schoharie during several different periods. Described here are measurements made August 10-17, 2004, Sept 1-9, 2004 and October 9-22, 2007. Instrumentation included ADCPs, a Nortek Vector, and UFI owned temperature loggers. The ADCPs were deployed on two different types of frames. During 2004 frames which oriented the ADCP upward looking were used. These frames were constructed of PVC and resembled small pallets, elevating the ADCP bottom approximately 0.05 m above the sediment surface. The Ashokan frame (Figure 3.45) was used during 2007. Deployment locations were made along the thalweg south of the Shandaken tunnel intake structure with a 1200 kHz ADCP in pulse coherent mode (Mode 11) and one deployment of a 600 kHz ADCP using Mode 1 capturing the entire water column velocity profile just north of the tunnel intake in September 2004.

Meteorological data was recorded by UFI maintained RUSS stations at 15 minute intervals in 2004 but was unavailable in 2007. The RUSS station in 2004
was deployed just north of the Shandaken tunnel intake and in front of the reservoir management building visible in Figure 3.2.

Discharge on Schoharie Creek is measured at a streamflow gage in Prattsville, NY (USGS gage 01350000) located approximately two miles upstream of the reservoir.

### 3.6.1 August 2004

Winds prior to and during the deployment period were extremely light and predominantly from the southwest. Maximum wind speed recorded was 7.2 m s\(^{-1}\), but the average wind speed was only 1.5 m s\(^{-1}\). Plots of wind speed and direction starting three days before data collection began are shown in Figure 3.70 with a wind rose shown in Figure 3.71. Light winds should result in smaller amplitude seiches during this deployment. Owens et al. (2011) suggests an approximately 2 day period for the baroclinic seiche for stratification conditions during the August and September 2004 deployments.

Schoharie Creek discharge and the reservoir surface elevation during the same ten day period as Figure 3.70 are shown in Figure 3.72. The reservoir was nearly full at the start of the deployment. A large discharge event on Day 13 filled the reservoir to capacity. A second event on Day 16 filled the reservoir beyond capacity, leading to elevated discharge at the Gilboa Dam and the north end of the reservoir.

A 1200 kHz Teledyne-RDI ADCP and a Nortek Vector were deployed on a bottom mounted frame looking up in the thalweg. The ADCP recorded 1200
velocity profile bursts divided into 175 0.02 m range cells (3.5 m profile length) every 30 minutes. Because of internal processing time the ping interval varied but averaged 1.4 Hz, resulting in approximately 12.5 minutes of data in each burst. The Nortek Vector sampled at 8 Hz every hour, collecting 4800 samples or six minutes of data. Battery and memory constraints limited the ADCP dataset to approximately 2.5 days. The Vector data is available for the entire 7 day deployment. Local water depth was 9-11 m.

Plots of the ADCP magnitude and direction are shown in Figure 3.73. During
this period the flow is fairly weak, generally less than 0.02 m s$^{-1}$. Directional and velocity shear is apparent throughout the deployment, with approximately 180° change in direction indicative of a baroclinic seiche. There are occasional patches of higher backscatter (Figure 3.74), such as on Days 11.25 and 12.25 near 1.5 m elevation, while the 12 hour time difference between these two patches suggests internal seicheing plays a role in their appearance.

The Vector data, measured closer to the bed at approximately 0.20 m elevation, shows more variability in magnitude than the ADCP data (Figure 3.75).
Figure 3.72: Reservoir elevation measured relative to the Gilboa Dam spillway crest (top) and Schoharie Creek discharge (bottom).

Except during slack periods caused by baroclinic forcing, flow is to the North into the main reservoir basin. Prior to the large runoff event on Day 13, there are increases in the near bed mean velocity which appear to be tied to baroclinic forcing and an increase in wind speed on Day 12. The discharge event on Day 13 results in a sustained increase in velocity for 12 hours before magnitude declines considerably, coinciding with an increase in winds to the south around Day 12-12.5. The Day 16 discharge increase results in an increase in velocity magnitude, while the second, smaller discharge event on Day 17 results in another, larger velocity increase. Velocities decay considerably in between
these two events, while winds increase briefly to the south before the second discharge increase. Despite several thousand cfs discharge volumes, moderate winds (around 4 m/s sustained) affect velocity magnitudes during this period.

Turbulence measured by the Vector during this period is shown in Figure 3.76. Intensities have been corrected for noise bias using Equation 2.1.3, resulting in background intensities $O(10^{-3}$ m s$^{-1}$). Background Reynolds stresses are near zero while dissipation levels are $10^{-9}$-$10^{-8}$ m$^2$ s$^{-3}$ with occasional $10^{-10}$ m$^2$ s$^{-3}$ levels coinciding with the slowest flows. During the discharge events, tur-
Figure 3.74: Average backscatter across the four beams of the the 1200 kHz ADCP deployed in the thalweg in August 2004.

Bulence increases significantly with much higher intensities (averaging 0.005 m s\(^{-1}\)) and non-zero Reynolds stresses peaking briefly near 0.1 Pa. Dissipation increases by 1-2 orders of magnitude, approaching 10\(^{-6}\) m\(^2\) s\(^{-3}\) levels during the discharge events.

The Vector SNR data is plotted in Figure 3.77. Prior to the Day 13 discharge event, SNR shows variation influenced by baroclinic forcing on a similar time scale as the ADCP backscatter. After Day 13, the baroclinic forcing component is seen superimposed on a linear decay of SNR which increased significantly dur-
Figure 3.75: Magnitude (–) and vertical velocity (–) during the August Schoharie deployment (top). Direction (bottom).

ing the discharge event. The linear decay of SNR suggests an exponential decay in scatterer concentration since SNR is in decibels, a logarithmic scale. The Day 16 event produces a similar peak in SNR, matching the magnitude of the Day 13 event despite its smaller discharge volume. Examination of the raw amplitude data (not shown, amplitude is a digital count of the return signal strength) shows similar peak amplitudes during these two events. Plotting SNR on a linear scale shows the Day 16 peak is almost twice as large as the Day 13 peak. Interpreting this difference in the absence of suspended sediment concentration grab samples and particle size distributions is difficult due to the dependence
of acoustic scattering strength on particle size.

3.6.2 September 2004

Winds during the deployment period had a mean speed of 1.9 m s\(^{-1}\), but saw maximum speeds of 8.5 m s\(^{-1}\) and several sustained, moderate wind speed pe-
Figure 3.77: SNR from one of the receivers of the Vector during the August deployment.

Wind speeds were predominantly around 4 m s\(^{-1}\). Wind directions were predominantly from the southwest (Figures 3.78 and 3.79). Starting on Day 5 there are sustained, moderate winds from the southwest followed by the strongest winds recorded during the deployment on Day 9. A second wind event, this time out of the northeast, occurs on Day 19, preceded by sustained but weak winds from the same direction. Because winds blow for several days with consistent speed and/or direction seiche activity is expected to be stronger then in August.

Schoharie Creek discharge and reservoir surface elevation are shown in Fig-
Figure 3.80. The reservoir was full at the start of the deployment, drawn down slightly over the first ten days and then filled to capacity by a small discharge event on Day 10. An extremely large discharge event on Day 18 filled the reservoir significantly beyond capacity for the remainder of the deployment. Owens et al. (2011) places this second discharge event as the tenth largest average daily flow in the Schoharie Creek streamflow record (beginning in 1902) at the time of publication in January 2011. The passage of Hurricane Irene and Tropical Storm Lee over the eastern New York region in Fall 2011 have likely pushed this event further down in the record.
A 600 kHz Teledyne-RDI ADCP was deployed north of the Shendaken Tunnel intake structure in the main reservoir body. Local water depth was approximately 17 m, dependent on reservoir level. This ADCP profiled the entire water column using Broadband processing (Mode 1 in the Teledyne-RDI naming scheme). Performance is good for mean velocities but not for turbulence owing to the higher noise than pulse coherent processing and large bin sizes utilized. The ADCP collected a burst of 1024 velocity profiles every hour in 0.5 m range cells.
Mean velocities (Figure 3.81, note color is on a log scale) throughout the water column are typically $O(10^{-3} \text{ m s}^{-1})$ with regular changes in direction due to a baroclinic seiche. Spectra of ADCP magnitude, flow direction, backscatter, and temperature are shown in Figure 3.82. There are two different periods suggested by this data, with magnitude and direction suggesting periods near 48 hours, while backscatter indicates a period of approximately 24 hours.

Owens et al. (2011) suggests a 48 hour period based on the 2 layer model used to predict baroclinic seiche periods in Onondaga Lake and the West Basin.
of Ashokan Reservoir in this dissertation. Wiegand and Chamberlain (1987) discusses the occurrence of a second mode vertical seiche in a three layer system, a simplification allowing simpler analysis of a continuous stratification in the hypolimnion. Extensive discussion of the theory and derivation of the predicted period for a three layer system is available in Mortimer (1952). Unfortunately, the data needed to predict the three layer system response in unavailable (namely a temperature profile allowing determination of layer depths and temperatures needed to determine densities). With a maximum depth of 40 m at the north end, comparable to the depths in Lake Windermere studied by Mortimer (1952), the bathymetry of Schoharie Reservoir is sufficiently deep to support development of a three layer system (most likely by formation of continuous stratification in the hypolimnion). Periods for the higher mode seiches predicted by Mortimer (1952) using a three layer model for Lake Windermere are approximately twice as long as for a two layer model, consistent with observations in Schoharie Reservoir. While no definitive proof exists of higher mode seiches, supporting evidence such as the velocity spectrum suggests they are probably excited within the basin.

Velocities increase significantly during the two discharge events, increasing by two orders of magnitude above the background seiche induced velocities. During the Day 10 event, the increase in velocity is confined largely to elevations of 5-10 m (equal to depths of 7-12 m). During the Day 18 event, the lower \( \frac{2}{3} \) of the water column see increases in velocity. At the start of this event and just after, at an elevation of 12 m, there are two distinct cores of higher velocity water. These two cores and the Day 10 event suggest density effects, likely caused by a temperature difference between Schoharie Creek water and the main basin water, play a roll in the fate of Schoharie Creek’s inflow.
A 1200 kHz Teledyne-RDI ADCP was deployed in the thalweg south of the Shandaken Tunnel intake structure in 11 m local water depth. The ADCP recorded a burst of 1024 profiles every hour at 2 Hz using Mode 11 (Teledyne-RDI’s pulse coherent mode). The profile is divided into seventy 0.04 m range cells for a total profile length of 3 m. Magnitude and direction are shown in Figure 3.83. Unfortunately, memory was expended prior to the Day 18 discharge event, although the smaller Day 10 event was captured. Mean velocities are $O(10^{-3} \text{ m s}^{-1})$ for most of the deployment, with direction showing the expected
Figure 3.82: Spectra for the ADCP magnitude (–), direction (––), backscatter (–) from bin 5 at an elevation of 3 m and ADCP onboard temperature (– –). The boxes represents periods of 20-25 hours and 40-48 hours.

baroclinic forcing. In contrast to the August deployment, there is significantly less directional shear in the velocity profile.

Evidence of buoyancy affecting the behavior of the Schoharie Creek inflow occurs during the Day 10 discharge event. A velocity increase occurs in the bottom 2 m of the profile on Day 10. Following this, there is a decrease in velocity in the bottom 1.5 m of the profile, while velocity remains elevated above 2 m. Temperature during this period (Figure 3.84) is useful in understanding the source
of various water masses, although it is not definitive. Examining the direction plot in Figure 3.83, just after the Day 10 discharge event, there is a small mass of water, still flowing north but with a slightly different direction. Based on temperature, this appears to be main basin water which was pushed south by the baroclinic seiche and then flushed back through the measurement region. Schoharie Creek water is approximately 1.5 °C warmer than the basin water, in support of buoyancy being at least a minor force in controlling the behavior of the Schoharie Creek inflow and accounting for the elevated velocity above 2 m, with the creek water displaced above the basin water due to the minor

Figure 3.83: Magnitude (top) and direction (bottom) measured by the 1200 kHz ADCP during September 2004 in Schoharie Reservoir.
density difference between the two. The displacement of the highest velocities away from the bed reduces stress available for mobilizing sediment, potentially having a significant impact on resuspension due to discharge events.

![Graph showing temperature variation over time](image)

Figure 3.84: Temperature measured by the ADCP in the thalweg during September 2004 in Schoharie Reservoir.

Backscatter (Figure 3.85) during this period shows the expected increase on Day 10 and a slow decay as this mass of water is flushed into the main basin and mixed with the reservoir water. The backscatter signal remains coherent, along with the temperature signal, for the remainder of the measurement period, potentially providing a means to trace this water mass and turbidity plumes entering the reservoir.
Boundary layer structure in the mean magnitude profiles is sometimes evident, permitting a least squares fit to Equations 3.2 and 3.3 as was done in Onondaga Lake and Esopus Creek. Plots of \( u_* \) and the scaling estimate are shown in Figure 3.86. Best fit values closely match scaling estimates, although many profiles do not show agreement with Equation 3.2, with only half of the profiles showing agreement.

Estimates of vertical turbulence intensity and turbulent dissipation are made using beam velocities and Taylor’s Frozen Turbulence Hypothesis. The Stacey
et al. (1999) variance method is used to estimate two of the Reynolds stress components. The ADCP during this deployment was oriented with a pair of beams approximately along and across the thalweg, allowing estimates of $u'w'$ and $v'w'$ stresses, where $u$ and $v$ are the along and cross thalweg velocity components.

Turbulence intensities are shown in Figure 3.87 and Reynolds stresses in Figure 3.88 for the ADCP dataset. Intensities show frequent near bed increases tied to the baroclinic seiche, with an increase occurring roughly every 24 hours in the bottom few bins of the profile on Days 2 through 9. Increases at the top of the
profile occur roughly 12 hours after (or before depending on the initial reference event) the near bed increases (e.g. Days 2.5, 3.5 and 4.5), also suggesting they are tied to the baroclinic seiche. The increase in turbulence intensity on Day 9 coincides with a strong wind event and the start of the Day 10 discharge event and its associated turbulence increase. Reynolds stress behavior is the largely the same as turbulence intensity, with peak stresses before Day 10 of 0.005 Pa. The Day 10 discharge event does not produce higher Reynolds stresses near the bed.

Figure 3.87: Beam (approximately vertical) turbulence intensities during September 2004 in Schoharie Reservoir.

Vector data is available from Day 1.9 - 6.8, missing both of the discharge
Figure 3.88: Reynolds stresses during September 2004 in Schoharie Reservoir, top: \(-\overline{\rho u'w'}\), bottom: \(-\overline{\rho v'w'}\).

Events and capturing primarily background flow conditions and turbulence (Figure 3.89). Velocities near the bed are higher than measurements from the 1200 kHz ADCP, but show similar variation with increases near Days 2, 3.5, and 6. Turbulence during this period is extremely weak despite these velocity increases, rarely exceeding background levels except during the Day 6 velocity increase.
3.6.3 October 2007

A final moored deployment in Schoharie was made in October of 2007 when the reservoir was drawn down 8 m (Figure 3.92). This is significantly lower than the 2004 datasets, exposing various shallows as mud flats, turning the thalweg into a continuation of Schoharie Creek and channelizing it’s flow. Met data is unavailable for this time period as the UFI maintained RUSS had been moved to another reservoir.
A temperature profile is available in the main basin from a series of Minilog temperature loggers owned and operated by UFI. Temperature profiles during this deployment are shown in Figure 3.93. The thermocline depth is approximately 10 m and remains fairly stable during this period. Strong stratification is present at the start of the deployment, with the surface gradually cooling from Day 15 on, similar to the trend seen in Onondaga Lake during this period. As the surface cools, wind mixing is able to mix out the temperature gradient at
Figure 3.91: SNR from one of the receivers of the Vector during the September 2004 deployment.

The thermocline, visualized in Figure 3.93 as the 15° isotherm moving to higher elevations after Day 15. The two layer model predicts a baroclinic seiche period of 11.5 hours, significantly shorter than 2004 because of the 8 m draw down and shorter basin length. This calculation was performed assuming average temperatures in the surface mixed layer (top 9 m) and hypolimnion (9 m and below) were 18°C and 6°C and a basin length of 5.6 km.

The 1200 kHz ADCP was deployed using the instrument frame from the summer 2007 Ashokan deployments in the thalweg. It recorded a burst of 1024
profiles every hour using Mode 11, divided into forty 0.02 m range cells, spanning a profile from approximately 0.7 to 1.5 m above the bed. Magnitude and direction are plotted in Figure 3.94. Because of the significant drawdown and an elevated discharge from Day 10-15, flow is consistently to the north in the bottom meter of the water column during this deployment. Baroclinic forcing influences magnitude throughout the profile while direction is primarily affected above 1 m. The baroclinic forcing is at times strong enough to stop and reverse flow in the bottom meter (Days 10, 19.5, and 22). Boundary layer fitting was performed and yielded $u_*$ estimate shown in Figure 3.95. Both Equation
3.2 and 3.3 produced similar $u_e$ estimates and consistent smooth wall values for roughness parameters.

Turbulence intensities are shown in Figure 3.96. Beam 3 shows the least abnormalities, i.e. no exceedingly large values near the surface or lower portion of the profile with reasonable orders of magnitude, of the four beams. Background intensities in beam 3 are near zero, but increase with the mean flow from Days 12 to 16 coinciding with elevated discharge. Beam 2 is comparable in quality, but occasionally reports very high intensity values (e.g. Day 13), while Beams 1
and 4 show contamination near the top and bottom of the profile respectively.

The battery case, a 0.30 m diameter cylinder, was mounted low on the instrument frame. Using a beam velocity spectrum from a range cell in the affected region, a peak at 0.025-0.03 Hz (33-40 second period) was identified. Based on a mean flow of 0.04 m s\(^{-1}\) and an expected critical Strouhal number value of \(St = 0.21\), where \(St = \frac{fD}{u}\), the expected vortex shedding frequency is 0.028 Hz (35 seconds). This suggests the battery case wake was affecting beam 4 significantly. A second peak in the beam velocity spectrum was identified at 0.25 Hz.

Figure 3.94: Magnitude (top) and direction (bottom) measured in the Schoharie thalweg during October 2007.
Figure 3.95: Best fit values of $u_*$ (●) and the mean flow scaling estimate (–) at the thalweg ADCP during October 2007 in Schoharie Reservoir.

This is roughly the frequency expected based on a Strouhal analysis using the width of one of the square legs of the instrument frame as a cylinder diameter. The calculated frequency is 0.17 Hz. The contamination in Beam 1 may be due to the mooring line which had significant slack to allow for an increase in reservoir surface elevation.

The Reynolds stress $-\rho u'\bar{w}'$ for the uncontaminated portion of the profile (based on turbulence intensity plots) is shown in Figure 3.97. Stress is elevated during the small discharge increase from Day 10-15 (peaking on Day 12), with
typical values during this period near 0.02 Pa. Using the definition of $u_e = \sqrt{\frac{\tau_{bed}}{\rho}}$ to estimate $\tau_{bed}$ (shown in Figure 3.99) and plotting $|\rho u'w'|$ from bin 35 (elevation 0.80 m) of the ADCP profile, stress away from the boundary is shown to be elevated above $\tau_{bed}$. There is a roughly linear decay in Reynolds stress as expected in a turbulent boundary layer (Ligrani, 1988). The instrument frame is likely contributing to the increased stress with its wake. Backscatter profiles (Figure 3.98) however, show a division in water masses in the middle of the profile throughout the deployment, but strongest from Days 12-14 coinciding with the higher stresses measured by the ADCP in this period as shown in Figures
3.6.4 Conclusions

Schoharie Reservoir’s water surface elevation is far more dynamic than Ashokan Reservoir owing to its smaller volume and significantly larger watershed relative to this volume. The dynamic reservoir elevation has the potential to significantly alter the flow within the reservoir. Baroclinic seiches, the primary forcing within the main reservoir body, are significantly altered by the

Figure 3.97: Reynolds stresses during October 2007, top: \( -\rho \bar{u} \bar{w}' \), bottom: \( -\rho \bar{v} \bar{w}' \).
Figure 3.98: Backscatter measured by the 1200 kHz ADCP during October 2007 in Schoharie Reservoir.

The reservoir level may also expose or flood various areas, creating mud flats and unarmored shoreline subject to wind wave erosion, drainage sheet flow or other potential resuspension mechanisms. A decreased reservoir level turns the thalweg, which at full capacity is fully submerged by a significant height of wa-
Figure 3.99: $\tau_{bed}$ estimated from best fit values of $u_*$ using the 1200 kHz ADCP data (■) and $|\rho u'w'|$ estimated from beam velocity data at bin 35, $z = 0.80$ m (△).

...ter, into a continuation of the Schoharie Creek stream channel. This alters the flow from primarily baroclinic driven oscillatory flow at full submergence to a superposition of baroclinic and barotropic driven flow (with the barotropic component due to the Schoharie Creek inflow), to a primarily turbulent open channel flow. However, the baroclinic forcing of the system is never fully removed since the thalweg flow is subcritical at all times (a velocity of several m s$^{-1}$ would be needed to achieve a supercritical Froude number) and connected to the main basin.
In the main basin, baroclinic forcing is the primary force driving mean flows. The full water column velocity profiles from the September 2004 600 kHz ADCP deployment show clear seiche activity (verified via a burst velocity spectrum), while also suggesting higher mode seiche activity. Large discharges, such as the Day 17-18 event in September 2004, briefly turn the reservoir into a plug flow system before baroclinic forcing quickly resumes control. During this event, baroclinic forcing is still apparent near the bed in the velocity data.

Buoyancy effects beyond the expected thermal stratification in the main basin occur frequently, directing Schoharie Creek water to a neutrally buoyant level within the reservoir (apparent on Day 10 in the 600 kHz ADCP data from September 2004). The density of Schoharie Creek water relative to the main basin water is expected to play an important role in the effect a discharge event has on the reservoir, primarily by guiding the creek water to a specific portion of the water column. As seen in September 2004 at the 1200 kHz ADCP, the Schoharie Creek inflow can be displaced from the bed, keeping velocities at the bed relatively weak and reducing the ability of elevated discharge to potentially resuspend material from the bed.

Flow in the thalweg (south of the Shendaken Tunnel intake) without significant reservoir drawdown behaves similarly to flow in the main reservoir in the absence of significant discharge as seen during August 2004 velocity records from the 1200 kHz ADCP and Nortek Vector. Baroclinic forcing is again predominant, although the thalweg acts as a minor flow constriction, leading to higher velocities near the bed as shown in the Vector data. Turbulence shows the effects of baroclinic forcing, with changes in intensity, Reynolds stresses, and dissipation shown in the Vector data, and intensity and Reynolds stress from the
1200 kHz ADCP data (August and September 2004).

During drawdown conditions, the thalweg becomes a turbulent open channel flow driven primarily by gravity and barotropic rather than baroclinic forces. Mean flows are predominantly to the north (into the basin), although the baroclinic seiche has the ability to slow and occasionally stop and reverse the flow (e.g. on Day 10 in October 2007). Backscatter, viewed as a surrogate for suspended sediment concentration, tends to increase during increases in discharge. The masses of high backscatter water remain coherent for several days after a discharge increase as they are washed back and forth by the baroclinic seiche and slowly mixed into the reservoir water.

With the exception of extremely large discharge events, such as the greater than 10000 cfs event observed in September 2004, baroclinic forcing remains the predominant force throughout the reservoir. Even with extremely large events, baroclinic forcing quickly reasserts itself within a day of the event. Reservoir behavior is complex owing to the many variables involved (wind speed, stratification, reservoir level, discharge), but will typically fall into baroclinic dominated regime or during large discharge events, a plug flow like system for brief periods. Smaller discharge events have the ability to affect portions of the reservoir, but are unable to completely stop or counteract the baroclinic forcing.

3.7 Conclusions

The various field campaigns discussed in this chapter provided observations of the behavior of medium sized basins subject to stratification and wind forcing. Baroclinic forcing drives mean flows within the main basins, generating velo-
ities $O(10^{-2} \text{ m s}^{-1})$. Despite the low energy of these systems, boundary generated shear, wind induced mixing, and instabilities and shear at the thermocline (or other water mass divisions involving tributary inflow waters) created turbulence. Turbulence intensities were shown to vary directly with mean flow strength, sometimes falling below detection limits for the instruments used. Dissipation levels were determined using isotropic turbulence relationships and varied over several orders of magnitude from $10^{-10} \text{ m}^2 \text{ s}^{-3}$ to $10^{-6} \text{ m}^2 \text{ s}^{-3}$ depending on measurement location within a basin.

Mean flows in the bottom boundary layer were shown to agree well with flat plate boundary layer theory. Boundary roughness effects were shown to be weak, with roughness Reynolds numbers barely above the threshold for transitional flow. While mean flows were found to agree well with boundary layer expectations, turbulence intensities and dissipation were shown to be poorly predicted by boundary layer scaling, with measured values below expectation. Stress levels outside of stream channels were extremely weak, suggesting internal resuspension of particles does not occur.

In Schoharie and Ashokan reservoirs, increases in discharge were shown to briefly control aspects of basin behavior and flow, but were unable to completely overwhelm baroclinic forcing. Changing reservoir water surface elevations were shown to change the physical characteristics of the basin such as length, layer depths, and bathymetry, influencing the basin response to wind forcing. A two layer model worked well to predict baroclinic forcing periods in all three basins. Evidence of a mode two vertical seiche in Schoharie Reservoir was observed, however measurements to confirm this were unavailable.

The field measurements highlight the complex nature of the BBL, but also
provide guidance on conditions for laboratory experiments examining turbulent scalar fluxes and sediment erosion.
4.1 Introduction

Scalar flux at the sediment-water interface and sediment scalar demand (e.g., oxygen or nitrate) is important in lake and reservoir management. In areas where high nutrient loads create eutrophic conditions, oxygen is depleted because of excess primary productivity (i.e., bacteria), altering the ecology of the bottom boundary layer. Sediment oxygen demand (SOD) is a common metric in BBL chemistry, with high SOD often leading to low dissolved oxygen (DO) levels and anoxic conditions. Low DO can cause fish kills, bacterial community changes and the production of undesirable compounds (e.g., hydrogen sulfide, ammonia, orthophosphate) (Beutel et al., 2007) and increased activity for heavy metals such as mercury (Mackenthun and Stefan, 1998).

Field and laboratory studies examine flux of oxygen at the sediment-water interface by profiling the concentration boundary layer (CBL, also called the diffusive boundary layer or DBL in literature). This is a region millimeters thick where scalar concentration transitions from a bulk water concentration \( C_{\text{inf}} \) to a concentration within the sediment. This concentration gradient drives scalar fluxes, while turbulent mixing will directly affect the thickness of the CBL and thus the sharpness of the driving gradient (Lorke et al., 2003b).

Lorke et al. (2003b) showed the influence of turbulence on CBL thickness using high resolution Doppler current profilers to measure the momentum bound-
ary layer and microelectrodes to measure dissolved oxygen. Mackenthun and Stefan (1998); Josiam and Stefan (1999); Jorgensen and Revsbech (1985); Hondzo (1998); O’Connor and Hondzo (2008) used in situ and laboratory measurements to show similar influence of turbulence on CBL thickness.

Horizontal advection plays a part in the flux rate (see discussion in Lorke et al. (2003b)), with non-stationary bulk concentrations ($C_{\text{inf}}$) routinely encountered and having a fairly large influence on CBL thickness. Lorke et al. (2003b) provides an estimate of 10% uncertainty in $C_{\text{inf}}$ resulting in 30% uncertainty in the thickness of the CBL. While $C_{\text{inf}}$ plays an important part in determining the CBL thickness, it is ultimately turbulence near the sediment-water interface which controls the CBL, especially in low energy systems such as the BBL of lakes and reservoirs (Lorke et al., 2003b).

Edberg and Hofsten (1973) noted a difference between in situ and laboratory measurements of SOD, reporting consistently lower values for laboratory measurements. They utilized simple magnetic stirrers to homogenize the overlying liquid in two chambers (one in situ and one laboratory) and provide the minimum 10 cm s$^{-1}$ current required to make accurate oxygen concentration measurements. They do not report mean velocities or turbulence levels in either chamber. Given the different diameters of the two chambers (the in situ system was over four times the diameter of the laboratory chamber) as well as the apparent difference in distance between the magnetic stirrers and the sediment-water interface, it is likely the two methods did not have matched mean velocity or turbulence levels.

Truax et al. (1995) compared in situ and laboratory measurements utilizing a closed loop system to recirculate water through chambers, finding sim-
ilar transfer rates when using fine sediment cores, but enhanced transfer rates with coarser sediments in the in situ systems. Mean velocities through both systems determined by unspecified measurements were approximately 0.30 m s\(^{-1}\). Different sized chambers were used, with the field chamber approximately two times larger than the laboratory chamber. The laboratory chamber contained three separate cores while the in situ chamber was flush with the bed. The complex flow around the cores likely influences measurements by creating different turbulent velocity fields. The authors attributed the differences in transfer rates to infiltration via porous media flow of oxygen rich water during the in situ measurements, a process that only occurred when the bed was composed of coarser sediments. Using planar laser induced fluorescence and PIV Reidenbach et al. (2010) showed bed roughness has a strong effect on turbulent mixing and mass transfer over rough boundaries (discussed in a more general sense by Raupach and Antonia (1991)). Their measurements showed increased pore water concentration for roughened beds under a variety of flow conditions, suggesting enhanced transfer due to higher turbulence was likely responsible for the higher transfer rate observed by Truax et al. (1995).

Beutel et al. (2007) developed a laboratory SOD measurement chamber forced by a recirculating peristaltic pump. A small jet (orifice diameter unreported) was introduced 2-3 cm above the sediment surface with the jet axis parallel to the surface. Measurements were made under quiescent, moderately (defined as jet velocities of 3-4 cm s\(^{-1}\)), and highly mixed (defined as jet velocities of 6-8 cm s\(^{-1}\)) conditions. The author points out that the facility is sub-optimal for the study due to horizontal spatial gradients of the mean flow. Results indicate an increase in SOD as mixing level increased, but no quantitative turbulence or mean velocity profile measurements are reported to allow relationship between
Lee et al. (2000) numerically model and make LDV measurements within a cylindrical chamber forced by two steady low Reynolds number jets ($Re_D=640$ where the length scale is the jet orifice diameter). They found a complex laminar flow, concluding flow structure is highly dependent on the chamber design. In particular they find a tangential, approximately axisymmetric, swirl flow just above the sediment-water interface with significant mean radial gradients in the bed stress, and hence $u_*$, with the jet velocity controlling the strength of mean flows and radial gradients.

Arenga and Lee (2005) used the Lee et al. (2000) chamber forced by peristaltic pumps to study SOD as a function of sediment characteristics and the variation of the hydrodynamic forcing. They varied hydrodynamic forcing by varying the peristaltic pump speed. Arenga and Lee (2005) rely on the previous simulation work of Lee et al. (2000), assuming laminar flow as forced by steady laminar jets, neglecting the pulsed flow characteristic of a peristaltic pump which will certainly alter chamber flow characteristics. While mean chamber velocities are compared to a semi-enclosed tidal bay, the flow within the chamber is regarded as laminar while the environmental flow will be turbulent.

Oldham et al. (2004) discuss the limitations of the above types of studies, primarily the lack of detailed hydrodynamic and turbulent characteristics reported within the chambers. They summarize the systems typically used to study benthic processes: mechanical grid stirred tanks, linear and annular recirculating flumes, and chambers stirred by radial or axial rotating impellers. Based on scaling arguments and experimental data, they develop simple expressions to estimate the friction velocity, $u_* \equiv \sqrt{\tau / \rho}$ where $\tau$ is the bed shear stress and $\rho$ is
the fluid density, for mechanically stirred systems commonly employed in SOD studies. The mechanically stirred facilities discussed by Oldham et al. (2004) all feature high mean flows which sets $u_*$ and turbulence levels.

In contrast to the enclosed chambers mentioned above, Hondzo (1998) used a linear recirculating flume and showed how changes in $u_*$ (i.e. changes in mean flow strength) influence oxygen fluxes in an open channel flow. Using a 2-D laser Doppler velocimeter (LDV), they made measurements of the Reynolds shear stress $\overline{u'w'}$ to estimate $u_*$ and derive a linear dependence of DO transfer rate on $u_*$. As discussed in Lorke et al. (2003b), the parameterization of turbulence via $u_*$ is appropriate for energetic systems, but does not work as well in low energy systems where a more direct measurement of the turbulence’s effect is needed. Lorke et al. (2003b) utilize the Batchelor length (or scale), estimated from dissipation measurements, to model mass transfer rather than a $u_*$ parameterization.

Given the low energy of typical lacustrine bottom boundary layers, the parameterization for CBL thickness presented in Lorke et al. (2003b) is more appropriate. This suggests valuable information can be obtained from experiments in an enclosed chamber, without the need to use a more complex experimental setup of a linear, recirculating flume.

4.2 Present Need

Researchers such as Hondzo (1998) and O’Connor and Hondzo (2008) worked in facilities which replicate flow structure (e.g. boundary layers) observed in natural flows such as streams and rivers. However, as seen in Chapter 3, Lorke
et al. (2003b) and other measurements in lake BBLs, the Law of the Wall (Equation 3.2) is often a poor model for flow structure, and \( u_* \), a poor parameterization of turbulence. When boundary roughness is involved, \( u_* \) will perform worse at parameterizing turbulence given the strong effects of roughness shape, size, and spacing on turbulence within the roughness sublayer. Raupach and Antonia (1991) states the turbulence to mean flow ratio can vary from 0.5 to 5 in rough wall boundary layers.

Chamber or microcosm facilities often have significant flow artifacts, e.g. radial gradients and strong mean flows, and are not calibrated to produce environmental turbulence levels (Beutel et al., 2007) or replicate flow structure (Truax et al., 1995). While open channel flumes are quite common and can often generate slow but well developed turbulent flows, they demand special considerations when conducting experiments to limit re-aeration of the overlying water. They are also typically not equipped to handle contaminated sediments such as those encountered in Onondaga Lake.

Onondaga Lake in particular requires a facility which is able to produce environmental turbulence levels based on field observations, handle mercury contaminated sediments, and allow researchers the ability to assess how changes in turbulence caused by various remediation methods will affect SOD. Auer et al. (2010) discusses proposed remediation methods which include an oxygen bubbler system and a nitrate slurry injection, which are expected to increase and decrease turbulence respectively. An open channel flow system is unfortunately not a viable option since it is expensive to modify for handling contaminated sediment cores, only permits testing of one core at a time reducing the range of conditions tested and the number of replicate experiments performed,
and would need to be modified to isolate the water surface from re-aeration. A small, low cost, turbulence chamber was developed to handle the above requirements and generate well characterized turbulence typical of the BBL of Onondaga Lake and other similar water bodies.

The turbulence chamber is modeled on systems developed by Webster et al. (2004) and Variano and Cowen (2004) which generate nearly homogenous, isotropic turbulence with near zero mean flows. The symmetric forcing employed by Webster et al. (2004) is impossible and physically unrealistic at the sediment-water interface, where the boundary imposes asymmetry. The one-sided forcing of Variano and Cowen (2004) is instead adapted for the chamber.

Turbulence is generated in a cylindrical chamber using peristaltic pumps. The pumps operate continuously but change direction randomly in a modification of the randomly actuated synthetic jet array (RASJA) of Variano and Cowen (2004, 2008). This system is easily adapted to larger or smaller facilities and different chamber geometries as needed. Multiple chambers, which are low cost and require only a few hours shop time to construct, may be run off of a single set of pumps providing scalability and replicate experiments.

Particle image velocimetry is used to quantify the hydrodynamics of the chamber, developing a calibration curve relating the peristaltic pump speed to the turbulent vertical intensity. While $\epsilon$ (or derived length scales) are more physically relevant (Lorke et al., 2003b), they depend on accurately measuring the fluctuations used to estimate turbulence intensity. Vertical turbulence intensity is thus used to characterize the facility, as it is robustly measured in the field by common acoustic velocity instruments like the Nortek Vector.
4.3 Methods and Facility Description

4.3.1 Chamber Design and Operation

The chamber consists of a section of 153 mm (6 in) outer diameter acrylic pipe with 6.35 mm (0.25 in) thick walls, a solid acrylic bottom plate with a machined groove to hold the pipe section in place, and a top plate with inlet/outlet ports (Figure 4.1). Chamber diameter was constrained by the size of sediment cores obtained in Onondaga Lake in initial studies using the chamber (Auer et al., 2010). Typical water column heights above sediment cores obtained for SOD experiments ranged from 80–120 mm (Auer, personal communication). During chamber characterization a water column height of 100 mm was used over an impermeable bed.

Cylindrical coordinates \((r, \theta, z)\) are defined such that the \(z\) axis is aligned with gravity and the chamber axis, positive upwards with \(z = 0\) at the bed. The radial \((r)\) and azimuthal \((\theta)\) coordinates lay on a circle circumscribed by the chamber wall (Figure 4.1). The jet array, mounted at \(z = 100\) mm, points in the \(-z\) direction creating a negative vertical velocity component initially. Velocities are measured in a plane located on the chamber diameter, the radial coordinate is analogous to the \(x\)-coordinate in a Cartesian system, and we adopt the notation \(u\) when discussing this velocity component. All elevations are reported relative to the bottom of the chamber.

The top plate contains six ports, spaced radially equidistant (50 mm) from one another around the top plate, 50 mm from the center of the chamber and 25 mm from the outer wall (Figure 4.1). Three peristaltic pumps (MasterFlex Model
Figure 4.1: Top: The chamber top plate showing the location of the six jet ports and the central access hole for probe insertion. Bottom: A side view of the chamber taken on an $r-z$ plane showing two opposing jet ports at the top and the coordinate system used when discussing chamber velocities. The $\theta$ component is directed into the page.
7523-70) were plumbed to the inlet/outlet ports using 6.35 mm (0.25 in) outer diameter, 3.18 mm (0.125 in) inner diameter Tygon tubing in a closed loop using opposing pairs of ports for each pump. Jet orifices lie in the $r - \theta$ plane, flush with the inner wall of the top cap of the chamber. This pump configuration was driven by the need to have a jet system that was inert to mercury, a common contaminant in Onondaga Lake sediments. Each pump can drive up to four chambers using appropriate pump heads.

Each pump is set to the same speed by the user at the start of an experiment. Pump direction is controlled by switching the state of an NPN 2N222 transistor via a MathWorks MATLAB controlled National Instruments DAQCard-6715 generating a 0–5 V signal. Direction changes utilize the Sunbathe algorithm detailed in Variano and Cowen (2008). The basic algorithm allows the user to select the mean and standard deviation of a Gaussian distribution from which the times between direction changes are drawn. The present experiments use a mean time between direction changes of 3.0 s and standard deviation of 1.0 s. Pump speed is altered to control turbulence intensity. Changing the diameter of the tubing used to plumb the system will change flow rates for a given pump speed. Any results presented here are specific to the chamber as tested.

For SOD experiments on Onondaga Lake sediments, a flushing cycle was programmed in the operating sequence allowing the water contained within the Tygon tubing loops leading to and from the peristaltic pumps to be flushed at user specified intervals. This flushing cycle was added at the request of researchers performing the SOD experiments and would not normally be included in chamber operation. As tested the system functioned in its random direction mode for two minutes before switching to a unidirectional flushing
mode for one minute, with this pattern repeating for the duration of an experiment. The random and flushing cycles are treated as two distinct phases when presenting results.

4.3.2 Velocity Measurements in the Chamber

Two-dimensional velocity fields are measured using particle image velocimetry (PIV). Image acquisition is handled by Boulder Imaging’s VisionNow software and recorded directly to disk for post-processing. The full chamber was imaged with a Dalsa-Coreco 1M30P CCD camera with a 1024 x 1024 pixel sensor, 12 µm square pixels, digitized at 12-bits per pixel, fitted with a Nikkor 50mm f/1.4 lens. A Coherent Innova 90 Argon-Ion laser was directed onto a Cambridge Technology Model 6860 galvanometer and scanned through the chamber in ~ 6 ms to illuminate the imaging plane. The field-of-view (FOV) contained the entire inner width and height of the chamber, although the chamber walls distort the images near them, removing 14 mm from each side from analysis.

A pair of images was acquired at 1 Hz with a Δt appropriate for the pump speed (Table 4.1). Values ranged from 500 ms for the lowest RPM to 10 ms for the largest. Image illumination and acquisition timing was controlled via a National Instruments PCI-6713 analog output card controlled by Mathworks MATLAB. The flow was seeded with hollow glass spheres (Potters Industries Sphericel #110P8) with a mean diameter of 11 µm and median specific gravity of 1.1. The particles are expected to passively follow the flow and were selected by drawing from the middle of a settled solution of water and particles. Stokes number for a typical flow of 5.0 x 10⁻³ m s⁻¹ (an expected typical turbulent velocity) is 3.4
indicating the particles will closely follow streamlines. Accumulation on the bed occurred, forming thin lines of particles. Secondary flow structures in the chamber or adhesion to the boundary could be responsible for the persistent formation of these lines of particles.

<table>
<thead>
<tr>
<th>Table 4.1: Δt for the pump characterization datasets</th>
</tr>
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<tbody>
<tr>
<td>Pump Speed (RPM)</td>
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<tr>
<td>Δt (sec)</td>
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</tbody>
</table>

Image pairs were processed in multiple passes on overlapping 32 x 32 pixel subwindows using the cross-correlation algorithm of Cowen and Monismith (1997) as implemented by Liao and Cowen (2005) as described in §2.3. The entire chamber width and height were imaged, with the last valid subwindow centered 18 pixels from the wall. Wall curvature affected PIV interrogation very near the wall due to image distortion.

Mean and Turbulent Flows

Horizontal homogeneity is examined by plotting a vertically averaged value of the various velocity fields in the calibration region from 10 to 20 mm above the bottom. Horizontal profiles are shown in Figures 4.2 through 4.7. The center of the imaged region is expected to lie approximately at the chamber center and is given a cross chamber distance of zero. While the mean flows show cross chamber variation, turbulence levels are uniform at all pump speeds. The mean flow variation and the large spikes in Figure 4.4 are due to jets remaining coherent rather than mixing together. Evidence of coherent jets are visible in all
profiles at pump speeds greater than 10 RPM as off center negative mean flows.

At pump speeds higher than 75 RPM, the jets impinge on the bottom based on observations made during SOD testing (Auer, personal communication). This is due to the relatively low ratio of water column height to jet orifice spacing, 2 for the tested chamber, dictated by constraints on chamber height and diameter. Variano and Cowen (2008) found a water column depth to jet orifice spacing ratio of about 6 was required to insure individual jets were stirred out by turbulence, much larger than the value here. This low ratio also leads to the observed jet signatures in Figures 4.4 through 4.7. The persistence of the jets is undesirable because of the strong mean velocities associated with a jet and the localized higher turbulence in these regions.

Because the near wall region was not measured due to distortion caused by the chamber curvature, the effect of sidewalls is not known. However they are expected to provide a no slip boundary condition and thus a decay to zero velocity.

The random and flushing operating cycles of the chamber produce different characteristic flows, with a stronger jet signature due to the continuous operation of the pumps in a single direction. Velocity time series are conditionally decomposed into random (2 minutes duration) and flushing (1 minute duration) blocks of data. Blocks did not include transition regions between each state, a typical time of 3-5 seconds. Blocks are thus contain slightly fewer samples than the block duration times the image pair acquisition rate. Each block of data is decomposed based on its block mean velocity producing a fluctuating velocity field. Conditional statistics are calculated by considering all of the data under the random or flushing cycles in a dataset. Eight cycles of the the three minute
random-flushing cycle were measured.

Conditional statistics are created by averaging across the imaged chamber region with a width of 112 mm (denoted by \( \langle \rangle \)) within the calibration region. Plots of \( \langle u \rangle, \langle u_{\text{rms}} \rangle, \langle w \rangle \) and \( \langle w_{\text{rms}} \rangle \) versus pump speed are shown in Figure 4.8. Above 10 RPM, mean velocities for flushing operation show a linear increase in magnitude with pump speed, while those for random operation have a pronounced curvature. Curves for turbulence intensity are similar between the two states but the random operation generates higher turbulence intensities above
Figure 4.3: Example profiles of $\bar{u}$ (—), $\bar{w}$ (— —), $u_{rms}$ (•), and $w_{rms}$ ( - ) for pump speed 6 RPM during random operation.

Dissipation

As shown by Lorke et al. (2003b), dissipation (and the derived Batchelor length scale) is a better predictor in low energy systems of the CBL thickness. It is important the chamber is able to produce well developed turbulence at energy levels typical of a lacustrine BBL. Using velocity fields in the calibration region, dissipation is estimated using spatial velocity spectra for each pump speed (Fig-
Figure 4.4: Example profiles of $u$ (—), $w$ (— —), $u_{rms}$ (− -), and $w_{rms}$ ( - -) for pump speed 10 RPM during random operation.

At pump speeds above 50 RPM, the spectra show an inertial subrange for $k_1 \approx 10^{-0.9}$. At lower pump speeds, there appears to be a $-\frac{5}{3}$ slope at wavenumbers near 1, with an increase in energy at the lower wavenumber where an inertial subrange is seen for higher pump speeds. Above 50 RPM the chamber is expected to be fully turbulent, while below this the results are inconclusive based on the spatial spectra.

Temporal spectra for the random operation mode are shown in Figure 4.10,
calculated at each interrogation point in the calibration region and averaged over the entire calibration region. At the lowest pump speeds, energy spikes occur at frequencies close to the mean on time of $t = 3$ s and higher harmonics. For pump speeds great than 10 RPM, these energy spikes disappear, entirely vanishing for pump speeds greater than 50 RPM. There is a $-\frac{5}{3}$ slope for pump speeds from 6 through 50 RPM. For 75 to 100 RPM the inertial subrange shifts to higher frequencies and is obscured by aliasing and noise. For low pump speeds, spatial spectra are inconclusive in determining whether turbulent flow exists, while the temporal spectra show well developed turbulence at 6 RPM.
Turbulent dissipation ($\varepsilon$) is calculated using Equation 2.9 using the spatial spectra shown in Figure 4.11. Values typical of a lacustrine BBL are seen for 10 RPM and less, with the potential for the 2 RPM data not being truly turbulent.
4.4 Calibration

Vertical turbulence intensity $\langle w_{rms} \rangle$ was chosen for calibration of the facility to environmental conditions. It is simple to measure this quantity in the field, while accurate estimation of the fluctuating velocity signal is a prerequisite to estimating dissipation and the Batchelor length scale. While many researchers use $u_*$ as a calibration or scaling parameter for scalar fluxes, as seen in Onondaga Lake (§3.4.6), turbulence is still present in the absence of an identifiable boundary layer structure, where $u_*$ estimated from the mean flow is also an inconsis-
tent predictor of turbulence levels.

The region from 10-20 mm above the bottom was chosen for the calibration region based on its proximity to the boundary and the constant vertical turbulence intensities observed there. The operational curve for turbulence intensity is shown in Figure 4.12 for random operation. Linear and quadratic fits for horizontal and vertical turbulent intensities are shown. The quadratic fit better models the data, with $r^2 > 0.99$ versus $r^2 = 0.98$ for the linear fits. In the region from 2-10 RPM (the region most representative of a typical lacustrine BBL), both
Figure 4.9: Kolmogorov scaled spatial velocity spectra for pump speeds of 2 (---), 6 (- -), 10 (···), 50 (···), 75 (---), and 100 (---) RPM. The Pao model spectrum is also plotted (---).

fits perform well.

Repeatability of results in the chamber is reasonable. Three datasets are available at 50 RPM and result in $\langle u_{rms} \rangle = 4.01, 5.05$ and 4.42 mm s$^{-1}$ and values of $\langle u_{rms} \rangle = 4.60, 5.36$ and 5.38 mm s$^{-1}$. Much of the variation is attributable to the location of the jet orifices in the top plate relative to the imaging plane and the short water column height.
Figure 4.10: Temporal velocity spectra for pump speeds of 2 (—), 6 (- -), 10 (···), 50 (− · −), 75 (—○—), and 100 (——♂——) RPM.

4.5 Discussion

The current chamber design is a balance between application constraints and operational considerations. Performance could be improved significantly by altering the chamber dimensions to allow a larger region for jet merging to occur. The peristaltic pump operation could be modified by lengthening the mean time between changes and changing the orifice layout and pump plumbing. However, in its initial form, the chamber is successful at reproducing turbulence similar to expectations for a lacustrine BBL.
At a pump speed of 10 RPM, the chamber reproduces typical values of vertical turbulence intensity and dissipation from Onondaga Lake. The asymmetric forcing of the chamber drives an observed anisotropy ratio \((\langle w_{rms}\rangle/\langle u_{rms}\rangle = 1.28\), averaged over 10–75 RPM), consistent with the results of Variano and Cowen (2008) and other facilities with asymmetric forcing. In both smooth and rough wall boundary layers, the anisotropy ratio would be slightly less than one in the near wall region where the CBL is located because horizontal turbulence intensities will be larger than vertical turbulence intensities (Ligrani, 1988; Raupach and Antonia, 1991). However, because the chamber is relatively small and this
Figure 4.12: The as tested chamber calibration curve for turbulence intensities, random operation $\langle u_{rms} \rangle$ (---) and $\langle w_{rms} \rangle$ (---). Confidence intervals are not shown for clarity.

ratio is nearly one in the chamber, the ability of anisotropy to creating variation in $C_\infty$ is expected to be minimal.

Temporal velocity spectra (Figure 4.10) show significant energy near the mean forcing period of the pumps ($t = 3$ sec). At 10 RPM, the energy concentration at the 3 second period has diminished and by 50 RPM there is no longer an increase in energy at the mean forcing period. Increasing the water column height would allow individual jets more time to merge and create a more homogeneous mean flow.
Testing the jet produced by a peristaltic pump at various speeds was used to examine the behavior of the chamber at lower speeds. Jet Reynolds numbers ($Re_j$) estimates are 280, 680 and 960 for pump speeds of 50, 75 and 100 RPM where $Re_j = \frac{U_{jet} d}{v}$, $U_{jet}$ was determined from volume flow rate measurements and the inner diameter $d$ of the tubing used to plumb the chamber. Flow visualization (Figure 4.13) of a single jet entering the chamber showed strong mixing indicative of turbulent flow beginning at 25 RPM. Below 25 RPM, the pulsed forcing of the peristaltic pump increases mixing, but truly turbulent flow does not develop. This suggests that jet interaction leads to development of turbulence at 6 and 10 RPM based on velocity spectra indicating turbulent flow.

![Jet visualizations at (clockwise from top left) 6, 10, 25, 50 RPM.](image)

Figure 4.13: Jet visualizations at (clockwise from top left) 6, 10, 25, 50 RPM.

Though conceptually similar to the random jet stirred tank of Variano and Cowen (2008) and the jet stirred tank of Webster et al. (2004), there are key differences between the chamber and these two systems. These differences affect both the flow and operation of the chamber. Aside from the reduced overall
degrees of freedom, the always on operation of the peristaltic pumps means a jet orifice is an alternating source and sink of momentum rather than just a source. While there are six orifices in the top cap, there are only three pumps used for the microcosm, reducing the degrees of freedom from 6 to nominally 3. The physically separated inlet and outlet and the reduced degrees of freedom create sources and sinks of momentum associated with each pump that are not co-located, generating a net mass transport from inlet to outlet within the microcosm, which is zero only when averaged over time or space containing a single pump’s inlet and outlet.

The performance of the low RPM operation of the chamber is important given the measured in-situ turbulence levels. At low RPM performance can be improved by tuning the mean time between direction changes to maximize turbulence and minimize mean flow. An early test of the system using a slightly larger chamber examined the dependence of turbulence intensity on mean time between direction changes. Results showed a linear increase in turbulence intensity and a linear decrease in mean flow as the mean time between direction changes increased. For continuity with ongoing SOD experiments, the mean time between direction changes was not altered for the calibration and testing presented here and remained fixed at three seconds.

Increasing the water column height will allow the jets to merge more fully. Based on the results of Variano and Cowen (2008), water column height would need to be 300 mm to achieve the water column depth to jet orifice spacing ratio of 6 to ensure individual jets fully merge. This has the added benefit of shifting the calibration curve so a higher pump speed will be needed to achieve a given turbulence intensity, potentially creating a more homogeneous velocity...
field and truly turbulent flow at lower intensities.

4.6 Conclusions

In-situ and laboratory measurements have shown the importance of turbulence in scalar fluxes, while many engineering applications need laboratory facilities to measure SOD under controlled, repeatable conditions representative of the turbulent benthic environment. Facilities meeting this need have been limited by a lack of direct measurement of turbulence quantities and no comparison to environmental turbulence levels and flows. Often, these facilities provide an estimate or simply qualitative assessments of mixing as moderate or strong.

The shear velocity, $u_\tau$, used by many researchers to characterize turbulent fluxes, is not a reliable representation of turbulence in the absence of boundary layer structure typically seen in low energy lacustrine environments. Using vertical turbulence intensity, $\sqrt{w'^2}$, as a comparison metric for laboratory and field studies captures relevant physical information available in $u_\tau$ directly (i.e. the turbulence intensity) but also permits the estimation of both dissipation and the Batchelor length scale for use in the alternative scaling proposed by Lorke et al. (2003b).

A laboratory facility was developed based on the concept of a randomly actuated array of jets using peristaltic pumps to drive the jet flow. The jet array was shown to produce turbulence levels typical of a lacustrine BBL over a homogeneous center region located from 10-20 mm above the bottom. This region was characterized for its mean and turbulent flows as well as dissipation levels estimated from velocity spectra. A calibration curve relating turbulence inten-
sity to pump speed for SOD testing was produced, allowing researchers the flexibility and control to develop SOD versus turbulence intensity curves. The chamber design is easily modified for given constraints, cheaply produced as a new facility given its minimal material content. The use of peristaltic pumps commonly found in laboratories allows multiple chambers to be run off of a single set of pumps, leading to more numerous replicate experiments as well.
Motivated by the turbidity issues in Ashokan and Schoharie Reservoirs, a series of experiments were carried out in a laboratory open channel flow to measure erosion rates of cohesive sediments, the dominant type found in both reservoir bottom sediments (Dusseau, 2008). Field observations (Chapter 3) showed weak, primarily baroclinically driven flow in the main reservoir basins while boundary layer structure was observed in the thalweg of Schoharie Reservoir and the Esopus Creek channel in Ashokan Reservoir. Boundary roughness, known to increase turbulence and enhance mass transport, played a role in the Esopus Creek channel, but was more limited in the main basin flows. Elevated discharge associated with precipitation events created distinctly different flow conditions, including elevated mean flows and stress levels. The potential for resuspension is considered to be most likely within the stream channels and thalweg during reservoir drawdown (i.e. when the thalweg will revert to an open channel flow state) and during large discharge events. Resuspension is expected to be minimally a factor under typical main basin flow conditions.

Owens et al. (2011) developed and tested a water quality model for Schoharie Reservoir incorporating a resuspension module. The resuspension module accounted for two primary sources of increased bed stress, wind waves and elevated reservoir currents, successfully simulating bed stresses measured in August 2004 by the Nortek Vector (i.e. §3.6.1). Owens et al. (2011) identifies resuspension in the thalweg as a secondary, but significant, source of particles to the water column during two specific flow conditions. The first is large discharge events when velocities and stresses are elevated regardless of reser-
voir elevation. The second potential period for resuspension identified by the model is during severe drawdown conditions when the thalweg is no longer submerged and returns to functioning as a stream channel with higher stress levels. The model predictions match expectation based on velocity measurements made in the basins and an observed lack of resuspension under most conditions.

The Owens et al. (2011) model utilizes three particle size classes, developed during testing of a turbidity model for Ashokan reservoir (Gelda and Effler, 2007a,b; Gelda et al., 2009). Gelda and Effler (2007b) showed three particle size classes provided the best balance of model performance and practicality for Ashokan Reservoir, two size classes generated slightly poorer model agreement with observations, while more than three size classes provided no clear benefit to model agreement. Because the two systems are interconnected, the three size classes, with Stokes equivalent sizes of 1.0, 3.1, and 8.3 µm, were adopted by Owens et al. (2011). The Stokes equivalent size is defined for irregular particles and represents the diameter of a sphere which would fall at the same velocity as observed for the irregularly shaped particle.

All of these particle size classes are within typical size ranges for cohesive sediments of less than ≈ 50 µm, i.e. clays and silts. The model was calibrated by comparing field observations of turbidity with the model output during various conditions as well as the previously mentioned stress comparison. Model comparisons using high discharge and severe drawdown observations of turbidity (no stress measurements were available) produced the best agreement with critical stress values ($\tau_{\text{critical}}$) for erosion of 0.08, 0.1, and 0.2 Pa for each class. The critical stress represents the lower limit for particle mobilization to occur. Stress
levels of this magnitude were rarely observed in the field under typical conditions, staying at \( O(10^{-2} \text{ Pa}) \) levels except in Esopus Creek. No data is available with direct observation during a large discharge event.

Identifying the actual \( \tau_{\text{critical}} \) for a cohesive sediment is extremely complex compared to a granular material, which is typically estimated following Shields (1936) analysis. In addition to the irregular, assymetric shapes encountered (clay particles are typically disc shaped), cohesive sediments also form amorphous flocs and various inter-particle forces increase the force necessary to initiate motion. Black et al. (2002) provides a comprehensive overview of the various challenges when working with cohesive sediments, including the potential for biologically produced extracellular polymeric substances (EPS) to glue sediment particles together. Winterwerp (2004) provides a more in depth discussion of the physics of cohesive sediments.

While attempts have been made to develop predictive models to estimate erosion rates, their applicability and use outside of the specific experiments where they were developed is limited. Bulk density has been identified by researchers as a parameter of interest in these predictive models, generally with an exponential relationship expected Jepsen et al. (1997).

Measuring \( \tau_{\text{critical}} \) is an unfortunately subjective process (Lavelle, 1987) made more difficult by the small size and numerous parameters affecting cohesive sediment stability (Huang et al., 2007). The method typically used for cohesive sediments is tracking the position of the sediment-water interface versus time at a known stress (McNeil et al., 1996). This method provides both an erosion rate and an estimate of \( \tau_{\text{critical}} \) based on whether erosion is detected. However, this method is limited by resolution of the sediment-water interface position.
and by the observation period over which experiments can be conducted. A very low erosion rate will require significantly longer to observe, particularly if the resolution for interface detection is coarse. Many experimental determinations of erosion rate have dealt with fairly high stress levels, minimum erosion scales greater than 1 mm, and observation periods of several hours. Even with improvements in sediment-water interface detection, measuring erosion rates at lower stress levels found in a lake or reservoir will be problematic because of the potential for long observation times needed to produce a detectable interface position change.

Several devices have been developed for in-situ measurements of erosion rates, including the Sea Carousel (Amos et al., 1992) and Ravens Flume (Ravens, 2007). Numerous other systems exist, but all follow the basic design of one of the two cited systems. These systems follow a fairly simple design, consisting of an enclosed chamber of some type (either a cylinder open at one end or a three walled duct), a sediment surface which forms the enclosing wall, and a method to generate a known stress. The response of the system is used to estimate erosion rates. Sea Carousel is based on the cylindrical dome while the Ravens Flume is a ducted system. Cylindrical systems use a turbulence grid or an annular flow to generate stress. The water isolated by the dome is monitored for suspended sediment concentration (SSC) and the change in SSC is used to infer erosion based on the surface area where the stress is applied.

The channel based systems work by pumping water through the system at a known rate expected to apply a specific average τ to the sediment surface based on boundary layer theory, numerical modeling, or turbulent duct flow analytical and empirical expressions. Erosion is typically inferred from suspended
sediment concentration measurements. Systems which rely on SSC to estimate erosion only measure suspended load and are not capable of bed load measurements.

Laboratory facilities have been developed to allow more detailed measurements and control of the applied stress and to estimate bed and suspended loads. These again fall into two basic designs. Annular and some straight open channel flumes are used with the entire flume bottom covered in sediment (Fukuda and Lick, 1980). More common for open channel flumes and closed duct systems, a sediment core with a known cross sectional area is mounted flush with the bottom of the flume and eroded (McNeil et al., 1996; Roberts et al., 2003). As the core is eroded, a vertical stage moves the sediment-water interface so it remains level with the flume bottom. Erosion is then determined by the change in interface position over time. These systems also allow determination of erosion rates with depth in the sediment. Erosion rates are expected to be lower deeper into a core because of the increased compaction caused by the overlying sediment.

Direct refinements of the McNeil et al. (1996) flume (Sedflume) have incorporated a larger duct (Roberts et al., 2003) (ASSET flume), and optical tracking of the sediment surface (Witt and Westrich, 2003) (SETEG flume). The ASET flume (not to be confused with the ASSET flume which is a different facility) is an application of the basic concept of Sedflume and its descendants, utilizing a turbulent open channel flow rather than a ducted flow. It incorporates real-time measurement of the sediment-water interface using an optical bed tracking system as well as acoustic Doppler velocimeters for simultaneous velocity measurements upstream and downstream of the sediment core (Lee and Wu, 2004).
Sedflume, ASSET, and SETEG flumes determine bed stress via analytical expressions for turbulent duct flow, numerical modeling or measurements without a sediment core mounted in the facility. Velocity measurements are typically not made because the sediment particles can interfere with measurement systems such as hot wires and laser Doppler velocimeters (LDV). Acoustic Doppler instrumentation, utilized in the following experiments and by Lee and Wu (2004), is robust to high SSC and interference from sediment particles, allowing data collection during erosion testing.

While Sedflume and its descendants have been widely used in engineering applications, its validity for measuring erosion rates in an unbiased manner has been questioned (Ravens, 2007; Jones and Gailani, 2009). Ravens (2007) suggests Sedflume (and similar facilities) will overestimate erosion rates in general due to the short test section length (1.35 m for Sedflume) and the small sediment core length (the last 0.15 m of SedfLume). The Ravens flume has a 1.1 m test section and overall length, with the entire bottom representing the sediment surface. Jones and Gailani (2009) suggests the comparison performed by Ravens (2007) was a comparison of two different measurements, with the Ravens flume measuring only suspended load while Sedflume measures total load (suspended plus bed load).

The question of test section length was addressed further by Ravens and Sindelar (2008) using a modular ducted flume with two test section lengths, one of 0.15 m (i.e. Sedflume) and one of 1.1 m (Ravens flume) and a cross section height and width of 0.11 and 0.13 m. Ravens and Sindelar (2008) found 35% greater erosion in the shorter test section at comparable bottom stresses for a monodisperse, fine grained quartz sediment. They concluded test sec-
tion length alone does not appear to significantly affect erosion measurements. Leading edge scour was expected to be much higher in the shorter test section but was not observed. Further discussion highlights the potentially significant effects of core handling on erosion rates, e.g. vibrations increasing compaction during collection and transport and the extrusion of the core during testing in Sedflume.

Sedflume and direct modifications (e.g. ASSET flume) utilize at 0.15 x 0.10 m test section. The ASET flume uses ASTM standard Shelby coring tubes with an outside diameter of 0.0762 m (inner diameter is unspecified). Huang et al. (2007) provides a summary of various straight channel erosion studies, with the maximum test section length specified as 0.60 m. With the exception of Ravens and Sindelar (2008) however, the dependence of erosion rates on test section length has not been investigated. While Ravens and Sindelar (2008) concluded test section length alone is not a significant factor in erosion rates, he utilized only five tests total, three with the longer test section and two with the shorter and measured fairly different behavior during all five tests.

The basic principle of Sedflume, tracking the sediment-water interface to measure erosion rate, is the most direct measurement of erosion rate possible. Defining the location of the sediment-water interface is problematic and assumes there will be uniform erosion over the core surface. This neglects the possibility of localized patches of erosion such as scour behind objects embedded in the core. Surface mapping, such as employed by Witt and Westrich (2003), is useful for measuring erosion with an irregular surface, but is limited by optical access to the facility, requiring a constraint on flow depth to minimize the optical path length. Flow depth will influence the structure of turbulence and limit
length scales in the flow (Davidson, 2004). While the exact relationship between turbulence length scales and erosion in unknown, the larger eddies in a turbulent flow contain more energy and have the potential to transport particles a further distance from the boundary, both of which can influence erosion by altering the local bed stress and particle concentration away from equilibrium.

The above erosion studies all utilize fairly high stress levels, with a typical minimum value around 0.2 Pa, in testing. This is a stress not routinely encountered in the BBL of lakes and reservoirs, although it is seen in certain instances and locations in Schoharie and Ashokan Reservoirs (e.g. stream channels). While erosion within the thalweg and creek channels undoubtedly occurs, the amount of sediment resuspension in the main basins is unknown. An 8 m long turbulent open channel flume in the DeFrees Hydraulics Laboratory was fitted with a sediment erosion apparatus to examine the behavior of reservoir bottom sediments under typical BBL conditions in the main basin and investigate erosion rates at higher stress levels in the thalweg and stream channels.

Bottom sediments from various sites in Schoharie Reservoir were collected for erosion testing in this facility. Measurements using PIV of the overlying velocity field are used to estimate the bed stress $\tau_{\text{wall}}$ while simultaneously optically tracking the location of the sediment-water interface. A pure kaolin clay (mean particle size 2 $\mu$m) was also tested to examine the behavior of a monodisperse sediment. Acoustic instrumentation, including a new profiling version of the Nortek Vectrino acoustic Doppler velocimeter, the Vectrino II, was used for higher stress measurements when optical access was degraded due to high turbidity.
5.1 Facility Description and Characterization

The 8 m flume utilized for these experiments is located in the Environmental Fluid Mechanics Teaching Laboratory section of the DeFrees Hydraulics Laboratory, School of Civil and Environmental Engineering, Cornell University. The flume has a 60 x 60 cm cross section, with maximum flow depths of approximately 50 cm. Flow is typically driven by a surface elevation gradient produced by the use of two variable speed centrifugal pumps capable of handling high suspended sediment loads. The flume side walls are made of glass, while the bottom panels are acrylic, permitting full optical access. The flume was designed and constructed by Engineering Laboratory Designs of Lake City, MN.

The inlet to the flume consists of two small mesh grids and a honeycomb grid to break up any large flow structures generated in the recirculation pipe network. A small diameter brass rod is glued to the flume bed at the start of the acrylic panels to trip the boundary layer into a turbulent state. Six meters downstream of this, a cutout in the bottom panel (Figure 5.1 can be used to mount instrumentation through and permits the installation the Sediment Erosion Apparatus discussed in §5.1.1). A 10 cm fixed weir was glued at the outlet to which either 10 or 20 cm extension could be clamped. Typical operating conditions for erosion tests involved a 20 cm weir height and 30 cm still water depth.

Because the flume was newly constructed measurements were carried out to understand the basic characteristics of the flume. Clear water boundary layer measurements utilizing PIV ($\Delta t = 0.015s$, image size approximately 140 x 140 cm, processed on a 16 x 64 subwindow grid with 50% overlap resulting in a vertical resolution of 1.12 mm or 5 plus units) produce the expected mean velocity.
profile (Figure 5.2). Turbulence and stress show the expected shape for a hy-
draulically smooth flow. The expected peak in streamwise turbulent intensity,
a decay in both turbulence components with increasing $z^+$, a decay in the verti-
cal turbulent intensity near the wall, and a constant stress region in the region
$50 < z^+ < 100$ are all observed. Integration of the Momentum Thickness integral

$$\theta = \int_0^{U_{\text{inf}}} \frac{u}{U_{\text{inf}}} \left(1 - \frac{\bar{u}}{U_{\text{inf}}} \right) dz$$

(5.1)
yields $\theta = 12.75$ and $Re_\theta = \frac{\theta U_{\text{inf}}}{v} = 1250$, where $U_{\text{inf}}$ is the free stream velocity ($U_{\text{inf}} = 112.5 \text{ mm s}^{-1}$)). This is only slightly less than the $Re_\theta = 1400$ used by Spalart (1988). Turbulence intensity and stress are slightly lower in magnitude than the DNS when normalized by $u_*$, despite the excellent agreement of the mean velocity profile, suggesting the constant stress region value of $\overline{u'w'}$ is 80% of $\tau_{\text{bed}}$ for this facility.

Figure 5.2: Moving clockwise from top left corner: Mean streamwise velocity profile (●), mean vertical velocity profile (△), streamwise (●) and vertical (△) turbulence intensities, and $\overline{u'w'}$ (●). Solid lines are the DNS result of Spalart (1988). $u_*$ was determined by fitting Equation 3.2 to the mean velocity profile without roughness effects.
5.1.1 Sediment Erosion Apparatus

The Sediment Erosion Apparatus was also designed by ELD and fits into the cutout in place of the blank acrylic plate. It consists of a small acrylic well approximately 0.24 m deep, with two pistons connected to a large flat plate, which is moved vertically by a stepper motor mounted below the apparatus. The two pistons pass through ferules to keep the electronic components below dry. The stepper motor is controlled by a computer via a serial communications port using the manufacturer’s supplied control software or an external control box with up and down buttons. Computer control can be automated via an application programming interface supplied by the manufacturer.

Acrylic cells with a false bottom are placed inside the well, secured via four screws at the corners. A series of eight thumb screws are used to prevent the sides from bowing out and leading to sediment leakage around the false bottom. The false bottom is a 0.02 m thick piece of acrylic with a groove machined around the edge. This groove holds a piece of rubber tubing and a Teflon wiper to seal against the side of the cells.

Two different size cells are available, a 23 x 23 cm square system and an 86 x 23 cm rectangular cell. The rectangular cell is used exclusively in the following experiments. Compared to similar erosion systems such as Sedflume, ASSET flume, SETEG and ASET discussed in the introduction, this cell is significantly longer and wider than other sediment test sections. The increased length is expected to minimize the potential for leading edge scour to affect measurements (Ravens and Sindelar, 2008).

A schematic drawing of the Sediment Erosion Apparatus with both end and
side views is shown in Figure 5.3. The fully assembled Sediment Erosion Apparatus including electronics and rectangular cell are shown in Figure 5.4.

5.1.2 Sediment Collection and Sediment Core Preparation

Natural sediments were collected on two different occasions from Schoharie Reservoir. The first collection using an Ekman dredge was largely unsuccessful at collecting sediment from areas away from shore due to the small dredge size. Sediments from near the shoreline were collected using a shovel. A second collection carried out by UFI was more successful and retrieved sediments from a variety of locations, largely by utilizing a larger dredge capable of penetrating further into the sediment bed. Natural sediments were stored wet in plastic garbage bag lined 5 gallon buckets. Chest freezers equipped with an automatic temperature control were used to maintain temperatures are 4°C for long term storage. One bucket was approximately enough sediment to fill the rectangular cell. The kaolin clay was sourced from Sheffield Pottery (www.sheffield-pottery.com) and is sold as EPK Kaolin.

The first sediment core was prepared outside the flume, but difficulties handling and mounting the heavy cell (approximately kg) in the flume and a catastrophic failure of the seal between the side wall and false bottom (Figure 5.5) necessitated significant modifications to the Sediment Erosion Apparatus. This failure prompted the addition of the thumb screws used to minimize bowing of the cell walls. At the same time, the seal between the wet well and the acrylic base the stepper motor and pistons were mounted to was modified to accept an O-ring and minimize leakage there.
Figure 5.3: Schematic drawings of the Sediment Erosion Apparatus.
Figure 5.4: Photos of the complete Sediment Erosion Apparatus fitted with rectangular cell.

Figure 5.5: Catastrophic failure of the false bottom to cell side wall seal.
After this initial failure of the system, sediment cores were prepared with the cell already installed in the flume. A mass of sediment was mixed with tap water (the same source used to eventually fill the flume) using a small hand mixer to ensure uniformity. After mixing, the core was allowed to settle for as little as one hour or as long as several days in an attempt to vary the sediment bulk density. The expected response was decreased erosion as settling time increased and the gravitational compaction increased.

5.1.3 Data Collection and Processing

A variety of instrumentation was used during the erosion tests, including multiple different PIV setups based around both Nd:YAG and Argon Ion lasers. Nortek Vectrino acoustic Doppler velocimeters (both a single point version and a pre-production profiling Vectrino II) were used to measure bulk flow properties. A SeaPoint Turbidity Meter (functionally similar to the D&A OBS-3 used in the field) was deployed during some tests to measure suspended sediment concentration.

A standard PIV setup is shown in Figure 5.6 and in use in Figure 5.7. The laser head was deployed on a table with various turning mirrors used to steer the beam above the flume. A light sheet was formed using a cylindrical lens with the Nd:YAG laser or a galvanometer (Cambridge Technologies Model 6860) with the Argon Ion laser. The Nd:YAG beam is also passed through a small converging lens with a fixed focal length to minimize the light sheet thickness.

Because the sediment core restricts optical access from below, the light sheet
Figure 5.6: A typical PIV setup. This is a clearwater experiment with the light sheet is directed up through the bottom.

is passed through the free surface. At lower flow rates, this is not a problem as the free surface is free of waves and turbulence. Once mean velocities increased to greater than 0.20 m s$^{-1}$, well beyond any measured velocity within the reservoir basins, free surface effects can be seen in the light sheet leading to out of plane losses in the PIV images. Free surface effects also cause the light sheet to move towards or away from the camera, affecting image calibration. To remove free surface effects and allow higher velocity flows to be used in erosion tests, an acrylic window with an upturned leading edge and short side walls (similar to a glass bottomed bucket) was placed in contact and parallel with the free surface (shown in use in Figure 5.8 and with the flume partially drained in Figure 5.9) to provide a smooth, non-refracting beam path.

Images were captured using one of three cameras, a Dalsa 1M30 (30 fps maximum rate) or 1M60 (60 fps maximum rate) or a Uniq UP-600 (60 fps maximum
Figure 5.7: A erosion test PIV setup in use. The turning mirrors are as bright spots within the image. The laser head is in the left foreground, and the beam path proceeds roughly from up from the center of the image, turns left half way to the top of the image, and then is directed onto various turning optics to form the light sheet in the background center of the image.

rate). Cameras were equipped with either a Nikon Nikor 60 mm f2.8 lens or a Nikon Nikor 105mm f1.8 lens. Because of the strong reflection off of the sediment surface, particularly when testing kaolin cores, a gradient neutral density filter was mounted at the front of the lens and positioned to attenuate light at the sediment-water interface. The two filters used provide the equivalent of closing the aperture 2 and 4 f-stops.

System timing was controlled through National Instruments analog output hardware (PCI-6229) programmed with timing signals generated using The Mathworks MATLAB Data Acquisition Toolbox. Images from all cameras were streamed to disk using Boulder Imaging’s VisionNow software suite and hardware.
Figure 5.8: The surface skimmer in use during an erosion test. Flow is from left to right.

Figure 5.9: The surface skimmer in use during an erosion test. Flow is from left to right.
PIV image pairs were processed with the Liao and Cowen (2005) implementation of the Cowen and Monismith (1997) cross-correlation algorithm. All image sets were processed on a regular grid with 32 x 32 subwindows and 50% overlap. Subwindow centers (grid nodes) were chosen to place sub windows as close to the sediment-water interface as possible without encountering the actual boundary. Tracking of the sediment-water interface was performed using the one of the intensity images generated during PIV data collection.

Velocity fields were assumed horizontally homogeneous and averaged in space and time to produce mean velocity profiles. Bed stress was estimated from the value of $u^2 = \tau_{bed}$ and from the turbulent Reynolds stress component $-\rho u'w'$ assuming a constant stress region equal to 85% of $\tau_{bed}$. Data quality control consisted of various convergence checks within the PIV code, a local median outlier filter Westerweel and Scarano (2005) on individual velocity fields, and an adaptive Gaussian filter on the combined space-time data series averaged to produce velocity profiles.

When PIV images were unavailable, typically because water clarity degraded image quality, a Nortek Vectrino deployed downstream of the test section sampling at mid-water column ($\approx 0.15$ m elevation) was used to infer $\tau_{bed}$ as 5% of the mean flow. The turbulent stress estimate is not used because of the measurement position well away from the bed and outside of the expected constant stress region.

For granular materials, the Shields number, $\theta = \frac{\tau_{bed}}{\rho_p \rho g d}$, where $\rho_p$ is the particle density, $\rho$ is the fluid density, $g$ is gravitational acceleration, and $d$ is particle diameter, is a useful parameter to understand particle behavior. Developed by Shields (1936) from dimensional arguments, the Shields parameter
non-dimensionalizes \( \tau_{bed} \) as allows prediction of sediment motion. By plotting a particle’s Shields parameter against the particle Reynolds number \( (Re_p = \frac{u_d}{\nu}) \) on a Shields diagram (Figure 5.10 permits assessment of particle motion. While not derived for cohesive sediments, the Shields parameter and diagram are used to assess the potential for particle motion and whether cohesive effects are significant (i.e. if particle motion is predicted by the Shields parameter but no erosion is observed, it is assumed cohesive effects are at play).

Figure 5.10: Shields curve taken from Shields (1936). Movement is expected for values of \( \theta \) above the crosshatched region, no movement for values below it. The width of the crosshatched region reflects the lack of a sharp or consistent transition between the two states.

5.1.4 Erosion Test Procedure

All erosion tests followed a similar procedure. A sediment core was prepared in the empty flume and allowed to settle. An acrylic lid weighed down with
approximately 40 pounds of weight covered the core and effectively isolated the surface during filling of the flume. The pumps were turned on to resuspend loose sediment while simultaneously draining the flume. This step helped improve optical clarity for the PIV measurements by removing some of the trapped sediment. The procedure was repeated several times if needed to improve optical clarity. Once the desired clarity was achieved, the flume was filled a final time to the desired still water depth.

With the cover in place, \( \Delta t \) was determined for the desired test conditions. The surface skimmer position was adjusted to contact the free surface and any last minute alignment in the laser beam path was performed. After stopping the flow, the acrylic cover was carefully removed, disturbing the sediment core as little as possible. The sediment-water interface was brought level with the flume bottom, allowing for some tilt depending on the surface preparation (skimming versus initial settling). Depending on surface condition, primarily related to the previous testing of a core and whether it was damaged by debris, a thin layer of sediment might be scraped off the surface to level the sediment-water interface and create a uniform surface. This was typically done prior to the fill and drain cycle so excess sediment could be washed out of the flume.

Depending on the flow speed being used in the erosion test, data collection was started after several minutes of the pumps running (slow speed tests) or prior to the pumps being started (high speed tests). Data from the first few minutes after startup is not typically analyzed because a changing water surface elevation altered the image calibration while free surface effects created an unusable light sheet. The startup flow period, discussed in the next section, provides qualitative insight into the processes involved in sediment resuspension.
and transport, allowing in some instances, processing select pairs with PIV to generate velocity fields suitable for identification of flow structures.

Testing continued for a set period, generally with the Vectrino and OBS being logged continuously if being used, and intermittent PIV datasets taken throughout a test cycle. Several long datasets were taken during select tests and will be discussed more fully.

5.2 Kaolin Core Erosion Tests

Several kaolin cores were tested prior to natural sediment core tests to examine the behavior of sediment consisting of a single particle size, eliminating some of the complexities of working with natural sediments. These tests also served as shakedown tests during which an experimental procedure could be developed and refined without wasting the natural sediment.

The first erosion test was performed after seven days of settling, leaving surface features formed during the settlement phase (Figure 5.11). Mean flow speeds of approximately 0.045 and 0.10 m s\(^{-1}\) were used. Data was collected for periods of approximately 10 and 20 minutes, however the flume was running continuously during this time resulting in approximately 1 hour of total stress applied to the core. These velocities were both slightly above typical velocities in the Schoharie and Ashokan reservoir basins. Resultant mean stress levels were approximately 0.5 - 1.0 \(x\) 10\(^{-2}\) Pa and 3.4 \(x\) 10\(^{-2}\) Pa (based on best fit values of \(u_c\)). For the 2 \(\mu\)m clay particles, assuming no cohesive effects, the Shields number and \(Re_p\) for the lower stress is 0.3 and 0.06, while for the higher stress these values are 1.1 and 0.12 for the higher stress. The Shields analysis suggests
a non-cohesive 2 μm particle would have probably been immobile at the lower stress and potentially mobile for the higher stress (this scaling is complicated by the fact Shields diagrams do not typically extend to such low \( Re_p \)).

No erosion occurred during these tests. Image 64 at approximately 32 seconds into the first dataset and image 2121 taken approximately one hour later in the second dataset are shown in Figure 5.12. There is no apparent change in the bed forms visible in the image.

![Figure 5.11: Surface features formed during the settling phase of the first kaolinite sediment core.](image)

This core was scraped smooth for subsequent testing, creating a flat surface with minimal features (Figure 5.13). Small ridges formed on the surface (Figure 5.13) after removal of the sediment core cover and prior to data collection. At slow speeds, these ridges persisted, but were wiped away during higher speed testing discussed below.
A second round of testing of the same core (approximately three weeks total consolidation time) using a smoothed bed was carried out two weeks after the first erosion tests. Velocities used were approximately 0.05, 0.15, 0.20, 0.30, and 0.37 m s\(^{-1}\). Testing occurred over a period of 5 hours, with varying time spent at each flow velocity. Results for the first two velocities tested were similar to the previous experiments (i.e. no observable erosion). At 0.20 m s\(^{-1}\) velocity, \(\tau_{bed}\) estimated from the best fit \(u_c\) was 7.0 \(\times 10^{-2}\) Pa, just below the critical stress predicted for the smallest size class by Owens et al. (2011). Erosion testing lasted...
Figure 5.13: (top) A smooth surface formed by scraping the sediment core with a thin metal sheet. (bottom) The same surface after filling the flume.
50 minutes at this stress.

The SeaPoint Turbidity Meter (functionally the same as the OBS-3 used in the field) was logged during this 50 minute period. PIV data, used to determine \( \tau_{bed} \) was collected for approximately 30 minutes within this period. A plot of the OBS data and a least squares curve fit of the form \( A - 1/t^p \), where \( A \) is an equilibrium SSC and \( p \) is a rate exponent, is shown in Figure 5.14. The best fit value for \( A \) is 0.0126 g L\(^{-1}\) with \( p = 0.95 \). While it is tempting to attribute this increase to erosion, the large amount of sediment trapped in the flume makes this unlikely.

![Figure 5.14: OBS record during the 50 minute long erosion test at a mean flow of 0.20 m s\(^{-1}\).]
A simple analysis of an erosion rate determined from the OBS record can be used to predict the expected sediment-water interface displacement which would need to occur to produce the increase in SSC observed. A linear fit to the OBS data for times greater than 2000 seconds, representative of the time period during which PIV images were collected, yields an estimated SSC increase of $7.5 \times 10^{-8}$ g (L s)$^{-1}$ or $2.7 \times 10^{-4}$ g (L h)$^{-1}$. The expected change in the sediment surface elevation can be estimated using this rate and the time needed to collect the PIV dataset.

The flume volume is estimated from the time taken to fill to the still water depth of 0.30 m. The water supply volume flow rate was determined by estimating the surface area of the flume above the bottom and measuring the time taken to change the surface elevation by 0.01 m several times. This resulted in a volume flow rate for the supply system of 0.5 m$^3$ min$^{-1}$ and a flume volume of 5.4 m$^3$ (5400 L) for a still water depth of 0.30 m. Assuming a density of 2600 g m$^{-3}$, this suggests over the course of one hour, a $5.9 \times 10^{-4}$ m$^3$ volume of kaolinite (assuming a density of 2600 g m$^{-3}$) would need to be put in suspension. Based on the surface area of the sediment core (approximately 0.2 m$^2$), roughly $3.0 \times 10^{-3}$ m of sediment would need to erode from the core during this period.

The PIV data obtained during this experiment covers a 30 minute period, or a $1.5 \times 10^{-3}$ m change in the sediment surface elevation. Tracking the sediment-water interface using the PIV images shows it does not move over the course of the 30 minute PIV dataset (Figure 5.15). The sediment-water interface in this plot has been determined simply by locating the maximum intensity in a column of pixels averaged over a $1.3 \times 10^{-3}$ m horizontal extent (50 pixels). This result suggests the OBS data is of no utility to estimate erosion in a system where
all sources and sinks of sediment are not known and easily monitored.

![Figure 5.15: Bed tracking results from the 30 minutes of PIV data collected during the 50 minute erosion test at a mean flow of 0.20 m s\(^{-1}\). The image has been inverted and cropped to show only the sediment-water interface. Individual intensity peaks are denoted by white •.](image)

PIV images were not collected during the final two, highest velocity tests because of free surface effects (the acrylic surface skimmer was not being utilized yet). Based on measurements from the Vectrino, mean values of \(\tau_{bed}\) obtained by estimating \(u_*\) as 5% of the mean flow are 0.15 and 0.19 Pa (nearly equal to the highest \(\tau_{critical}\) used by Owens et al. (2011)). The 0.30 m s\(^{-1}\) velocity did not erode any sediment. The sediment-water interface showed no change during
this period.

A second round of kaolin cores was tested with significantly shorter settling times, only one to two hours, but produced similar results. One test with a mean velocity of 0.30 m s\(^{-1}\) and mean stress of 0.17 Pa (estimated from both \(u_\ast\) values and the measured Reynolds stresses of PIV velocity data), did not produce a detectable change in the sediment-water interface over the course of one hour. This test did produce interesting qualitative results during the first 5 minutes of the experiment during the transition from quiescent to turbulent conditions.

On the surface of the sediment core during quiescent conditions, a very loose layer of unconsolidated sediment forms referred to as the fluff layer Schaaff et al. (2006). This layer is extremely thin, consisting entirely of freshly settled particles which are easily re-suspended owing to their small size and lack of cohesion. During the startup up phase of the flow, the fluff layer sediment is observed to form a thin sheet of suspended sediment just above the eventual sediment-water interface, showing as a semi-translucent region near the bed with higher pixel intensities.

Robinson (1991) discusses a variety of coherent motions (or vortices) formed in turbulent boundary layers. A fairly wide variety of structures, primarily what are generally called ejections (near wall fluid ejected into the outer flow), are visualized by the clay suspension as long as the near bed layer remains. Eventually, the fluff layer sediment is expended as mixing of the water in the tail box, pumps, and pipe network feeding the head box distributes the sediment throughout the water column. The fluff layer persistence is also limited by the small surface area of the sediment core. Fluff layers could form on an acrylic surface, however the surface charge likely alters the typical observed behavior over
a sediment core. It should also be noted for a fluff layer to form in the flume, fairly quiescent conditions must exist as well as sufficient suspended sediment in the near wall region capable of settling out in a relatively short period of time.

Visualization of fluff layer sediment transport into the outer flow by boundary layer structures (i.e. ejections) is shown in Figure 5.16. Visually striking, the role these structures play in mass transport is important. O’Connor and Hondzo (2008) discusses their role in scalar transport, and based on these visualizations they seem to play an important role in turbulent mass transport as well. The frequency of occurrence and intensity of ejections (along with the occurrence and intensity of sweeps moving outer fluid into the near wall region) would seem to govern the mass transfer rate in a boundary layer flow (what would normally be parametrized via the $\overline{u'w'}$ Reynolds stress or eventually an eddy diffusivity. The PIV determined velocity field is superimposed on the middle image of Figure 5.16. The velocity field has been decomposed using the image mean velocity following Adrian et al. (2000) recommendations to allow easier visualization of the vortex velocity structure.

Resuspension of non-fluff layer sediment might occur if a high speed sweep of fluid encounters the bed and briefly elevates the local stress enough to loosen sediment particles. An ejection type motion occurring close enough and soon enough after the sweep would be able to resuspend these loosened particles in a similar manner to the fluff layer sediments. A flow where these motions occur more frequently (for instance if they are tied to physical features of the boundary such as roughness elements) should then have a higher erosion rate compared to a similar flow without any features.

For a smooth boundary between a mono-disperse sediment and water, how-
ever, it appears extremely high stress levels, higher than are easily produced in a laboratory open channel flow without free surface effects (waves) are required to erode any significant quantity of sediment.

5.3 Natural Sediment Erosion Tests

Using a similar procedure to the kaolin erosion tests, several natural sediment cores were tested for erosion. Because of the minimal erosion at low speeds seen during the kaolinite tests, primarily higher velocities (0.20 - 0.70 m s$^{-1}$)
were used in these tests. Settling and consolidation time varied, beginning with longer times and gradually shortening to allow only one hour of settling and consolidation.

Erosion in natural cores was observed, but it extremely difficult to quantify because of the inhomogeneous response of the natural sediment. For instance, one of the early erosion tests was at a mean flow of approximately 0.25 m s$^{-1}$ and a mean stress of 0.15 Pa (based on the best fit $u_*$ value). For particles of 1, 10 and 50 $\mu$m diameter, sizes suggested by the grain size analysis of Dusseau (2008) to be predominant in this sediment, Shields numbers are approximately 10, 1, and 0.2 with $Re_p$ values of 0.01, 0.12, and 0.61. This suggests all three particles sizes should be mobile, while no erosion was observed.

In the region where PIV images were taken, there was no movement of the sediment-water interface (Figure 5.17, averaged over a 50 pixel, 2 mm horizontal region). However, before and after images of the core surface (Figure 5.18) show the inhomogeneous response of the surface to an applied stress (note these are from a different test series, but utilized the same sediment core and slightly higher velocities).

The heterogeneous response of the natural sediment core is due to the various size classes in the sediment (Dusseau (2008)). Despite efforts to homogenize the cores, differential settling and other forces created cores with physical differences depending on $x$-$y$ location in the core, as well as the expected differences with depth. PIV data collection was not possible during these tests due to high turbidity caused by suspended sediment from the core and trapped sediment from previous tests. Mean flows measured with the Vectrino were 0.30 - 0.80 m s$^{-1}$ (Figure 5.19), resulting in $\tau_{bed}$ in the range of 0.20 - 1.60 Pa based on estimates
Figure 5.17: Bed tracking results from the natural sediment erosion test at a mean flow of 0.25 m s\(^{-1}\). The image has been inverted and cropped to show only the sediment-water interface. Individual intensity peaks are denoted by white •.

Despite these high stresses, sections of the core show little erosion while adjacent areas have been eroded considerably. Depending on the measurement location, an erosion rate of zero or several millimeters per hour would be obtained. Erosion is confined to regions where primarily non-cohesive sediments (fine sands) occur and to the downstream wake regions of hard objects such as
Figure 5.18: (top) View from above of the surface of the natural sediment core showing a smooth, uniform surface before testing. (bottom) View from above showing the inhomogeneous response of the surface. The right photo is a closer view of the upper left corner of the core visible in the before photo.
twigs, forming fairly large and deep scour pits or channels. Based on the photos in Figure 5.20, these channels formed at the downstream end of the core and slowly progressed upstream with debris visible in the channels. The sides of these channels have very sharp edges delineating them from the laterally adjacent and largely untouched sediments. The upstream end resembles a headcut, with a sudden, sharp change in elevation resembling a small cliff.

Figure 5.19: Velocity time history of a long duration natural sediment test.

The source and cause of these eroded channels is unknown. They were not observed during the kaolinite core tests and were not observed for lower speed erosion tests. High turbidity during testing obscured any view of the sediment
surface, so the flow velocity when these channels first formed is undetermined (i.e. there are velocity measurements available, however they are of little use in answering these questions since the sediment surface was obscured). Their growth rate is also unknown for this reason.

The channels are uniformly spaced across the sediment core surface (the topmost channel was likely disrupted by the twig shown in Figure 5.20. One possibility is a local increase in stress as a result of streamwise vortices. Robinson (1991) suggests the spacing of low-speed streaks commonly observed in laboratory boundary layers should be 100 viscous length scales \( \frac{\nu}{u_*} \). Because these streaks were observed at the conclusion of the erosion testing, they can reasonably be assumed to have either formed or continued to develop during the highest velocity tested of 0.80 m s\(^{-1}\). This yields \( u_* \approx 0.04 \) m s\(^{-1}\) and the viscous
length scale \( \frac{\nu}{\kappa} \approx 5 \times 10^{-3} \) m. This is smaller than the erosion channel spacing by a factor of about 20. While this erosion doesn’t seem to be tied to low speed streaks associated with the boundary layer, streamwise streaks of sediment were observed over the acrylic bottom panels up and downstream of the sediment core, suggesting a larger scale feature of the flow was responsible.

Despite the obvious difficulties present in studying erosion of a natural core, a final core settled for only one hour was tested. Testing began with a mean flow of 0.20 m s\(^{-1}\) (\(\tau_{bed} = 0.1\) Pa) for a period of 30 minutes. There was no change in the bed position during this time. The flow was then increased to a mean speed of approximately 0.48 m s\(^{-1}\), resulting in a mean stress of \(\tau_{bed} = 0.36\) Pa, and remained at this level for the next 5.5 hours. Erosion was observed during this period, captured by optical bed tracking in a 3 hour PIV dataset. An earlier PIV dataset showed no erosion over approximately 5 minutes. The PIV dataset collection times are shown on the velocity history plot for this experiment in Figure 5.21.

Optical bed tracking is performed similar to before, with the brightest (highest intensity point) taken as the location of the sediment-water interface. Based on the physical calibration of the image with 1 pixel equal to 0.042 \(\times 10^{-3}\) m, this places the resolution of the bed position at 0.021 \(\times 10^{-3}\) m corresponding to an accuracy of 0.5 pixels. The least squares fitting of the mean velocity profile over the entire dataset suggests the flow feels the sediment-water interface approximately 1.63 \(\times 10^{-3}\) m above the average intensity peak location (defined as half of the difference between the start and end pixel locations). Allowing for roughness effects (there are various seams between flume panels, the wet well, sediment cell, sediment itself) and the error in determining the elevation offset
using only the velocity profile, this is a very reasonable result and confirms the intensity peak is a reliable estimate of the location of the sediment-water interface (even if it is potentially biased). Assuming the bias is constant in time (a reasonable assumption assuming the reflective characteristics of the sediment are constant and illumination intensity does not vary), the relative position of the sediment-water interface can be determined from an arbitrary starting location.

Figure 5.22 shows the estimated location of the sediment-water interface (rel-
ative to its start at $t = 0$) over the course of the three hour dataset. The total interface displacement is $1.88 \times 10^{-3}$ m over 3.15 hours, yielding an erosion rate of $5.97 \times 10^{-4}$ m hr$^{-1}$ at an average stress of 0.36 Pa. For the five minute data set taken prior to the three hour set, this erosion rate suggests an interface displacement of only 1 pixel.

Figure 5.22: Bed track results from the final natural sediment erosion test. Individual bed locations are plotted as • with a best fit line (– –) plotted on top of this data.

During the course of this experiment, various small dunes composed mainly of fine sand migrated through the measurement location. These result in the spikes in sediment-water interface location seen in Figure 5.22 at 25 minutes and
again at 150 minutes. These dunes were typical $2-3 \times 10^{-3}$ m high and covered a streamwise distance of approximately 0.05 m (an entire dune did not fit into the PIV field of view with a width of 0.043 m). The dunes moved fairly slowly, generally taking about two minutes to traverse the field of view. Saltation of individual particles at the downstream crest is visible in the PIV images. A series of images showing the first dune observed pass through the field of view is shown in Figure 5.23.

Figure 5.23: A sequence of four PIV images showing the passage of the first dune observed in the bed tracking experiment shown in Figure 5.22.
5.3.1 Natural Sediment Test Conclusions

While an erosion rate was measured during the last test it is difficult to draw many conclusions from it. Without multiple estimates at different stresses, no relationship between stress and erosion rate can be established. While fundamental interest in developing such a relationship exists, the challenges of measuring the expected very small changes in position of the sediment-water interface in a laboratory setting, specifically under conditions typical of the environmental flow of interest are numerous. Very long experiments are needed to allow interface displacements to exceed the potential noise levels of measurement systems, while variations in environmental conditions (such as temperature) will alter the cohesiveness of the sediment and its erodibility.

Given the results under conditions typical of the Schoharie and Ashokan basins (and their tributaries), there is little doubt in the conclusion of Owens et al. (2011) characterizing resuspension as a secondary, but potentially significant source of particles to the water column from the thalweg. Given the expected baroclinic forcing in the main basin, stress capable of resuspending the small, turbidity causing particles does not seem likely. In the thalweg, under certain conditions such as high streamflow or severe drawdown, sufficient stress will exist to resuspend these small particles. Even during these specific conditions, scour due to the wake of various objects on the bed seems far more likely to result in resuspension than a stress applied to a smooth surface.
5.3.2 Erosion tests with the Vectrino II

The Vectrino II is an evolution of the Vectrino. While there are numerous advancements in internal processing and electronics (Craig et al., 2011), from an end user perspective the distinguishing feature of this instrument is its ability to measure velocities over a short profile around the sample volume location of the Vectrino (0.05 m from the central transducer). This velocity profile is divided into 1 mm range cells, extending from 0.04-0.07 m from the central transducer, compared to the single sample volume of the Vectrino with a typical height of 0.007 m and a minimum height of 0.0025 m.

An erosion test using kaolinite was performed during validation testing of the Vectrino II to examine performance of the interleaved bottom distance measurement. Conditions for this erosion test were similar to the final test discussed in §5.2, namely a mean velocity of approximately 0.30 m s\(^{-1}\) and a pure kaolinite sediment core.

One promising feature of the Vectrino II for erosion studies, particularly once optical access is limited by turbidity, is the simultaneous collection of velocity and range to bottom data. Using a computer controlled stage (Velmex slide model SPMA 1512 P40 - S1.5, driven by an Animatics Smart Motor Model SM1720D85C Version 4.15T) with 10 \(\mu\)m repeatability of position, distance to bottom data was taken over acrylic and sand surfaces. Based on the expected stage position and the measured distance to bottom, the Vectrino II (and Vectrino, tested in a similar manner in a separate experiment) is capable of resolving the boundary position fairly precisely, with the residual error between expected and measured distance over a smooth acrylic bed of 0.33 \(x\) \(10^{-3}\) m and over a smooth sand bed of 0.40 \(x\) \(10^{-3}\) m. Plots of the expected distance versus
measured distance are shown in Figure 5.24.

Figure 5.24: (left) Measured and expected distances measured over an acrylic boundary. (right) Measured and expected distances measured over a sand bed.

Using an early pre-production unit, measurements of the mean velocity profile in a turbulent boundary layer were made and compared to collocated PIV measurements. Results are shown in Figure 5.25. The result is quite good (and has subsequently been improved as internal processing and probe calibration has been refined). The mean profile is suitable for a fit to Equation 3.2. Turbulence profiles show odd shapes but are of reasonable orders of magnitude. The odd profile shapes are a result of the small head size, calibration procedure and the complex performance of bistatic, pulse-coherent systems Zedel (2008). The center of the profile, centered approximately 0.05 m from the central transducer and corresponding to the traditional acoustic velocimeter sample volume, agree the best with the PIV data and boundary layer expectations.

A kaolinite core was prepared and allowed to settle for approximately two hours before testing began. Mean velocities were expected in the 0.30-0.40 m s\(^{-1}\) range, and the Vectrino II probe head was positioned approximately 0.07 m from the sediment-water interface to start. Velocity was profiled over a range
of 40-70 mm from the central transducer in 1 mm cells, 31 total cells. Velocities were recorded at 10 Hz to simplify post-processing. Distance to the bottom was monitored at 1 Hz and restricted to an expected range between 0.06 and 0.10 m from the central transducer.

The mean stream wise velocity profiles were fit to the Equation 3.2 to obtain $u_*$ and further estimate $\tau_{bed}$. The mean bed stress estimate from $u_*$ and the bottom position versus time are shown in Figure 5.26. Looking at the first 150 minutes of testing, there is a linear change in the elevation of the surface, re-
resulting in 4 mm of apparent erosion at an average stress of 0.30 Pa. This yields an erosion rate of $1.5 \times 10^{-3}$ m hr$^{-1}$. For comparison, the natural sediment core yield a rate of $0.6 \times 10^{-4}$ m hr$^{-1}$, thirty times less than was measured here at the same average stress. The highest kaolin test prior to this showed no erosion at a stress approximately half the value here.

![Graph of bed stress and sediment surface position](image)

Figure 5.26: Average bed stress estimated from best fit $u_*$ values (top) and sediment surface position (bottom) during the Vectrino II sediment erosion test.

At $t = 165$ minutes, the mean flow was increased. Prior to this, from $t = 130$ minutes, the bed position was fairly stable showed little change (indicating the above erosion rate estimate should actually be somewhat higher). The corre-
sponding increase in stress with this velocity increase initiates erosion again, with an estimate rate of $3.5 \times 10^{-3} \text{ m hr}^{-1}$.

These wildly different and significantly higher erosion rates are somewhat unexpected given the other erosion test results. However, in this case there was damage to the false bottom and an improper seal against the side wall, allowing sediment to leak underneath the false bottom and enhance the erosion rate. This accounts for the significantly higher erosion rate than observed during previous tests at similar stress levels. Because the instrument was on loan, a repeat of the experiment could unfortunately not be carried out.

### 5.4 Conclusions

Multiple sediment cores were tested in an attempt to determine a relationship between erosion rate and bed stress and determine specific erosion rates for reservoir bottom sediments. Testing was carried out in a specially constructed facility using a laboratory open channel flume. A sediment erosion apparatus permitted the construction of artificial cores (as opposed to cores sourced intact from the field) for testing. Various settling and consolidation periods were tested, from as little as one hour to several weeks.

Results show significant stresses are required to erode any sediment, far in excess of typical baroclinically driven flows occurring in lake and reservoir bottom boundary layers. Behavior of a mono-disperse sediment compared to a poly-disperse natural sediment was dramatically different. The mono-disperse sediment showed no erosion, even at extremely high stress levels. The natural sediment eroded, but behavior was quite different depending on the local
sediment type, with granular materials more readily eroded. Erosion also was
tied to objects in the sediment such as twigs, small pebbles or potentially sharp
gradients in sediment type. Channels formed in a sediment core at high veloc-
ity show a regular spacing, but this spacing is much larger than the spacing of
stream wise velocity streaks observed in laboratory boundary layer flows. Tying
these features to either a sediment type or a flow effect can not be accomplished
at this time.

Development of surface mapping techniques to estimate a volume erosion
rate (rather than an estimate obtained from a point or line measurement) would
account for the spatially inhomogeneous erosion observed in the natural sed-
iments. However, as most of these techniques are based around a laser and
optical data collection, their utility in a highly turbid flow characteristic of high
stress environments in the laboratory is questionable.

A novel acoustic Doppler profiler shows some promise for future erosion
tests, being more robust to suspended sediment, with sufficient accuracy and
resolution of the velocity field and boundary distance for erosion measure-
ments. It unfortunately is limited to a point distance measurement similar to
the optical tracking employed with the PIV images.

Cohesive sediment erosion is an extremely complex problem with numer-
ous challenges. While it is possible to erode sediments, limitations of current
measurement techniques do not allow identification of a specific physical mech-
anism for erosion, calling into question the validity of observed erosion rates in
natural sediments (e.g. scour seems more likely given the localization of erosion
near hard objects embedded in the sediment).
CHAPTER 6
MAPPING TURBIDITY PLUMES WITH ACOUSTIC BACKSCATTER

The field measurements detailed in Chapter 3 and Chapter 5 as well as the work of Owens et al. (2011); Effler et al. (1998, 2006a) all indicate significant discharge events are required for internal resuspension in Schoharie and Ashokan reservoirs. The behavior of the Schoharie and Esopus Creek discharges will be primarily controlled by buoyancy effects, directing it to the surface, as an interlayer flow or as a plunging, density driven current along the thalweg and reservoir bottom. In terms of potential resuspension, this last scenario would generate the highest stresses. However, regardless of location in the water column, large discharge events have the ability to mobilize significant quantities of sediment within the upstream channels which are then transported into reservoir basins. Determining the behavior, fate, and persistence of these inflow plumes is crucial to understanding the sources of turbidity within these reservoirs.

6.1 Using acoustics to monitor suspended sediment concentration

Acoustics are attractive for measurements in water because of their long range and remote sensing capabilities. There are numerous problems and limitations (succinctly described in Lohrmann (2001)) when using an acoustic device (especially one not specifically designed for acoustic backscatter measurements like an ADCP) for monitoring SSC.

For examining trends, simply correcting an amplitude profile for range at-
Attenuation is usually sufficient. This provides an instrument dependent measurement of relative particle concentration with a linear scaling over a fairly wide range of SSC (given as approximately 1–10000 mg L$^{-1}$ by Lohrmann (2001)). Developing an absolute calibration to estimate SSC is far more involved and is not typically performed because of the extensive system characterization involved.

Further complicating the use of acoustics for SSC monitoring is the frequency dependent response of a system relative to particle size. At a given acoustic frequency, systems will have a maximum sensitivity to a specific particle size at $ka = 1$, where $k$ is the acoustic wave number, $k = \frac{2\pi f}{c}$ with $f$ the acoustic frequency and $c$ the speed of sound (typically taken as 1500 m s$^{-1}$) and $a$ the particle radius. Below the maximum sensitivity radius, amplitude response is proportional to $a^{-4}$ and above the maximum sensitivity radius to $a^{-1}$.

While acoustic systems will be sensitive to a wide range of particles, they will typically be most sensitive to larger particles at typical instrument frequencies. Because all commercially available ADCPs operate at a single frequency, this puts some limitations on their utility to monitor the SSC of non-ideal particles. Particularly in the reservoirs where particle diameters of interest are $O(1-10 \, \mu m)$ with an acoustic return approximately 11 orders of magnitude below the maximum sensitivity diameters of 600 and 1200 kHz systems of 800 and 400 $\mu m$.

### 6.2 Event and Background Sampling

Boat based surveys of acoustic backscatter were conducted in the summer and fall of 2007 on both Schoharie and Ashokan Reservoirs. Surveys were per-
formed using Teledyne-RDI Workhorse Monitor ADCPs operating at 600 and 1200 kHz from a variety of vessels depending on location. The 600 kHz system was used most frequently because of its greater range (up to 50 m under ideal conditions).

Correction of the return signal intensity is performed using the following equation (Lohrmann, 2001)

\[
\text{backscatter} = \text{amplitude} \times 0.43 + 20\log_{10}(R) + 2\alpha_w R + 20R \int \alpha_p dr
\]

Where the first term is the a scaling of the return intensity to decibels (the 0.43 factor carries units of dB count\(^{-1}\), where count is a unitless intensity value), the second term account for beam spreading, the third term is attenuation due to water absorption, \(\alpha_w\) is the water absorption coefficient and \(R\) is the along beam range, and the final term is attenuation due to particle absorption and \(\alpha_p\) is the particle attenuation coefficient. Both attenuation coefficients carry units of dB m\(^{-1}\). The particle absorption term is typically neglected since it is much less than the other two correction terms for low scatterer concentrations.

A typical instrument setup used involved 0.5 m range cells and bottom tracking to determine boat speed over the ground. Boat velocities were typically 1 m s\(^{-1}\). Acoustic backscatter is corrected for range attenuation using typical values suggested by the literature for the attenuation due to acoustic spreading and water absorption (\(\alpha_w = 0.06\) for the 600 kHz system), neglecting particle attenuation. Backscatter is then averaged in each range cell across the four acoustic beams. Boat position was monitored with a Garmin GPS76 mounted directly above the ADCP, which was itself mounted on a portable frame constructed of
wood and aluminum. This frame suspends the ADCP in a downward looking manner over the side of the boat, at sufficient depth and distance from the boat to minimize contamination of velocity in the closest range cells. A picture of the frame with ADCP deployed is shown in Figure 6.1.

Figure 6.1: ADCP in the deployed position for a bottom tracking survey. The GPS is visible at the intersection of the two pipes forming the mount.

Background surveys were made along the thalweg of Ashokan Reservoir in August 2007 and Schoharie Reservoir in October 2007. Throughout the fall of 2007, no significant runoff events occurred so survey data is solely for background stream flow conditions. In addition to the acoustic backscatter measurements, in October 2007 UFI utilized a SeaBird CTD equipped with an optical sensor for monitoring beam attenuation coefficient (BAC), shown to be a reliable tracer for turbidity in the reservoirs Effler et al. (2006b). Comparisons between profiles of BAC and acoustic backscatter are made to attempt to deter-
mine a relationship between acoustic backscatter and turbidity.

### 6.3 Ashokan Background Survey

A longitudinal transect running from within the Esopus Creek channel to the dam forming the eastern wall of the West Basin was performed August 1, 2007. Three lateral transects were performed near the eastern end, in the center and near the western end just outside the Esopus Creek mouth. Transect lines are shown superimposed on bathymetry contours in Figure 6.2.

The three lateral transect backscatter contours are shown in Figure 6.3. Range cell location was not corrected, relative to the GPS position, to account for the beam angle when averaging. Near sharp changes in bathymetry (around kilometer 1.25 on transect line 1 in Figure 6.3), the beam spread results in each of the beams seeing the rising or falling bathymetry at different surface positions. This shows as slight increases in backscatter following the bathymetry contours and is more apparent at greater depths where beam spread is larger.

The surface mixed layer has similar backscatter properties on all three transect lines, with fairly uniform backscatter levels throughout this region. There is some minor structure visible in the surface mixed layer, with stronger backscatter appearing near the surface and what could be vertical mixing near the thermocline on transect line 2.

Below the thermocline, transect lines 1 and 2 show significantly weaker backscatter (consistent with measurements discussed in §3.5.3). Transect line 2 shows mid hypolimnion pockets of slightly higher backscatter. These elevated
Figure 6.2: (left) Longitudinal transect line running from within Esopus Creek to the eastern end of the West Basin. The • mark 1 km distances from the start. (right) The three lateral transect lines, numbered 1 through 3 going from east to west.
regions are potentially the influence of a small tributary inflow just north of the transect line in the small lobe. The attenuation at depth in the deepest portions of Transect Line 1 is more likely due to range attenuation than a physical change in backscatter characteristics given the depths here are near the maximum range of the instrument.

Transect line 3 shows higher backscatter below the thermocline, a consequence of both proximity to Esopus Creek and the reduced cross section and consequent higher flow velocities (i.e. more particle are in suspension because of higher turbulence and a shorter time since exiting Esopus Creek). None of the
lateral transects show obvious gradients along the transect line (i.e. the West Basin appears to be primarily 2D even with fairly strong lateral bathymetry changes).

The longitudinal transect backscatter is shown in Figure 6.4. The backscatter in the surface mixed layer is consistent with observations from the lateral transects. There is a fairly obvious increase near Esopus Creek (left end of transect) which largely disappears within 1 km of the mouth. There is a strong near surface backscatter also observed in the lateral transects. It appears this feature might be tied to Esopus Creek, with potentially some lateral variation around the 1-1.5 km where there also appears to be a small vertical gradient. A portion of this vertical gradient is due to the boat course veering into shallower waters (the sharp decrease in bathymetry at 1 km).

The thermocline shows some evidence of internal seiching, with a slight elevation from kilometer 2-3. This could also be past wind mixing given the shallow thermocline location. The connection to the East Basin also seems to influence the thermocline, with some mixing/erosion of the thermocline occurring in the eastern (right) half of the transect.

From kilometer 2-3 of the transect there is a distinct plunging backscatter plume which eventually enters the water column at kilometer 3 as a detached plume. This behavior is likely due to buoyancy effects, although it is surprising in mid-summer to see what is likely creek water behave as a plunging, density driven flow when it should be warmer than the hypolimnion water.

Despite the apparent drop in backscatter in the hypolimnion along the transect line, it is impossible to draw any definitive conclusions on turbidity from
Figure 6.4: Range corrected acoustic backscatter from the longitudinal transect. Transect started in Esopus Creek (along track distance is zero).
this dataset as no relationship between acoustic backscatter at 600 kHz and turbidity can be established. Data taken in Schoharie Reservoir along with the BAC data should permit development of this relationship.

6.4 Schoharie Surveys

Backscatter surveys were made on October 8-9 and October 22-23, 2007 on Schoharie Reservoir. The reservoir level was drawn down approximately 8 m on these dates (Figure 3.92). No surface meteorological conditions are available, but notes taken during the survey indicate light winds were present during all surveys. A minor discharge event occurred on Days 20-21, while flow on Days 8-9 discharge from Schoharie Creek was minimal.

The surveys on Days 8-9 were primarily background surveys similar to those conducted in Ashokan Reservoir in July 2007. While the entire basin was surveyed, behavior in the southern end between the Schoharie Creek mouth and the Shandaken Tunnel intake structure is of the most interest. The area to the north where the basin gradually widens and deepens is of secondary interest due to the baroclinic forcing which will routinely advect water from this area back to the intake region.

Figure 6.5 shows a transect conducted along the thalweg from within the Schoharie Creek mouth to the approximate Shandaken Tunnel intake structure. Two CTD casts equipped to collect BAC data were performed during this transect to compare acoustic backscatter and BAC profiles. These two casts are marked in Figure 6.5 by yellow •.
Figure 6.5: Aerial photo of the souther end of Schoharie Reservoir showing the Schoharie Creek to Shandaken Tunnel Intake transect line. ◇ mark 500 meter intervals of along track distance. Yellow ● mark the two CTD cast locations.
Backscatter plotted versus the along transect distance, calculated as the straight line vector length between positions reports, is shown in Figure 6.6. This survey was conducted with a Teledyne-RDI 1200 kHz Workhorse Monitor ADCP. In addition to the backscatter data, velocity data using what Teledyne-RDI calls Mode 12, a high ping rate broadband processing mode, was collected. This velocity data, while not as accurate as the Mode 11 data typically collected during moored deployments, is more accurate (owing to significantly more averaging) than the data collected during the Ashokan survey.

![Figure 6.6: Acoustic backscatter data collected during along the transect line shown in Figure 6.5. The two CTD casts including BAC measurements are shown as vertical lines near 750 and 1750 m.](image)

Near the Schoharie Creek mouth, there is a fairly strong plume of high backscatter water. A picture looking at the low flow dam taken during the sur-
vey is shown in Figure 6.7. There is a surface debris line where the creek water is actively mixing with the reservoir water. In Figure 6.6 this is the region of high backscatter around 100 m into the transect. This backscatter plume persists for the first 1000 m of the transect, probably helped by the low water level and physical constraints of the thalweg. This backscatter plume is discernible for most of the 3 km transect line between a depth of 2-4 m.

Figure 6.7: The low flow dam on Schoharie Creek before it enters the reservoir.

Profile and scatter plots of BAC and backscatter from the two casts conducted with the CTD are shown in Figure 6.8. For the profiles, the depth averaged value has been subtracted from each profile. There is unfortunately no obvious relationship between the two quantities during these two casts nor in any other cast made in Schoharie. This, as observed in the main basin deploy-
ment in Ashokan reservoir and discussed in the introduction to this chapter, is due to the minimal sensitivity of the ADCP acoustic frequency to the small, turbidity causing particles.

Figure 6.8: (top) BAC (●) versus acoustic backscatter from the 1200 kHz ADCP (△) for the two CTD casts conducted on the Schoharie Creek to Shandaken Tunnel transect. (bottom) Scatter plots of acoustic backscatter versus BAC.

The profiles shown at the top of Figure 6.8, there are obvious differences in the acoustic and optical properties of the water. The first cast BAC shows a fairly flat profile, with small increases near the surface and bottom. The backscatter however shows a strong increase (the creekwater plume) at 2 m, then decreases sharply below this. The second cast has similar behavior, with the backscat-
ter plume now occurring between 2-4 m. The BAC profile shows an increase near the bed, while the backscatter actually decreases here before being contaminated by bottom returns.

The Day 22-23 surveys yielded little better results. Numerous transects were performed in the Schoharie Creek mouth, running both along and across the thalweg. The most interesting result from these surveys was the horizontal separation of the creek and reservoir water masses. A photo from the surface (Figure 6.9 shows this clearly both by water color and by the surface debris.

![Figure 6.9: View looking across Schoharie Creek just downstream from the low flow dam. The dam is out of the frame to the right, flow is right to left. The transect shown in Figure 6.10 began in front of the far wall.](image)

Acoustic backscatter shows a similar separation of the water masses, persist-
ing throughout the water column (Figure 6.10). The transect moved east to west across the channel (Figure 6.11). High backscatter water on the east side of the channel is from Schoharie Creek, while the low backscatter water on the west side is reservoir water.

Figure 6.10: Backscatter along the transect line shown in Figure 6.11, moving from east to west (left to right).

Multiple CTD casts were conducted in this area, with locations shown in Figure 6.11. The BAC profiles from these casts are shown in Figure 6.12. The drastic change in backscatter, water color, and physical character of the water does not occur in the BAC profiles (symbols in Figure 6.12 are keyed to those marking locations in Figure 6.11). The only profile showing an increase in BAC
is located on the west side of the channel (marked by °). This cast was likely contaminated by prop wash stirring up sediment from the bed unfortunately.

6.5 Conclusions

Acoustic backscatter is a promising, but limited, means of mapping various water masses. In Ashokan and Schoharie Reservoirs, inflow creek water has some-
times drastically different backscatter characteristics from the reservoir water. Despite these differences, acoustic backscatter from an ADCP is unfortunately not useful for mapping turbidity plumes because the acoustic systems are minimally sensitive to the small turbidity causing particles. A lack of significant discharge events during the Fall 2007 field campaign provided no data on other characteristics of turbidity plumes measurable remotely with an ADCP from the surface.
CHAPTER 7
CONCLUSIONS

The interaction of turbulent flow with the sediment-water interface is a complicated topic encompassing a wide range of processes. As discussed in the introduction, flow is one of the primary forces determining the ability of organisms to use habitats for feeding, shelter, and reproduction, fluxes of dissolved compounds (nutrients, contaminants, etc.) and the flux and retention of particles (Nowell and Jumars, 1984; Lorke et al., 2003b; Mackenthun and Stefan, 1998).

Velocity measurements in the bottom boundary layer of several medium sized basins during the stratified season showed flow is generally weak and under most circumstances directly controlled by a baroclinic seiche set up by wind forcing. Short duration, strong winds $O(10 \text{ m s}^{-1})$ generated almost immediate response in the thermocline position. Moderate winds $O(5 \text{ m s}^{-1})$ occurred more frequently, and if sustained over several hours also generated seiches.

During the observation periods, mean flow strength is tied to the amplitude of internal seiches. In Onondaga Lake, which responded more to wind forcing than the two reservoirs studied, higher mean flows were measured because of this. Schoharie and Ashokan Reservoirs have periodic flow at baroclinic time scales, but weaker mean flows because seiche amplitudes were not as large. Reservoir management for drinking water supply likely contributes to the weaker response to wind, but this effect is not quantified. Importantly, observation periods were relatively short (typically 1-2 weeks) and extremely limited in terms of monitoring the various inputs and outputs to the systems. A more thorough measurement campaign is needed to fully understand the complex flow within these systems.
Periods when baroclinic forcing did not control the mean flow in the two reservoirs were associated with high discharge. The largest event captured in Schoharie Reservoir increased velocities within the main basin and way from boundaries by an order of magnitude from 0.01 m s\(^{-1}\) to 0.10 m s\(^{-1}\) (September 2004), shifting flow from baroclinically driven to a plug flow, barotropically driven system.

Turbulence at the boundaries was measured using a variety of instruments. It was predominantly low energy with dissipation O(10\(^{-8}\) m\(^2\) s\(^{-3}\)), but falling to O(10\(^{-10}\) m\(^2\) s\(^{-3}\)) during the weakest mean flows. Turbulence varied directly with mean flow strength suggesting boundary shear was the primary source. Despite near zero mean velocities during flow reversal, turbulent velocity spectra and structure functions showed the inertial subrange \(a \frac{5}{3} \) and 23 slope in the spectra and structure functions respectively). An examination of Doppler noise in turbulence statistics allowed correction of turbulence intensities and dissipation estimates for noise bias in the low energy flows encountered in the field.

A turbulent chamber for use in scalar flux studies allowing a controlled and repeatable flow was developed. This chamber provides a unique facility for researchers to study turbulent mass transport in low energy environments. It produces turbulence using an array of jets driven by peristaltic pumps, randomly changing direction to generate homogeneous turbulence in a region located 10-20 mm above the sediment-water interface. Using the Onondaga Lake measurements as guides for typical turbulence levels, the chamber turbulence levels were characterized using quantitative imaging. A curve relating pump speed to turbulence was developed.
The laboratory sediment erosion studies examined the potential for resuspension of clays, silts, and fine sands using bottom sediments from Schoharie Reservoir and a kaolin clay. The kaolin cores were used to examine behavior of a mono-disperse sediment. Based on laboratory tests of both the natural sediments and kaolin, at typical stress levels measured in the bottom boundary layer of both Schoharie and Ashokan Reservoirs, resuspension of fine cohesive particles is not expected to occur.

Vortices in the near wall region were shown to eject unconsolidated particles away from the boundary into the outer flow. Because this transport mechanism was confined to the first few minutes of an erosion test and only acted on unconsolidated particles, it is an insignificant source of erosion in the laboratory experiments. Tests of the natural cores showed small debris like twigs and small pebbles generated local scour in the objects wake. Given the large amount of debris washing into the reservoirs, this is potentially a large source of particles and resuspension, but one exceedingly difficult to quantify in the laboratory.

Attempts to use acoustic backscatter to map turbidity plumes had limited success. No clear relationship between acoustic backscatter of tributary inflow water and turbidity could be established for background flow and small discharge events observed during the fall 2007 field measurements. Larger discharge events may generate a different signature which can be measured remotely by acoustics from the surface.

7.1 Future Work

On a technical front, there are two main areas which bear continued attention.
Acoustic Doppler instrumentation was used extensively in this dissertation in both the laboratory and field. A variety of methods to screen and reduce noise bias in the measurements, particularly in low energy or sub-optimal measurement conditions were utilized.

As more and more researchers utilize these instruments, particularly instruments equipped with pulse coherent processing, a better understanding of the performance and measurement capabilities of instruments is needed, especially with respect to potential bias in turbulence statistics. Pulse coherent instruments are routinely used in non-traditional settings, such as on moving platforms, in the laboratory, and in higher energy flows, understanding their performance becomes much more important.

Quantitative imaging is an extremely powerful tool to examine flows because it can provide both Eulerian and Lagrangian perspectives on the flow. The main drawback to quantitative imaging measurements is often the amount of time needed to process a dataset to extract velocity fields. As image sensors become larger and storage becomes cheaper (allowing larger datasets to be collected), reducing processing time by parallelizing image processing and by utilizing GPU based processing will become increasingly important. The possibility of realtime (or at least near realtime) analysis is a distinct possibility with current technology. For a modest image size (1134 x 485) using native libraries for OS X, processing of an image pair through three total passes was reduced to approximately 0.6 seconds on a 32 x 32 grid with 75% overlap. In many applications, this processing speed would be adequate to process data in realtime and provide the experimentalist feedback, allowing adjustment of image pair collection parameters in a dynamic setting.
The behavior of stream inflows on all three basins is another area worth investigating. Buoyancy controls the behavior of the stream inflows, diverting it to the surface, along the thermocline or into the hypolimnion (where it may run along the bottom or within the water column. Each of these scenarios will result in a different effect on basin water quality and impact on the BBL. While there is some evidence within the Onondaga Lake South Deep dataset for a minor bottom current, no significant inflow events were observed running along the bottom where they would have the greatest effect on scalar fluxes and sediment resuspension and transport.

The effect of bed roughness to modify flow and affect scalar fluxes, despite being negligible in most of the flows measured, could easily become more significant if mean flows increased such as during a density driven bottom current. Exploring the relationship between bed roughness a turbulent mass fluxes is a promising area of work in both the field and laboratory. Experiments to further characterize field boundary layers and boundary roughness can be easily translated to the laboratory, while understanding the ability of velocity measurements to provide quantitative information on boundary roughness is work best performed in a controlled setting. To date, there has been little research done on randomly arrayed roughness elements and how their flow effects might differ from a regular array.

Finally, the use of acoustic backscatter to track turbidity plumes and water masses is certainly a worthwhile pursuit given the numerous uses of sonar systems already established. However, as demonstrated here, ADCPs are severely limited in the abilities because they are optimized for velocity measurement and not backscatter measurement. A dedicated backscatter system with a low noise
amplifier is likely needed to track the small turbidity causing particles in the reservoirs, and because the range of a system with a frequency high enough to hear echos is limited, surface surveys might not be the optimal method for this type of work. Development of a forward or side scatter system with separate transmitters and receivers (or ideally a forward and backscatter system where near field is monitored via backscatter and far field via forward scatter) might prove the optimal method since it will reduce attenuation and extend range. A control volume analysis of sediment flux would be possible by measuring a vertical and horizontal profile at either end of a channel, treating the sides and bottom as sediment sinks, and measuring the mass in and out of the ends. This approach is not without its challenges, likely requiring significant development work since no commercial systems are set up for this type of measurement.


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