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Boundary Layer Turbulence and Sediment Transport in an Inter-Tidal Mudflat Channel.

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BOUNDARY LAYER TURBULENCE AND SEDIMENT TRANSPORT
IN AN INTER-TIDAL MUDFLAT CHANNEL

A Dissertation
Submit[ed] to the Graduate Faculty of the
Louisiana State University and
Agricultural and Mechanical College
in partial fulfillment of the
requirements for the degree of
Doctor of Philosophy

in

The Department of Oceanography and Coastal Sciences

by

Lun Xu
B.E., North China Institute of Water Resources and Hydro-Power, 1982
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August, 1996
To my maternal grandmother,

who gave me her all, but got nothing from me
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Abstract

Mudflats along the west coast of South Korea are influenced by high tidal ranges and strong winter monsoonal storms, and are drained by a network of incised channels. In a selected inter-tidal mudflat channel of Namyang Bay, seven levels of 3-D turbulent velocity data, five levels of water temperature data, and two levels of sediment concentration data were collected to analyze turbulent boundary layer flow with stratification. Using time series analysis techniques, the basic turbulent parameters such as turbulent intensity (variance of $u'$), vertical momentum flux, $<w'u'>$, and variance of $w'$ were obtained. The Reynolds stress distribution obtained from the "eddy correlation" technique clearly shows the presence of a constant stress layer, the thickness of which is less than in previous results. Three methods of calculating turbulent energy dissipation rate (i.e., "Kolmogorov law", "turbulent scale", and "integration") are summarized, and three basic methods for estimating bottom shear stress (i.e., "velocity shear", "eddy correlation", and "inertial dissipation") are evaluated. The final results indicate that: 1) the shear stress estimated from the "eddy correlation" technique is less than from the other two methods; 2) the results from the "velocity shear" method are strongly related to mean flow velocities; and 3) the "inertial dissipation" method can give satisfactory results when the turbulent velocity fluctuations are measured within the constant stress layer (lower 100-150 cm) and the
corresponding Reynolds number is greater than the critical Reynolds number (Re_c=3000). In the meantime, sediment transport processes on mudflats have been of concern, which lead to formation of a new conceptual processing model. Three types of roughness length variations with bottom shear stress: proportional, constant, and reverse proportional are summarized. The sediment composition changes could exert an important influence on roughness length. Finally, a modified "inertial dissipation" method has been derived in which non-neutral stratification caused by sediment resuspension is considered. The results of this method indicate that bottom shear stress can be reduced by about 6-35% in the presence of non-neutral stratification induced by sediment resuspension.
Chapter 1

Introduction
§1.1 Background and Objectives

In marine and fluvial systems, a clear knowledge of flow in the bottom boundary layer is important in all fluid-dynamics-related fields including coastal and environmental engineering, marine geology, and estuarine ecology. However in many cases, the velocity field is influenced by bed sediment movement and flow stratification induced by suspended sediment. In these situations, bottom shear stress reflects the interactions between bed sediment and the current velocity field. The stress originating at the boundary has important effects both on the sediment bed, where it influences the erosion and deposition of sediment and the creation of bed forms, and on the water, where it determines the profile of the mean velocity within the boundary layer (Bowden & Ferguson, 1979).

The interaction of the current velocity field with the sediment distribution field by means of current shear stress has two main aspects. First, near-bed particle motions induced by bed shear stress cause a variation in the apparent roughness of the sediment bed (Smith & Mclean, 1977). On the other hand, the presence of stable stratification induced by sediment resuspension inhibits vertical momentum and mass transfer, thereby reducing the shear stress near the bottom and altering the vertical distribution of suspended sediment (Smith & Mclean, 1979; Adams & Weatherly 1981; Wiberg & Smith, 1983).

The direct measurement of bottom shear stress presents serious technical difficulties (Bowden & Ferguson, 1979; Gust, 1988). Because of those difficulties, bottom shear stress normally is derived from velocity measurements made in the bottom boundary layer. There are three methods of estimating time-averaged bed
shear stress: 1) velocity shear, 2) eddy correlation, and 3) inertial dissipation (e.g., Green, 1992). The "velocity shear" method is used most extensively. In this method, bed shear stress is determined from measurements of the near-bed mean-velocity shear through the application of the "law of the wall" (e.g., Hinze, 1975). The "eddy correlation" method exploits the Reynolds stress concept which, in the case of interest here, is the mean product of measured horizontal and vertical velocity fluctuations. The "inertial dissipation" method was introduced recently in oceanographic fields (Grant, 1984; Gross & Nowell, 1985; Huntley, 1988a; Green, 1992). Based on the spectra of turbulent velocity fluctuations, this method permits the relationship between friction velocity and energy dissipation to be established. Thus, the "inertial dissipation" method is related closely to the turbulent characteristics of the flow.

In 1926, Prandtl began a lecture as follows: "What I am about to say on the phenomena of turbulent flows is still far from conclusive. It concerns, rather, the first steps in a new path which I hope will be followed by many others". Nearly fifty years later Cebeci (1974) stated that "an understanding of turbulence is still far from being complete, and its theory lacks solid foundations as well". While the situation has improved significantly during the past 20 years, much remains to be learned of the turbulent phenomenon. Most of our understanding of turbulent flows comes from observations. The field experiments have been of great importance in advancing our knowledge of planetary boundary layers, i.e., the boundary layers at the earth surface and the ocean bottom (Gluhovsky & Agee, 1994).

Based on this background, I have attempted to obtain a better insight into: 1) features of boundary layer turbulence and methods for estimating the bottom shear
stress, 2) features of sediment movement under different conditions, and 3) interactions between the current velocity field and suspended sediment. Using available data, the focus of this study is on the determination of bottom shear stress under conditions of non-neutral stratification. The major objectives are to: 1) evaluate the three methods for estimating the bottom shear stress, with emphasis on the "inertial dissipation" method, 2) understand sediment transport mechanisms and processes on a high-tide-range mudflat, 3) develop an improved technique for evaluating the effect of sediment resuspension on bottom shear stress.

§1.2 Data information

The west coast of Korea is fronted by broad intertidal sand and mudflats which have formed in a high-tide-range environment that is subjected periodically to intense winter-season storm surges from monsoonal winds (Wells et al., 1992; Adams et al., 1992). The data for this study were acquired in an inter-tidal channel of Namyang Bay, an embayment on the west coast of South Korea. Namyang Bay has an area of about 60 km². Broad mudflats that ring the bay are incised by numerous distributary channels. There are two reasons for choosing this study area. First, there exists a broad mudflat composed of marine sediments uncontaminated by river sediment input. Second, with the high tidal range, there are numerous well-developed distributary channels allowing the role of distributary channels on mudflats to be studied easily.

Primary environmental aspects of the Namyang Bay study area are as follows:

1. Fluvial sediment input is small.
2. There exists a high-tide-range environment (4-9 m) which is subjected periodically to intense winter season storm surges from monsoonal winds.
3. Drainage channels and small tributary channels are well developed on the mudflat. The larger intertidal channels are 2-5 km long, 1-5 m deep, and 10-100 m wide.

4. Sediment accumulation rate on the mudflat is high (0.15–2.0 cm/yr). Intertidal and shallow subtidal sediments range from <40% to >90% mud (Wells et al., 1990).

Data were acquired at three measuring stations along a major distributary channel (see Fig. 1-1). The information presented in this dissertation is based on data from the two stations at the lower region of the channel (station 2 and 3). Data from station 2 are one-month-long time series data with a 20 minute sampling interval, which will be used mainly for the analysis of sediment transport. Data from station 3 are composed of seven levels of high frequency velocity fluctuations, five levels of temperature fluctuations and two levels of sediment concentrations, which will be used to examine turbulent characteristics and buoyancy effects of suspended sediment separately. A more detailed data description will be presented in each chapter.

§1.3 Organization of the manuscript

The organization of this dissertation is as follows:

Each chapter begins with a discussion of the background and objectives for that particular chapter followed by presentation of data acquisition methods. The core of each chapter is the presentation of results which also includes derivation of relevant physical concepts and analyses of data. Finally, each chapter is closed with a discussion and summary. Appendices are in the last section of the dissertation.
Fig. 1-1 Study area in Namyang Bay, west coast of South Korea. The three sampling stations are along a distributary channel.
The arrangement of each chapter is as follows:

Included in Chapter 1 is a brief description of the study area, objectives and organization of the dissertation. Turbulent current velocity fluctuation data are introduced in Chapter 2 and are supplemented by a discussion of the characteristics of the three methods for estimating bottom shear stress. The features of sediment movement and associated transport processes on the mudflat are the principal topic of Chapter 3. The results of Chapters 2 and 3 will be used in Chapter 4 to examine the effects of sediment resuspension on the determination of shear stress in the bottom boundary layer. Finally, Chapter 5 will present the conclusions of this study.
Chapter 2

Characteristics of turbulence and levels of bottom shear stress in a high tidal range inter-tidal channel
§2.1 Introduction

On the west coast of South Korea, there are broad intertidal sand and mudflats which have formed in a high-tide-range environment (Wells et al., 1990). The mudflats are incised by numerous channels that act as drainage and sediment dispersal routes, primarily during ebb flows (Adams et al., 1990a). Because the mudflat channels play an important role in sediment transport, a study of turbulent characteristics in the channel is imperative for understanding current energy transformations and for determining bottom shear stress under the action of wind waves, tidal currents and sediment resuspension. With these types of interactions, the turbulent characteristics of the channel currents should be crucial.

The effect of bottom shear stress on the vertical velocity structure in a bottom boundary layer has been long recognized (Smith & Long, 1976; Grant, 1984). With the action of tides, waves and current stratification effects, both the vertical velocity structure and turbulent energy distribution will be altered. The turbulent velocity data are composed of full frequency band information which includes residual current, tidal, wave and turbulent fluctuation data. Through spectral analysis, these components can be shown either in the frequency domain or wave number domain (assuming frozen turbulence). Recent studies of turbulent flow in tidal currents include those by George (1994), Green (1992), Huntley (1988a,b), and Gross & Nowell (1985).

§ 2.2 Methods

In chapter 1, three methods for estimating time-averaged bed shear stress were introduced. They have different characteristics. For the "velocity shear" method, the
bottom shear stress is determined from measurements of near-bed velocities and application of a wall-law model:

\[ u = \frac{u_* \ln \frac{z}{z_0}}{k} \quad (2-1) \]

where \( u \) is the flow velocity at height \( z \) above bottom; \( k \) is the von Karman constant; \( u_* \) is the shear velocity; and \( z_0 \) is the roughness length. When several levels of flow velocity in the lower portion of the turbulent boundary layer are obtained, \( u_* \) and \( z_0 \) can be calculated from Eq. (2-1) using regression analysis. Finally, the bottom shear stress, \( \tau_0 \), can be determined from the relationship, \( \tau_0 = \rho u_*^2 \) where \( \rho \) is the density of water. Although this method is most widely used, it is valid only in steady, neutral, unstratified flow and in the absence of waves. Also, velocity data must be obtained within the logarithmic layer, a lower portion of the turbulent boundary layer.

The "eddy correlation" technique requires accurate measurements of horizontal and vertical velocity fluctuations in the constant stress layer, the lower portion of the logarithmic layer (e.g., Huntley, 1988a). The time-mean of the product of those fluctuations is the Reynolds stress, \( <w' u'> \), which is related to bed shear stress by the expression

\[ \tau_0 = \rho <w' u'>, \quad (2-2) \]

where \( u' \) and \( w' \) are the horizontal and vertical velocity fluctuations, and the brackets represent an ensemble average. This method is useful when waves are present, but may
suffer from misalignment of the current meter with the components of the flow when wave orbital velocities are superimposed on a steady flow (Huntley 1988a).

The "inertial dissipation" method makes use of the relationship between the friction velocity and dissipation rate of kinetic energy, namely,

\[ \tau_e = \rho (k z \varepsilon)^{2/3} \]  \hspace{1cm} (2-3)

where \( \varepsilon \) is the dissipation rate of kinetic energy. This method was first proposed by Deacon (1959) and realized by Taylor (1961) in meteorological fields. Later, this method was extensively applied in meteorology by Smith (1967), Weiler and Burling (1967), Miyake et al. (1970), Smith (1970), Pond et al. (1971), Hicks and Dyer (1972), Stegen et al. (1973), Dyer (1975), and Leavitt and Paulson (1975). Champagne et al. (1977) reviewed and further developed this method. Grant et al. (1984), Gross & Nowell (1985), and Huntley (1988b) applied the method to oceanic bottom boundary layers. The inertial dissipation technique depends only on a small band of wave numbers within the inertial subrange and is insensitive to noise or any signal outside that band (Grant et al., 1984).

To date, the relative efficacy of these three methods has not been well documented. The results of Gross & Nowell (1985) indicated that the estimation of shear velocity, \( u_* \), based on the dissipation rate derived from the magnitude of the inertial subrange of the spectra, agreed to within 10 percent with estimates of the shear velocity from the mean velocity profiles. Johnson et al. (1994) found that the dissipation measurements appear to underestimate the stress, and the magnitudes of the
stress from the dissipation estimates are smaller than those of the logarithmic method by about a factor of 3.

The work in this chapter includes the following three aspects: The first is to describe the basic characteristics of turbulent velocity data in a high tidal range intertidal channel. The second is to describe and assess the three techniques for estimating dissipation rate. The last is to evaluate the three methods for estimating the bottom shear stress based on data collected on the mudflat (station 3).

§ 2.3 Data acquisition

In this chapter, only data from station 3, located near the channel mouth, are considered. The measurements were made with the Boundary Layer Interaction Profiling System (BLIPS) (Adams et al., 1990b). This system includes the following components:

1. Eight pairs of Marsh-McBirney electromagnetic current meters (EMCM) with 2-axis, 0.1-sec response time, for measuring fluctuating velocity data.
2. Five thermistors to measure the vertical temperature profile. The thermistors have resolution of about 0.01°C, accuracy of 0.04°C and operating range of -5°C to 45°C. The sensor location distances from the channel bed are 35, 70, 125, 175 and 200 cm.
3. The channel turbidities were measured with Seatech 5-cm path length beam transmissometers at z=100 cm and z=200 cm. These short path length transmissometers provide accurate suspended sediment concentrations over the range of 1-250 mg/l.
4. A Paroscientific quartz pressure transducer was installed near bottom (z=30 cm) to measure water depth. The configuration of BLIPS sensors is shown in Table 2-1.
<table>
<thead>
<tr>
<th>No</th>
<th>Z(cm)</th>
<th>Velocity</th>
<th>Temperature</th>
<th>Sediment</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>10</td>
<td>$u', w'$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>25</td>
<td>$u', w'$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>35</td>
<td></td>
<td>T1</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>45</td>
<td>$u', w'$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>70</td>
<td>$u', w'$</td>
<td>T2</td>
<td>C1</td>
</tr>
<tr>
<td>6</td>
<td>100</td>
<td>$u', w'$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>125</td>
<td>$u', v'$</td>
<td>T3</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>150</td>
<td>$u', w'$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>175</td>
<td></td>
<td>T4</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>200</td>
<td>$u', w'$</td>
<td>T5</td>
<td>C2</td>
</tr>
</tbody>
</table>
Data measurements at station 3 extended from 12:00 May 15 to 9:00 May 27, 1992 (The data from 4:00 May 17 to 13:00 May 17 are missing). At the beginning of each hour, measurements were taken four times per second (sampling frequency = 4 Hz) for 8 min-32 sec producing a total of 2048 data points. Scan time of the sensor was 1/8 second. This measuring time period is less than 12 minutes, which is the longest record whose stationarity would permit valid calculation of all of turbulent velocity fluctuation (Soulsby, 1980).

Under the action of a cyclic tidal current, there exist more or less unsteady features in turbulent velocity data. Thus the data have to be processed before further analyses. This procedure can be summarized as follows:

a). De-bug—Remove data spikes that are usually much larger or smaller than adjacent data. In this study, 3-std cutoff was used.

b). De-mean—Reduce all data sets to have a zero mean value.

c). De-trend—Remove the residuals with longer periods by high-pass filtering.

In this study, each sampling period is 512 seconds and the sampling frequency is 4 Hz. The cutoff frequencies for band pass filtering have been defined as follows:

i. $f > 4 \times 4/2048 = 7.81 \times 10^{-3} \text{ s}^{-1}$ for de-trending ($T=128$ s)

ii. $f > 4 \times 40/2048 = 7.81 \times 10^{-2} \text{ s}^{-1}$ for removing long period residuals ($T=12.8$ s)

iii. $f > 4 \times 512/2048 = 1.0 \text{ s}^{-1}$ for removing short period residuals ($T=1$ s)

Fig. 2-1, shows the results from a typical application of this filtering. The detailed procedure of filtering is presented in Appendix A.
Fig. 2-1. A typical example of the raw velocity fluctuation data processing procedure: 1) De-bug, 2) De-mean, and 3) High-pass.
§ 2.4 Results

2.4.1 Spatial and temporal variation of mean velocity data.

As mean velocity values are necessary for analysis of turbulence, the computations of this important quantity were made for each of the seven levels. Fig. 2-2 shows a typical time variation of vertical mean velocity profiles. The mean velocities are closely coupled with the variation of tidal level and show a typical standing wave situation (Fig. 2-3). The logarithmic velocity profile formulation was used extensively to express the vertical velocity profile. By fitting seven levels of measured mean flow velocities to a log-velocity profile, Eq. (2-1), one can determine that the friction velocity, \( u_* \), and channel bed roughness length, \( z_0 \), in the bottom boundary layer. Bottom shear stress is directly related to friction velocity by the relation \( \tau_o = \rho u_*^2 \). The results indicate that friction velocity is tied closely to tidal level variation (see Fig. 2-4). For roughness length, however, a strong relationship was not obvious. Basically, the roughness length is somewhat inversely proportional to the variation of the mean current velocity. In the next chapter, this phenomenon will be discussed in more detail. Here, owing to its extensive acceptance, the "shear velocity" method is mainly used to compare with other calculations of bottom shear stress. The mean friction velocity \( u_* \) is 1.47 cm/s; and mean roughness length is 0.10 cm.

2.4.2 Energy spectrum of turbulent data

The spectral analysis technique has been fundamental to turbulence research. The turbulent energy spectrum is composed of large- and small-length-scale
Fig. 2-2 A typical tidal cycle was selected for the indication of velocity profile (top figure), mean flow velocity (mid-figure) and tidal level (bottom figure) variations at station 3 during 12:00 - 20:00 May 15, 1992.
Fig. 2-3 Temporal variation of tidal level (upper Figure) and mean flow velocity at 100 cm height (lower figure) at station 3 during 14:00 5/19 - 9:00 5/27, 1992 (n=188 hours).
Fig. 2-4 Temporal variation of friction velocity, $u_*$ (upper figure) and roughness length, $z_0$ (lower figure) at station 3 during 14:00 5/19 - 9:00 5/27, 1992 (n=188 hours).
components. The large-length-scale (or low wave-number) part produces most of the
turbulent energy; dissipation rate here is small. Large-length-scale turbulence is
anisotropic and depends strongly on the mean current flow. The small-length-scale part
(or high wave-number part) produces little turbulent energy, however, dissipation rate
is large. Because small-scale turbulence is relatively homogeneous and isotropic, it can
be treated as independent of the mean current and the analysis thus is simplified
(Tennekes & Lumley, hereinafter T&L, 1972). Since the study by Grant (1962), little
effort has been devoted to a study of turbulence in inter-tidal channels.

The energy spectrum was defined as the Fourier transform of the
autocorrelation. Thus the spectrum $\phi_{ij}(k)$, can be given by

$$\phi_{ij}(k) = \frac{1}{(2\pi)^3} \int \int \int \exp(-ik \cdot r) R_{ij}(r) \, dr,$$

where the correlation $R_{ij}(r)$ is

$$R_{ij}(r) = u_i(x, t) u_j(x+r, t),$$

with the assumption of homogeneity; $k$ is the wave number; $r$ is the spatial lag; $u_i$ and
$u_j$ are the tensor expressions of three components of flow velocity. The above equation
is referred to as the 3-D spectrum. At high wave numbers, small eddies tend to have
about the same size in all directions (T&L, p250). Because of the extreme complexity
of the 3-D spectrum, a 1-D spectrum is extensively used in practice, namely,

$$\phi_{ii}(k) = \frac{1}{\pi} \int R_{ii}(x, 0, 0) \, e^{-ikx} \, dx.$$
The 1-D spectrum is usually expressed in radian wave number domain rather than frequency domain. In this case, Taylor's hypothesis (frozen-turbulence approximation) (Hinze, 1959) is introduced to convert from the frequency domain to the wave-number domain if the turbulent velocity fluctuation is much less than the mean flow velocity (i.e., \( u'/U << 1 \)). This relation can be expressed as:

\[
\phi(k) = \frac{\phi(f)}{2\pi fU},
\]

where wave-number \( k = 2\pi f/U \); \( f \) is the frequency; and \( U \) is the mean flow velocity. The value of the spectrum at a given frequency or wave-number is the mean energy at that scale.

The "inertial dissipation" method relies strongly on the existence of the "inertial subrange", the range of wave numbers for which \( \phi \propto k^{53} \) is valid (T&L, p265). The physical meaning of the "inertial subrange" can be explained as the range in which no energy is added by the mean flow and no energy is taken out by viscous dissipation, so that the energy flux across each wave number is constant (T&L, p267). From the available data, a typical series of data (12:00-20:00, May 15, 1992) were selected for spectral analysis. Using the above-mentioned technique, some characteristics of the spectrum, especially for the inertial subrange, were found as follows: (See Appendix B for the confidence interval of the spectrum)
1. The local Reynolds number, $Re_c$, increased with sensor elevation, which resulted in an expansion of the inertial subrange frequency band (see Fig. 2-5). This phenomenon strongly indicates that the local Reynolds number controls the development of the inertial subrange. When the Reynolds number is large enough, the inertial subrange can exist as a wave-number band where no energy is added by the mean flow and no energy is taken out by viscous dissipation (T&L, p.267). The results from T&L indicate that the inertial subrange in the spectrum of turbulence will increase with the Reynolds number. The results obtained here substantiate those of T&L. For sensors above level 6 ($z=150$ cm), the inertial subrange was absent. There exists a relatively broad inertial subrange at level 5 ($z=100$ cm) where the local $Re$ is around 4,000-15,000, which satisfies the critical Reynolds number of Huntley (1988b), $Re_c=3000$, and is close to the minimum condition of T&L, $Re>4,000$ (T&L, p.266).

2. Compared to $u'$ spectra (Fig. 2-5), $w'$ spectra do not present such a broad "inertial subrange". Some authors (e.g., Grant et al., 1984; Champagne et al., 1977; Williams and Paulson, 1977) have recommended the use of the vertical flow spectra rather than the streamwise flow spectra because the former are less contaminated by wave motion in the inertial subrange. The data presented here show the opposite effect. The streamwise spectra have a broader inertial subrange than do the vertical flow spectra (Fig. 2-6). However, $w'$ spectra at level 5 also present the best developed inertial subrange.

3. Variation of spectral density with tidal cycle indicates that time-dependence can affect the character of the inertial subrange. Fig. 2-7 shows the variation of
Fig. 2-5 (1) Vertical variation of $u'$ spectrum, $S_p$ ($\text{cm}^2/\text{s}^3$), on the wave number $k$ ($1/\text{cm}$), domain at station 3 at 13:00, May 15, 1992. The last two letters: $u_1$, $u_2$, $u_3$, and $u_4$, represent the velocity at $z=10$, 25, 45, and 70 cm above bottom.
Fig. 2-5 (2) Vertical variation of u' spectrum, $S_p (cm^3/s^3)$, on the wave number $k$ (1/cm), domain at station 3 at 13:00, May 15, 1992. The last two letters: u5, u6, and u7, represent the velocity at $z=100$, 150, and 200 cm above bottom.
Fig. 2-6 (1) Vertical variation of $w'$ spectrum, $S_p$ (cm$^2$/s$^3$), on the wave number $k$ (1/cm), domain at station 3 at 13:00, May 15, 1992. The last two letters: w1, w2, w3, and w4, represent the velocity at z=10, 25, 45, and 70 cm above bottom.
Fig. 2-6 (2) Vertical variation of $w'$ spectrum, $S_p$ (cm$^2$/s$^2$), on the wave number $k$ (1/cm), domain at station 3 at 13:00, May 15, 1992. The last two letters: w5, w6, and w7, represent the velocity at $z=100$, 150, and 200 cm above bottom.
Fig. 2-7 (1) Temporal variation of $u'$ spectrum, $S_p$ (cm$^2$/s$^3$), on the wave number $k$ (1/cm), domain at station 3. The last two letters, u4, represent the velocity at $z=100$ cm above bottom. The numbers 1512, 1513, 1514, and 1515 represent the sampling time, 12:00, 13:00, 14:00, and 15:00, May 15, 1992.
Fig. 2-7 (2) Temporal variation of $u'$ spectrum, $S_p$ (cm$^3$/s$^2$), on the wave number $k$ (1/cm), domain at station 3. The last two letter, u4, represent the velocity at $z=100$ cm above bottom. The number: 1516, 1517, 1518, and 1519 represent the sampling time, 16:00, 17:00, 18:00, and 19:00, May 15, 1992.
spectral density with the tidal cycle (refers to Fig. 2-2). The effect of a steady current on the inertial subrange can be divided into three situations. The data (xk1512, xk1514, xk1517, and xk1518) are in the relatively steady accelerating /decelerating period, so they have a broader "inertial subrange". The data (xk1513 and xk1519) are in the unsteady accelerating period (variable temporal gradient) and the inertial subrange is narrow. During the period of changing current direction (xk1515, xk1516), the inertial subrange was degraded.

By analyzing spatial and temporal variations of the energy spectra for turbulent velocity fluctuations, it can be seen that steady conditions and the Reynolds number of the flow are relevant to the existence of an inertial subrange.

2.4.3 Turbulent momentum flux (or Reynolds Stress)

The turbulent momentum flux can be obtained from at least two techniques. One is directly from the definition of \( \langle w' u' \rangle \), which is the streamwise component of the Reynolds stress. For this study, vertical velocity fluctuations, \( w' \), and streamwise velocity fluctuations, \( u' \), have been measured at field sites. Thus, \( \langle w' u' \rangle \) can be directly estimated. The term \( \langle w' u' \rangle \) can also be derived from the integral of the cospectra of \( u' \) and \( w' \). Cospectral densities, which indicate the contribution to the covariance in unit intervals of cyclic frequency \( f \) or wave number \( k \), can be defined as

\[
\langle u' w' \rangle = \int_0^\infty \phi_{uw}(f) \, df = \int_0^\infty \phi_{uw}(k) \, dk. \tag{2-8}
\]

In practice, this relation can be further expressed as (e.g., Weiler and Burling, 1967)
Thus the relation of $f \phi_{uw}(f)$ vs. log $f$ or $k \phi_{uw}(k)$ vs. log $k$ can be plotted. The covariance $<w'u'>$ represents the vertical turbulent flux of horizontal turbulent momentum or the horizontal flux of vertical momentum by assuming $\rho=1$ g cm$^{-3}$ (Hinze, 1959). By definition, $\tau=\rho<w'u'>$, turbulent momentum flux reflects the Reynolds stress. The temporal and vertical variations of turbulent momentum flux with a tidal cycle have been shown in Figs. 2-8 and 2-9.

Compared with the variance of $u'$ and $w'$, the momentum flux $<w'u'>$ clearly decreases with elevation (Fig. 2-10). Because of the sensor alignment problem at the different measuring levels, the momentum flux data $<w'u'>$ were not as well correlated as the variance of $u'$ and $w'$, but do support the existence of a constant stress layer in the lower 100-150 cm above bottom, which agrees well with the results of the spectrum distribution (Fig. 2-5). In other words, the location of the constant stress layer matches that of a broad "inertial subrange". This result should be very useful for further study of the constant stress layer. When the relative height $(z/h)$ was taken as the vertical coordinate, the constant stress layer was defined as the lower quarter of the profile (Fig. 2-11).

2.4.4 Turbulent intensity

The turbulent intensity is the standard deviation of the turbulent velocity fluctuations ($u'$, $v'$, and $w'$). The calculation of the turbulent intensity is the same as the above-mentioned momentum flux calculation (Eq. (2-8) or (2-9)). It is important
Fig. 2-8 Temporal variation of momentum flux, $w'u'$, at station 3 during 12:00
5/15 - 3:00 5/17, 1992 (n=40 hours; see Appendix C for detailed values).
Fig. 2-9 Vertical distribution of momentum flux, $<w'u'>$, measured at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (n=40 hours, see Appendix C for detailed values).
Fig. 2-10 Comparison of three co-variances, $<u'u'>$, $<w'u'>$, and $<w'w'>$, at station 3 during 12:00-19:00 5/15, 1992 (n=8).
Fig. 2-11 Vertical distribution of momentum flux, $<w'u'>$, with relative height at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (n=40; see Appendix C for detailed values).
to note that this method works only in the absence of a dominant wave velocity peak in the spectrum (George et al., 1994). Otherwise the orbital wave velocity would dominate the spectrum and would not represent turbulent velocity. In that case, turbulent intensity can be estimated from the dissipation rate (George et al., 1994). The data measured here does not show any obvious differences between the two methods.

In order to compare with the Reynolds stress, $<w'u'>$, the variances of the turbulent velocity fluctuations, $<u'u'>$ and $<w'w'>$, are represented rather than turbulent intensity in this study. Figs. 2-12, 2-13, and 2-14 show the vertical and temporal variations of turbulent intensities squared. Some common features are as follows:

1) The amplitude of turbulent intensities varies directly with the mean current velocity, namely, large and small mean current velocities are associated with large and small turbulent intensities respectively.

2) Turbulent intensity squared, $<u'u'>$, decreases with increasing elevation (see Fig. 2-12). At low mean current velocities, turbulent intensity squared is relatively uniform with height above the bed. There is a steep vertical slope of the distribution when mean current velocities are larger than 40 cm/s.

3) The temporal variation of turbulent intensity squared, $<u'u'>$, in contrast to the temporal variation of momentum flux, is in phase with the mean flow velocities (see Fig. 2-13 and Appendix D).
Fig. 2-12 Vertical distribution of turbulent intensity squared, $<u'u'>$, at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (n=40; see Appendix D for detailed values).
Fig. 2-13 Temporal variation of turbulent intensity squared, $<u' u'>$, at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (n=40; see Appendix D for detailed values).
Fig. 2-14 Vertical distribution of turbulent intensity squared, $\langle w'w' \rangle$, at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (n=40; see Appendix E for detailed values).
4) The profiles of vertical turbulent intensity squared, $<w'w'>$, remain relatively uniform with height above the bed (see Fig. 2-14). The temporal variation of $<w'w'>$ is also in phase with the mean flow velocities (Fig. 2-15).

2.4.5 Energy dissipation rate

When the kinetic energy of turbulence decays, it is converted into thermal energy as small eddies are strained by viscous stresses. This phenomenon is called the dissipation of the turbulent kinetic energy. Because of its importance, turbulence dissipation has been the focus of numerous recent studies (e.g., George, 1994; Green, 1992; Huntley, 1988a,b). In the "inertial dissipation" method, the energy dissipation rate is necessary for the calculation of the bottom shear stress. Basically, there are three methods to obtain the energy dissipation rate. In the following section, these three methods will be introduced and evaluated.

2.4.5.1 Method 1-- determination of dissipation rate by Kolmogorov law

With the assumptions of locally isotropic, horizontally homogenous, stationary surface-layer turbulence conditions, a kinetic energy balance can be expressed as:

$$P(=<w'\ u'>\ \frac{\partial U}{\partial z})+\varepsilon +I_m = 0$$  \hspace{1cm} (2-10)

(Wyngaard & Cote, 1971; Champagne et al., 1977), where $P$ and $\varepsilon$ represent the shear production and the dissipation rate of turbulent kinetic energy; and $I_m$ is the measured imbalance, a substantial gain term under unstable conditions (Champagne et al., 1977). If $I_m=0$, shear production is balanced by energy dissipation and Eq.(2-10) is reduced to two terms.
Fig. 2-15 Temporal variation of turbulent intensity squared, $<w'w'>$, at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (n=40; see Appendix E for detailed values).
The velocity profile in the bottom boundary layer obeys the "law of the wall", namely,

\[ \frac{\partial U}{\partial z} = \frac{u_*}{kz} \]  

(2-11)

where } \( u_* \) is related to the vertical flux of momentum as

\[ u_*^3 = -\langle w' u' \rangle. \]  

(2-12)

Equating Eqs. (2-10) when } \( I_m = 0 \), (2-11) and (2-12), the expression for dissipation rate, } \( \varepsilon \), can be obtained,

\[ \varepsilon = \frac{u_*^3}{kz}. \]  

(2-13)

Thus, the dissipation rate of turbulent kinetic energy and shear stress (} \( \tau_0 = \rho u_*^2 \)) are directly associated. If there exists an inertial subrange where the flux of energy from low to high wave numbers is equal to the dissipation rate, the three dimensional inertial subrange spectrum } \( E(k) \) will satisfy

\[ E(k) = \alpha \varepsilon^{2/3} k^{-5/3} \]  

(2-14)

(e.g., T&L p265) where } \( k \) is the radian wave number and } \( \alpha \) is the Kolmogorov constant. In practice, the result can be extended to the one-dimensional spectrum, namely

\[ \phi_i(k_i) = \alpha \varepsilon^{2/3} k_i^{-5/3}, \]  

(2-15)

where } \( i = 1, 2, \) or } \( 3 \) which means that any of the three orthogonal components of turbulent velocities can be used (Huntley, 1988a). As previously mentioned, the
streamwise flow spectra were found to be less contaminated by waves compared to vertical flow spectra. Therefore, the streamwise flow spectra were used for obtaining the energy dissipation rate in this study. For convenience, \( k \) will represent the 1-D wave number throughout this dissertation.

### 2.4.5.2 Method 2—determination of dissipation rate by the concept of turbulent characteristic scale

From the fundamental concepts of turbulent theory, the dissipation rate can be related to the velocity and length scales. Taylor (1935) established the relation

\[
e = \frac{u^3}{l},
\]

(2-16)

where \( u \) and \( l \) are the velocity and length scales of the large-scale turbulence. This relationship implies that viscous dissipation of energy can be estimated from the large-scale dynamics. Here, the velocity scale \( u \) can be represented by the turbulent intensity, \( \sqrt{u'^2} \) (George et al., 1994). The length scale, \( l \), scales with the largest eddies (T&L, p20). Svendsen (1987) defined it as \( 0.2h < l < 0.3h \), where \( h \) is water depth. In this study, \( l = 0.25h \) is adopted (George, 1994). Because the data were used only when the tidal level was above the top sensor (200 cm), the value of the length scale is within the range of 50 cm to 150 cm (maximum water depth = 700 cm).

### 2.4.5.3 Method 3—determination of dissipation rate by turbulence isotropy

The third method of calculating turbulent energy dissipation rate is with the second moments of the spectra using the isotropic relation:
(Champagne et al., 1976), where \( v \) is the kinematic viscosity of water, and \( \phi_{\text{m}}(k) \) is the spectral density in the streamwise direction. Once \( \phi_{\text{m}}(k) \) is obtained, the calculation of the above integral will be straightforward.

With regard to tidal cycle variations, a comparison of the three methods for estimating dissipation rate is shown in Fig. 2-16. Basically, they have similar vertical distributions, namely, dissipation rate decreases with elevation and varies directly with mean current velocities. But a large mean current velocity causes a large gradient of dissipation rate profile, and a low mean current velocity is associated with a relatively uniform vertical distribution of dissipation rate. The determination of friction velocity depends on the vertical distribution of dissipation rate in the "inertial dissipation" method. However, within the range of 50 cm to 150 cm, the profile of the dissipation rate is relatively uniform. So, this kind of distribution will not significantly affect the determination of shear stress. Comparing with the three methods derived from the different physical concepts, the results presented here indicate that 1) A good agreement among the three methods exists in the case when mean current velocity is large (\( u > 20 \) cm/s). 2) The calculation of method 1, "Kolmogorov law", depends on values from the inertial subrange of the spectrum. However, the spectrum in the inertial subrange is not an exact straight line. Thus, the calculation may produce an error. 3) For method 2 (turbulent scale), a problem exists as to how to set the turbulent length scale, \( l = \alpha h \), where \( \alpha \) is a constant parameter. In this study, the results indicate that when \( \alpha = 0.25 \),
Fig. 2-16 The results of three dissipation rate estimations compared with corresponding mean current velocity profiles at station 3 during 12:00-19:00 5/15, 1992, where $e_1$ is the result from the "Kolmogorov law" method; $e_2$ is the result from the "turbulent scale" method; and $e_3$ is one from the "integration" method.
the bottom shear stresses from method 2 have the best agreement with the results from
the "shear velocity" method. This relation will be discussed in more detail later. 4) Method 3 (the integration method using Eq. 2-17) is relatively easy to use for
calculation purposes. The statistical relations among these three methods (called e1, e2
and e3) have been shown in Fig. 2-17 and 2-18 separately.

2.4.6 Effects of Reynolds Number

From this study, it has been found that the Reynolds number controlled the
development of the "inertial subrange". Actually, the Reynolds number represents the
relative importance of inertial forces to viscous forces. Two forms of Reynolds number
in the bottom boundary layer can be: "bulk" Reynolds numbers, Re=U(δ)δ/ν, in which
the scales are those of the mean flow, δ is the thickness of bottom boundary layer, and
"turbulent" or "local" Reynolds numbers, Reₜ=ku₂/ν, in which the scales are those of
the turbulence (Bradshaw, 1978). Reₜ reflects the local mean shear in which a
logarithmic boundary layer velocity profile is proportional to 1/z. It also represents the
situation in which the length scales for mean production and advection are separated
from inertial-range length scales (T&L, Chapter 8).

A turbulent boundary layer is a necessary condition for the existence of an
inertial subrange where the shear production can be balanced by turbulent energy
dissipation (T&L, Chapter 8). In order for a turbulent boundary layer to exist, the local
Reynolds number has to be greater than the critical Reynolds number, Reₑ=3000,
namely, Reₜ=ku₂/ν>Reₑ (Huntley, 1988a). The kinematic viscosity, ν, will depend on
the water temperature. For data from the first sampling period (5/15 12:0-5/17 3:0),
Fig. 2-17 The statistical relation between the dissipation rate determined from the "Kolmogorov law" method, $e_1$, and that from the "turbulent scale" method, $e_2$, at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (n=40).
Fig. 2-18 The statistical relation between dissipation rate determined from the "Kolmogorov law" method, $e_1$, and that from the "integration" method, $e_3$, at station 3 during 5/15 12:00-5/17 3:00 ($n=40$).
average temperature was 17.4°C, with a range of 16°C to 21.3°C. The position of the critical Re, the critical height, z_c, has been shown in Fig. 2-19. The concept of the critical height, z_c, can help us to know where the "inertial subrange" exists. The reason why some critical heights were above 100 cm is that flow velocities were very small during the high water level period. Generally, however, the position of the critical Re is below 50 cm from the bottom. The mean vertical Re distribution has been calculated (see Fig. 2-20), and its mean value is listed in table 2-2.

<table>
<thead>
<tr>
<th>z(cm)</th>
<th>35</th>
<th>70</th>
<th>125</th>
<th>175</th>
<th>200</th>
</tr>
</thead>
<tbody>
<tr>
<td>Re_i</td>
<td>1806</td>
<td>3631</td>
<td>6450</td>
<td>9048</td>
<td>10225</td>
</tr>
</tbody>
</table>

The critical height z_c must satisfy the relation (Gross & Nowell, 1985)

\[ z_c = \frac{Re_z \sqrt{(\kappa u_\star)}}{v} \]  \hspace{1cm} (2-18)

When 3000 was taken as Re_z, the average critical height z_c was 56 cm. Gross & Nowell (1985) found that below 100 cm, dissipation seems to be underestimated by the "inertial dissipation" method relative to production. And this underestimate was associated with lower values of Reynolds number near the bed. For this study, the mean critical height is less than 100 cm. The critical height difference between this study and that presented by Gross & Nowell (1985) may be due to the strength of the
Fig. 2-19 The location of the critical height (of the critical Reynolds number) compared with the corresponding friction velocity, $u_*$, and tidal level.
Fig. 2-20 Vertical distribution of Reynolds number in the lower portion of the bottom boundary layer (z<200 cm) at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (n=40).
high-tide-range current and mean shear stress in this study. In addition, the data presented here indicates that buoyancy flux, a topic of chapter 4, would be another factor influencing the energy transport in this region.

2.4.7 Bottom shear stress

The determination of the bottom shear stress, $\tau_0$, is tremendously important both for practical and theoretical reasons. With relatively accurate in situ measured data available, a comparison of $\tau_0$ calculated by the three methods can be made. In order to present their relationships clearly, the three methods are indicated in table 2-3. For the "velocity shear" method (method 1), $\tau_0$ can be determined by $\tau_0 = \rho u^2$. The friction velocity $u_*$ would be obtained from the logarithmic velocity profile using levels of measured mean flow velocities.

For the "eddy correlation" technique (method 2), $\tau_0$ can be determined as $\tau_0 = -\rho <w'u'>$. As shown in Fig. 2-9, where $<w'u'>$ is the co-variance of velocity fluctuations, the values of $<w'u'>$ decrease with increasing elevation. For this method, the only area of interest is the constant stress layer where the flow shear stress is equal to the bottom shear stress. Thus, the height of the constant stress layer must be established before shear stress can be estimated.

For the "inertial dissipation" method (method 3), the determination of bottom shear stress will depend on the availability of dissipation rate information. Once the dissipation rate has been obtained, the bottom shear stress can be determined by Eq. (2-3). Because dissipation rate can be determined in different ways, the shear stress
Table 2-3 Definition of symbols of three methods for calculating bottom shear stress

<table>
<thead>
<tr>
<th>Method #</th>
<th>Method description</th>
<th>Sub-Method (Dissipation rate)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Method 1</td>
<td>Velocity shear (Eq.2-1)</td>
<td>$u_*$</td>
</tr>
<tr>
<td>Method 2</td>
<td>Eddy correlation (Eq.2-2)</td>
<td>$u_*$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1. $u_{r1}$: $e_1$ - Eq.(2-15)</td>
</tr>
<tr>
<td>Method 3</td>
<td>Inertial dissipation</td>
<td>$u_*$</td>
</tr>
<tr>
<td></td>
<td>(Eq.2-3: $u_{r3} = (\kappa z e_3)^{1/3}$)</td>
<td>2. $u_{r2}$: $e_2$ - Eq.(2-16)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3. $u_{r3}$: $e_3$ - Eq.(2-17)</td>
</tr>
</tbody>
</table>

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estimates calculated by method 3 may vary. The values, $u_{e_{1b}}$, $u_{e_{3b}}$ and $u_{e_{3c}}$, corresponding to the three dissipation estimations will be introduced separately (Table 2-3).

Data from a typical tide cycle (May 15 12:00-19:00, 1992) are used to compare the three methods. Because the "shear velocity" method has been widely used, the other two methods will be compared with it. By applying these data to the three methods, some interesting points can be found (refer to Fig. 2-21). The relations between these three methods are also listed in Tab. 2-4, where the means of friction velocity at levels 4, 5 and 6 for methods 2 and 3 were normalized with the values from method 1. Similarly, the normalized friction velocity profiles calculated with the three methods are also shown in Fig. (2-22).

These relations can be summarized as follows:

1. Reynolds stress obtained by the "eddy correlation" method shows a normal vertical shear stress distribution, where a constant stress layer exists below $z=100-150$ cm above bottom, and above that its value decreases with height. The absolute values from this method are smaller than the results estimated from the other methods.

2. For the "inertial dissipation" method, the results from the turbulent scale sub-method ($u_{e_{1b}}$) correspond most closely to those from the classical "shear velocity" method ($u_{e_{1}}$) (see Fig. 2-23); the two energy spectrum methods($u_{e_{3a}}$ & $u_{e_{3c}}$) also show good agreement with the "shear velocity" method during the high mean velocity period ($u > 20.7$ cm s$^{-1}$). But when tidal currents approach a sluggish period (current direction is changing), there are many differences between them (refer to k1515 and
Fig. 2-21 Comparison of the results from four friction velocity estimations at station 3 during 12:00-19:00 5/15, 1992, where \( u_z \) is the result from the "eddy correlation" method; \( u_{3a} \), \( u_{3b} \), and \( u_{3c} \) refer to the results which dissipation obtained from the "Kolmogorov law", "turbulent scale", and "integration" method separately.
## Table 2-4 Comparison of different methods for calculating friction velocity
(May 15 12:00-19:00, n=1-8)

<table>
<thead>
<tr>
<th>No.</th>
<th>Data</th>
<th>u (cm/s)</th>
<th>$u_{-2}/u_1$</th>
<th>$u_{-3a}/u_1$</th>
<th>$u_{-3b}/u_1$</th>
<th>$u_{-3c}/u_1$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>k1512</td>
<td>48.5</td>
<td>0.163</td>
<td>0.890</td>
<td>1.062</td>
<td>0.818</td>
</tr>
<tr>
<td>2</td>
<td>k1513</td>
<td>63.5</td>
<td>0.491</td>
<td>0.768</td>
<td>1.012</td>
<td>0.793</td>
</tr>
<tr>
<td>3</td>
<td>k1514</td>
<td>26.9</td>
<td>0.152</td>
<td>1.172</td>
<td>0.895</td>
<td>1.434</td>
</tr>
<tr>
<td>4</td>
<td>k1515</td>
<td>12.8</td>
<td>0.114</td>
<td>3.113</td>
<td>2.394</td>
<td>4.782</td>
</tr>
<tr>
<td>5</td>
<td>k1516</td>
<td>7.6</td>
<td>0.015</td>
<td>4.861</td>
<td>0.906</td>
<td>3.282</td>
</tr>
<tr>
<td>6</td>
<td>k1517</td>
<td>20.7</td>
<td>0.286</td>
<td>2.242</td>
<td>1.293</td>
<td>2.379</td>
</tr>
<tr>
<td>7</td>
<td>k1518</td>
<td>30.8</td>
<td>0.286</td>
<td>1.348</td>
<td>1.165</td>
<td>1.518</td>
</tr>
<tr>
<td>8</td>
<td>k1519</td>
<td>56.6</td>
<td>0.181</td>
<td>0.789</td>
<td>0.754</td>
<td>0.668</td>
</tr>
</tbody>
</table>
Fig. 2-22 The relative value of four friction velocity estimations compared with corresponding classical "shear velocity" method at station 3 during 12:00-19:00 5/15, where $u_1$ is the friction velocity from the "shear velocity" method; $u_2$ is the result from the "eddy correlation" method; $u_{3a}$, $u_{3b}$, and $u_{3c}$ refer to the results which dissipation obtained from the "Kolmogorov law", "turbulent scale", and "integration" method separately.
Fig. 2-23 The statistical relation of friction velocity estimations between the "shear velocity" method, $u_*$, with the "turbulent scale" method, $u_{3b}$, at station 3 during 12:00 5/15 - 3:00 5/17, 1992.
There are reasonable explanations for those differences. First, the low mean velocity implies low Reynolds numbers. The spectra during that period have shown a deterioration of the inertial subrange. Second, the velocity profile was not fully logarithmic and correlation coefficients were low. Third, the buoyancy term may not be neglected during the periods of lower mean current velocity. The measured data showed that the suspended sediment concentration gradient did not decrease too much although the concentration value approached its lowest value (only half of its highest value) during the low velocity period (more details in Chapter 4).

As we know, the buoyancy term depends mainly on the density gradient rather than density itself. Considering the low shear production during this time period, the buoyancy term should be taken into account in the turbulent energy budget.

3. For the "inertial dissipation" method, the two energy spectrum methods (\(u_{13}\) & \(u_{23}\)) remained relatively steady with the tidal variation; while the results from the turbulent scale method (\(u_{\tau_b}\)) were strongly related to tidal cycle because this method is closely associated with the water depth.

4. Vertical distribution of shear stress indicates that near \(z=100\) cm, the bottom shear stresses from the "inertial dissipation" method are in good agreement with those from the "shear velocity" method. For the "inertial dissipation" method, only one level of velocity data is required to determine the bottom shear stress. Therefore \(z=100\) cm appears to be an optimal placement of the sensors. The results from the three different inertial dissipation methods indicate that shear stress increases with elevation in the lower 150 cm. An increase with height is inconsistent with a constant stress
theory and the results of Reynolds stress profiles. Two reasons are possible: 1) The Reynolds number near bottom is less than the critical Reynolds number. Gross & Nowell (1985) found that spectral estimations of \( u_* \) from dissipation using wall scaling agree quite well with mean profile estimates of \( u_* \). Close to the wall, however, the agreement broke down because of inadequate separation between energy production and viscous dissipation scales. The results presented here also support this explanation. 2) The existence of buoyancy forces induced by temperature and suspended sediment concentration gradient could influence the dissipation and the magnitude of the friction velocity. When the buoyancy term can not be neglected, use of the "inertial dissipation" methods may cause errors in the calculation of \( u_* \) (see Chapter 4).

\section{Discussion}

\subsection{The constant stress layer}

The "inertial dissipation" method is based on the assumptions that there exists an inertial subrange where the local \( \text{Re} > \text{Re}_c \) and a constant stress layer (Huntley, 1988a). Turbulent measurements for using the "inertial dissipation" method must be made in the constant stress layer. The data presented here indicate that it is relatively easy to satisfy \( \text{Re}_p > \text{Re}_c \) when measurements are above the critical height \( z_c = 0.56 \text{ m} \). The second condition, however, is not as straightforward. According to Huntley (1988a), the thickness of the constant stress layer is one-half the thickness of the logarithmic layer \( \delta_{\log} \), namely,
where $\delta_{log}$ is considered to be 10% of the thickness of the bottom boundary layer, $K$ is the thickness coefficient (see the detail in Chapter 4), and $f$ is the Coriolis parameter. At 37° N latitude, $f=8.77 \times 10^{-5}$ s$^{-1}$. When the mean friction velocity $u_* = 1.47$ cm/s is substituted into the above relation, $\delta_{a}=3.35$ m. This condition seems very easy to satisfy, too. But the vertical distribution of Reynolds stress from the data measured here indicates that the thickness of the real constant stress layer is much less than this value ($\delta_{a}=3.35$ m). For the situation in the inter-tidal mudflat channel, Eq. (2-19) seems to overestimate this thickness. This overestimate is probably due to approximations in the determination of boundary layer thickness.

2.5.2 Comments on the relation of friction velocity to dissipation rate

The cornerstone of the "inertial dissipation" method is Eq. (2-13) which is formed by the introduction of two basic concepts. One is the approximation of the kinetic energy balance equation. If a neutral, locally isotropic, horizontally homogenous, stationary bottom turbulence can be assumed, the kinetic energy balance can be expressed as Eq. (2-10) (when $I_w=0$), namely, the dissipation of turbulent kinetic energy by viscosity is equal to the energy production by velocity shear. The other is the well-known "law of the wall", namely Eq. (2-11). Both of which are based on the condition of neutral stratification. In the tidal channel, the vertical temperature, salinity and sediment concentration gradients could contribute to the
density stratification. Clearly, neutral stratification can not always be assumed and the buoyancy flux term can not always be neglected in the kinetic energy budget equation.

§ 2.6 Summary

Based on the foregoing analysis, this chapter can be summarized as follows:

1. Reynolds stress profiles constructed from "eddy correlation" calculations typically exhibited a constant stress layer in the region, \( z \leq 100 - 150 \) cm. Reynolds stress calculated by the "eddy correlation" technique was lower than the results estimated from the other two methods.

2. The "inertial dissipation" method for obtaining bottom shear stress is useful when the measured data is within the constant stress layer. For this study, the practical thickness of the constant stress layer is less than that given by Equation (2-19). For the non-neutral stratification situation this method fails to satisfy the basic assumption of the kinematic energy balance equation.

3. When the dissipation rate was calculated from the basic turbulence scale concept, \( \varepsilon = u^3/l \), the calculated bottom shear stress showed good agreement with that of the classical "shear velocity" method. Using this scale concept, the "inertial dissipation" method could be simplified. The turbulent length scale, \( l \), was set as a function of water depth, \( l = 0.25 \) h.

4. The local Reynolds number strongly controls the development of the inertial subrange. At the 100 cm level there existed the broadest inertial subrange. \( Re \) was within the range 4,000-15,000. When the critical Reynolds number, \( Re_c \), was set as 3000, the average critical height, \( z_c \), was 56 cm. The results of calculation indicate

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that for \( z < 56 \text{ cm} \), the "inertial dissipation" method will underestimate bottom shear stress.

5. In this study, shear stress calculated by the "inertial dissipation" method using the data at \( z = 100 \text{ cm} \) showed the best agreement with the "shear velocity" method. Additional studies are needed to verify this relation.

6. The three different methods for calculating dissipation rate yield similar vertical distributions. Energy dissipation rates varied with mean current velocities. However large mean velocity resulted in larger vertical gradients of dissipation rate. Low mean velocity usually resulted in a relatively uniform vertical distribution of dissipation rate.

7. Contrasted to the vertical momentum flux profiles, the turbulent intensity profile was relatively uniform. When the mean current velocity was large enough, the gradient of the profile increased.

8. Turbulent velocity spectra demonstrate that near high tide when currents were weak, the "inertial subrange" in the spectra could be severely diminished because of the unsteady flow and low Re.

9. Mean turbulent velocity from the shear velocity method showed that friction velocity data correlated well with tidal height. For 184 data points, \( u_{\text{mean}} = 1.47 \text{ cm s}^{-1} \).
Chapter 3

Sediment transport and mean current characteristics in the intertidal mudflat channel
§3.1 Introduction

In order to understand the effect of sediment resuspension on the bottom shear stress, a knowledge of sediment transport features and mudflat processes is important. In chapter 2, we mainly discussed the boundary layer turbulence and the bottom shear stress. Now the focus will be on the sediment movement on the mudflat. Intertidal mudflats are a consequence of interactions between currents and sediments in the nearshore zone. Owing to unsteady characteristics of tides and waves, and the addition of non-neutral stratification caused by vertical gradients of sediment concentration, salinity and temperature, the problem is complex. Many of the related problems such as intertidal mudflat processes, sediment exchange in shallow water, and wave induced mud transport are still far from being fully understood (Dyer, 1989).

With a high-tidal environment, the west coast of the Korean peninsula exhibits a broad intertidal mudflat incised by numerous tributary channels (Fig. 3.1). The in situ measured data indicated that the mudflats exhibit high deposition rates (0.15-2.0 cm/yr) (Wells et al., 1990). Park & Khim (1992) believed that Southeastern Yellow Sea mud is largely associated with the discharges of the Keum and Yongsan Rivers, a few hundred kilometers south of Namyang Bay. The source of sediment necessary to account for the observed high deposition rates on the Namyang Bay mudflats is an interesting and important problem. Wells and Park (1992) proposed a model (Hereafter referred to as the W&P model) of the Keum River Estuary, which indicated a low sediment influx from the river but high influx from the tidal flats to the coastal
Fig. 3-1 With a high-tide-range environment, the study area presents a broad intertidal mudflat incised by numerous distributary channel.
areas during winter, at a time when north winds generate large waves in the Yellow Sea, and the reverse in summer, when winds from the south produce little wave activity. Coastal mud would thus be eroded in winter and replenished in summer. The W&P model outlined the general sediment source variation with season in this area. In order to explain mudflat sediment transport processes, more detailed analysis of the interaction between the current and bed sediment is necessary.

In this chapter, emphasis will be on data from station 2. The current velocity data were taken from three levels (z=20, 45 and 100 cm above bottom). Optical Backscatter Sensor (OBS) measurements were made at two levels (z=45 and 100 cm). Sampling was conducted from 0:00, May 11 to 15:00, June 9, 1992 for 20 minutes every hour. During this period, flow kinematics and associated suspended sediment distributions were acquired over a complete spring-neap-spring tidal cycle, which is composed of a period of maximum tidal current flow as well as periods of flow acceleration and deceleration on each side of the maximum. The current velocity, sediment concentration and tide level data during the sampling period are presented in Figures 3-2 and 3-3.

The Korean peninsula is strongly affected by monsoonal winds. The winter monsoon season, from October to April, is dominated by northerly winds which drive a flow to the south (Wells & Park, 1992; Watts, 1969). The summer monsoon season from June to September, is characterized by light and variable south winds (Huh, 1982). The data to be analyzed are within the latter period; winds near the end of the
Fig. 3-2 Temporal variation of mean current velocities at 3 levels, $u_1$, $u_2$, and $u_3$ ($z=20$, 45, and 100 cm above bottom separately) at station 2 during 0:00 5/11 -15:00 6/9, 1992 (corresponding to total 712 hours).
Fig. 3-3 Temporal variation of sediment concentrations at 2 levels, C1 and C2 at z = 45 and 100 cm above bottom and tidal level elevations at station 2 during 0:00 5/11 - 15:00 6/9, 1992 (n=712 hours).
measuring period obviously increased. Detailed wind data are not available at present, thus a detailed study of wind effects on sediment is not possible.

Visual observations show the presence of broad distributary channels in the study area. Mudflat channels play a very important role in the transport of sediment. Knowledge of the function of a mudflat channel for sediment transport is obviously relevant. Based on the data from a mudflat channel, this chapter has two objectives:

1. To understand basic characteristics of channel sediment movement in the context of high tidal range variation, the following aspects are addressed:
   a) Horizontal bed sediment distribution in the mudflat channel
   b) Relationships of vertical sediment concentration profile to bottom shear stress, $\tau_0$, and sediment critical stress, $\tau_c$, and the resuspension coefficient, $\gamma_0$
   c) Relationship of current velocity profile to suspended sediment concentration profile in which Richardson number, $R_i$, is a relevant factor
   d) Mean flow velocity profile and its variation through the tidal cycle where the key factors are friction velocity, $u_*$, and roughness length, $z_o$
   e) Relationships of bottom shear stress to mean current velocity, in which the drag coefficient, $c_d$, is a relevant indicator
   f) Effect of residual current on sediment transport

2. To describe sediment transport on the mudflat including:
   a) Sediment transport rate in a mudflat distributary channel
   b) Sediment transport process analysis on the mudflat surface
These results are essential for understanding the effects of sediment resuspension on current stability, a problem that will be examined in Chapter 4.

§3.2 Sediment deposition characteristics on mudflats

3.2.1 Horizontal distribution of sediment

To delineate horizontal distributions of sediment, sixteen surface sediment sampling sites were occupied along the studied distributary channel. Nine samples are from the mudflat surface and seven are from the channel bed. Comparison of samples showed that the channel bed sediments were much coarser than those on the mudflat surface. Additionally, grain size decreased in a downstream direction (see Fig. 3-4). The distribution characteristics can be described as follows:

When the tidal level reaches its peak, the fine grain sediment will fall on the bed owing to the presence of very low flow velocity. The settling fine material will be consolidated on the surface bed and form a cohesive sediment dominated surface bed with high dry density. This kind of bed is very difficult to erode. With falling tide levels, upstream erosion occurs in distributary channels which are prominent on the mudflat. The resultant upstream channel erosion is enhanced because of channel meandering. Hydraulic jumps at step changes in topography are obvious evidence of this erosion (Adams et al., 1990). The slopes of channel beds are much steeper than those of the mudflat surface. Thus the flow in the channel will carry the eroded fine material downstream. Because of the relatively higher velocity in channels, channel
Fig. 3-4 Sixteen surface sediment sampling sites were established along a channel tributary with 9 samples from the mudflat surface and 7 samples from the channel bed. Analysis of the samples showed that the sediment grain size (values (mm) are shown in parentheses) was significantly larger in the channel bed than on the mudflat surface, and generally decreased in a downstream direction on the mudflat surface.
beds will retain the coarser material only. This result fundamentally reflects a typical characteristic of cohesive sediment: it is readily transported because of its fine grain size, but due to the presence of inter-particle electric bond, it is hard to erode (Partheniades, 1993). On the other hand, the current velocity will decrease in the seaward direction because of the presence of a weaker bed slope. Therefore sediment grain size becomes finer downstream.

3.2.2 Vertical distribution of suspended sediment and reference concentration

The determination of a reference concentration, \( C_0 \), is a key to predicting the vertical sediment concentration profile. By means of a reverse calculation of \textit{in situ} measured concentration data, estimated reference concentration results indicate that the reference concentration basically varies with the bottom shear stress. These results also are consistent with the relationship,

\[
C_0 = \frac{\gamma_0 C_b S}{1 + \gamma_0 S}, \quad (3-1)
\]

presented by Smith and Mclean (1977b), where \( S = (\tau_0 - \tau_c)/\tau_c \); \( \tau_c \) is the critical shear stress of the bed sediment; \( C_0 \) is the concentration of sediment in the bed (=0.6); \( \gamma_0 \) is the resuspension coefficient. Because \( 1 + \gamma_0 S = 1 \) (\( \gamma_0 < 10^{-3} \)), Eq. (3-1) can be approximately expressed as:

\[
C_0 = \gamma_1 \frac{\tau_0 - \tau_c}{\tau_c}, \quad (3-2)
\]

where \( \gamma_1 = C_b \gamma_0 \). For \( \gamma_0 \), different studies have yielded different values. Dyer (1980)
found \( \gamma_0 = 7.8 \times 10^{-5} \) for a steady tidal flow. The result obtained by Smith & Mclean is \( 2.4 \times 10^{-3} \); Glenn & Grant (1983) found this value to be around \( 2.0 \times 10^{-3} \). Recent observations by Bedford (1994) support the results found by Glenn & Grant (1983). For this study, the resuspension coefficient is found to be \( 3.65 \times 10^{-5} \). The relationship of reference concentration \( C_0 \) and relative shear stress \( S = (\tau_0 - \tau_c) / \tau_c \) is shown in Fig. 3-5. The measured data indicate that reference concentration is closely related to bottom shear stress in the first half of the measurement period (first 14.6 days) when winds were weak or absent (see Fig. 3-6). Upon arrival of summer monsoon season (early June), the correlation between reference concentration and bottom shear stress deteriorated as sediment resuspended from the mudflat surface flowed into the distributary channel. Compared with previous studies, values of the resuspension coefficient obtained from this study is smaller. A possible explanation is the existence of a large sediment size difference between the mudflat surface and the channel bed. As mentioned before, the cohesive sediment on the mudflat surface is highly erosion-resistant. During the weak wind period, no major sediment on the mudflat surface could flow into the distributary channel. Thus, the suspended sediment concentration in the channel should be low and directly related to the bottom shear stress. So the reference concentration will be associated well with relative shear stress. However during the strong wind period, the sediment on the mudflat surface will be largely eroded by wind waves into distributary channel. That portion of suspended sediment is less related to the channel bottom shear stress, which was indicated with bad
Fig. 3-5 The relation of reference concentration $C_0$ and relative shear stress, $(\tau - \tau_c)/\tau_c$, at station 2 during 0:00 5/11 - 12:00 5/23, 1992.
Fig. 3-6 The relation of reference concentration $C_0$ with friction velocity, $u_*$. The upper figure shows the data measured at station 2 during the first time period (0:00 5/11 - 12:00 5/23) when it was in the fair weather situation. The lower figure shows data measured during the second time period (2:00 5/28 - 5:00 6/9) when it has reached the beginning of the summer monsoon season.
correlation between the reference concentration and relative shear stress. This is an unique feature in this study.

3.2.3 Critical shear stress of bed sediment

The critical shear stress, $\tau_c$, can be determined by the Shields Curve (e.g., Raudkivi, 1990), based on the relation:

$$\frac{\tau_c}{\gamma (\rho_s-\rho) d} = f \left( \frac{u_* d}{v} \right), \quad (3-3)$$

where $\rho_s$ and $\rho$ are the sediment grain and water density respectively; $d$ is the mean sediment diameter, taken as 0.12 mm for channel bed sediment in this study; $v$ is the kinematic viscosity ($=1.08 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ at $T_{\text{mean}}=17.4^\circ \text{C}$). With these known values, $\tau_c = 1.5 \text{ dyne cm}^{-2}$. This result is based on two assumptions. First, the mean sediment size sampled from the channel bed was taken as representative of sediment size because we are concerned mainly with the erosion of bed sediment. If bed sediment can be eroded, the fine-grained suspended sediment could not be deposited. Second, for the later calculation of vertical sediment concentration distribution using the Rouse formula, the mean sediment size on the mudflat surface was taken as the representative sediment size because the largest percentage of suspended sediment in the channel originated on the mudflat surface.

3.2.4 Richardson Number

The Richardson Number is a good indicator of flow stability. There are two kinds of Richardson numbers: Flux Richardson number and Gradient Richardson
Number. By definition, the ratio of buoyancy production to shear production is called the flux Richardson number, \( R_f \), namely,

\[
R_f = -\frac{\rho}{\rho_0} \frac{\langle \rho'w' \rangle}{\langle w'u' \rangle} \frac{\partial u}{\partial z}
\]

In the case where the sediment concentration gradient is the main contribution to the density gradient, the density of the sediment/water mixture can be expressed as:

\[
\rho = \rho_0 + (\rho_s - \rho_0)c,
\]

where \( c \) is the volume concentration of sediment with units of volume per volume. Another form of this important parameter is the gradient Richardson Number, defined as

\[
R_i = -g \frac{\rho_s - \rho_0}{\rho_0} \left( \frac{\partial c}{\partial z} / \left( \frac{\partial u}{\partial z} \right)^2 \right). \tag{3-5}
\]

Applying the multi-level sediment and current velocity data presented in this study to Eq. (3-5), the Richardson number can be estimated. From the temporal variation of \( R_i \) (see Fig. 3-7), it is found that during neap tides (low bottom shear stress) flow is unstable or neutral, while during spring tides (high bottom shear stress), flow is stably stratified. This variation can be explained as follows:

During neap tides, only very fine material is resuspended and the vertical sediment concentration distribution is relatively uniform. Compared with upstream bottom erosion, during this low tidal level period, the dominant erosion process would be channel flank slumping. This would result in a high sediment concentration in the.
Fig. 3-7 Temporal variation of gradient Richardson number, $R_i$, at station 2 during 0:00 5/11 - 15:00 6/9, 1992 (total 712 hours). The unstable stratification (negative value) in the middle of time series data matched neap tide period.
upper layers rather than at the channel bottom because the sediment is introduced near the water surface. With the low settling velocity of fine grain sediment, the temporary formation of unstable stratification would be possible.

§3.3 Hydrodynamic characteristics of mean channel current

3.3.1 Friction velocity $u_*$ and roughness length $z_0$

With the high tide level variation, the vertical velocity profile will change its shape as friction velocity and roughness length vary over the tidal cycle. In chapter 2, three methods to calculate the bottom shear stress were introduced. Here, the "shear velocity" method will be applied based on flow velocity data sampled from station 2. For all in situ measured data except those measurements taken when a sensor was out of the water, the temporal variation of friction velocity and roughness length are shown in Fig. 3-8, and the statistical results are summarized in Table 3-1. From this calculation, it is found that maximum value of friction velocity varies with tidal range. Roughness length, however, remains approximately constant. Fig. 3-9 suggests that the variance of roughness length varies inversely with shear stress. This relationship indicates that suspended sediment from the mudflat surface is less readily exchanged with bed sediment under conditions of high bottom shear stress. Thus bed sediment could maintain a relatively uniform size composition and bed form which are key factors to producing a constant roughness length.

Roughness lengths obtained in this study are similar to those presented by Gross and Nowell (1983, Fig. 5). Smith and Mclean (1977a) found that roughness...
<table>
<thead>
<tr>
<th>data</th>
<th>Data No.</th>
<th>% of total</th>
<th>$z_0$ (cm)</th>
<th>$u_*$ (cm/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>all available data</td>
<td>483</td>
<td>67.8</td>
<td>1.286</td>
<td>3.48</td>
</tr>
<tr>
<td>data ($r^2&gt;0.95$)</td>
<td>331</td>
<td>46.4</td>
<td>0.483</td>
<td>2.73</td>
</tr>
<tr>
<td>data ($r^2&gt;0.996$)</td>
<td>85</td>
<td>11.9</td>
<td>0.178</td>
<td>2.39</td>
</tr>
</tbody>
</table>
Fig. 3-8 Temporal variation of friction velocity, $u_*$ (upper figure) and roughness length, $z_0$ (lower figure) at station 2 during 0:00 5/11 - 15:00 6/9 (total 712 hours).
Fig. 3-9 The relationship of friction velocities, $u_*$, and roughness length, $z_0$, with high correlation coefficients ($r^2 > 0.996$) at station 2 (upper figure). The lower figure indicates the variance variation of $z_0$ with $u_*$. 

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lengths are proportional to bottom shear stresses. This relation can be described as follows:

\[ z_0 = \alpha_0 \frac{\tau_0 - \tau_c}{(\rho_s - \rho) g} + z_u, \quad (3-6) \]

where \( \alpha_0 \) is a constant and \( z_0 \) is the value obtained from the experiments of Nikuradse (1933). Green (1990) found roughness length to be a function of the thickness of the viscous sublayer in wave-current interacting tidal boundary layers, namely,

\[ z_0 = \frac{v}{g u_*}. \quad (3-7) \]

Eq. (3-7) expresses an inverse relationship to bottom shear stress. In the case of low-tide-range, with relatively low current velocity, it has been found that roughness length is inversely related to the bottom shear stress owing to variation of sediment composition and bottom bio-disturbance (Adams et al., 1996). That may suggest roughness length is not only a function of bottom shear stress, as the above two relations indicate, but also a function of characteristics of sediment movement, i.e., the variation of bed form morphology and sediment composition. In this case, although sand ripples are present on the channel bed, the suspended fine material did not participate in the flow-bed exchange process. Therefore the moving bed form would keep its shape. Based on this understanding, it may be concluded that variation of the bed material composition is one of the main factors controlling the variation of roughness length. Thus, it is implied that the variation of sediment source and composition must be considered before applying the roughness length formulas.
3.3.2 Drag coefficient $C_d$

Through regression analysis of logarithmic velocity profiles, the shear velocity and $u_{100}$ can be calculated. Finally, $C_d = (u_0/u_{100})^3$ will be determined. The results (only for $r^2 > 0.996$) indicate that as $u_{100}$ increases, variance of $C_d$ decreases (see Fig. 3-10). Basically, the drag coefficient is approximately constant with $C_d = 4.215 \times 10^{-3}$ (refer to upper figure of Fig. 3-10). Here from another standpoint, it is indicated that the relatively constant drag coefficient is consistent with an unvarying roughness length. Actually drag coefficient is an indicator of bed friction.

3.3.3 Variation of velocity profile with tidal cycle

The tidal level variation will strongly affect the vertical velocity profile. Thus logarithmic velocity profiles did not always exist. The correlation between the measured data and a logarithmic profile shows some interesting relationships with the tidal level:

1). Data are poorly correlated at (see Fig. 3-11):
   a. low tidal level (some of the sensors were out of the water)
   b. peak tidal level (current is changing direction)
   c. periods of high current velocity but with significant acceleration/ deceleration

2). Data are well correlated at (see Fig. 3-12):
   peak velocity, especially during the flood velocity peak (low temporal velocity gradient). The general trend is that data have high correlation coefficients during maximum flood velocities and deteriorate when the velocity decreases.
Fig. 3-10 Temporal variation of drag coefficient $C_d$ at station 2 during 0:00 5/11 - 15:00 6/9 (upper figure). The lower figure indicates the relation of drag coefficient with the current velocity at $z=100$ cm.
Fig. 3-11 Temporal variation of correlation coefficients for logarithmic velocity profiles and corresponding tidal level variations at station 2 during 4:00 5/15 - 0:00 5/21 (corresponding to the data number from 100 to 240 in total data set), which mainly indicates where low correlations occurred.
Fig. 3-12 Temporal variation of correlation coefficient for logarithmic velocity profiles and corresponding mean velocity variations at station 2 during 12:00 5/23 - 8:00 5/29 (corresponding to data numbers from 300 to 440 in total data set), which mainly indicate where high correlations occurred.
By analogy to a method derived by Monin & Obukhov (1954) for non-neutral stratification flow, Soulsby & Dyer (1981) introduced a dimensionless parameter:

\[ \zeta = \frac{z}{\Lambda} = \frac{\partial u_*/\partial t}{u_*|u_*|} \]

(3-8)

where \( \Lambda \) is an acceleration length scale. Velocity profiles can be determined by

\[ u = \frac{u_*}{K} \left( \ln \frac{z}{z_0} - \frac{z-z_0}{\gamma \Lambda} \right) \]

(3-9)

where \( \gamma \) is a linear coefficient. For steady flow, \( \Lambda \to \infty \), and the above equation becomes the logarithmic velocity formula. For an accelerating current, \( u_* \) will decrease; for a decelerating current \( u_* \) will increase (Soulsby & Dyer, 1981). During periods of maximum velocity, \( u_* \) variation approaches zero, so flow is approximately steady and the logarithmic velocity profile fits the measured velocity profile well.

3.3.4 Residual current

Through application of several low pass filters (with cutoff periods of 14.8, 3.0 days, and 23.7, 12.9, 11.8, 1.0 hours), it was found that the lower the cutoff frequency, the more dominant the net seaward residual current in the channel. After application of a low-pass filter with a 14.8 day (\( > \) the period of lunar fortnightly) cutoff period, the residual current can be divided into three periods (see Fig. 3-13), which are:
Fig. 3-13 (1) Residual current after low-pass filtering (t>1.0, 11.8, 12.9 hours) at station 2 during 0:00 5/11 - 15:00 6/9 (total 712 hours).
Fig. 3-13 (2) Residual current after low-pass filtering (t> 23.7, 72.0, 356 hours) at station 2 during 0:00 5/11 - 15:00 6/9 (total 712 hours).
i. first 8 day period which is dominated by seaward residual flow

ii. next 12.5 day period which exhibits a landward current

iii. final 9 day period when current returns to seaward direction

This residual pattern corresponds well to the sediment transport rate in the channel (compare to Fig. 3-15), which suggests that the direction of the residual current is a dominant factor in determining the net sediment transport on the mudflat. When upstream channel erosion caused by the ebb current matches the net seaward residual current, net seaward sediment transport from the mudflat will increase. It is difficult to explain, however, why sediment concentration increased during the last two day period when the bottom shear stress increased very little. It is our present understanding that the last sampling period (Early June) marked the beginning of the summer monsoon season when wind speed would increase dramatically. Thus wave-induced bottom shear stress has to be considered in the estimation of total bottom shear stress.

§3.4 Calculation of sediment transport rate in the channel

To better understand inter-tidal mudflat sediment transport processes, the variation of sediment transport rate in the major channel must be known. Calculations, based on the data from station 2, were made for this purpose. The calculation of sediment transport rate can be divided into the following steps.
3.4.1 Calculations to determine u. and \( z_0 \)

To determine \( u_0 \) and \( z_0 \), current data from three elevations have been fitted to a logarithmic velocity formula which can be expressed as:

\[
\ln z = \frac{k}{u_0} u + \ln z_0.
\]  

(3-10)

It also can be written as the linear expression:

\[
y = \alpha x + \beta
\]  

(3-11)

where \( y = \ln z \), \( x = u \), \( u = k/\alpha \) and \( z_0 = e^\beta \). Based on the \textit{in situ} measured flow velocity profiles, the coefficients \( \alpha \), \( \beta \) can be obtained from a standard linear regression analysis (e.g., Brownlee, 1960). In practice, the regression coefficients \( r^2 \) of the data to a logarithmic velocity profile are not always satisfactory. If the \( r^2 \) has a low value, say \( r^2 < 0.95 \), the logarithmic velocity formula does not work well (Gross & Nowell, 1983). When this situation occurred, the mean current velocity could be adopted. As most of the low \( r^2 \) data were associated with low current velocities, this substitution has little effect on the computation of the sediment transport rate.

3.4.2 Calculation of vertical concentration profile using the Rouse formula (1935)

The Rouse formula derived from the diffusion equation can be expressed as:

\[
\frac{C}{C_0} = \left( \frac{H - z}{z_0} \right)^{\frac{\omega}{K_x u_0}} \]  

(3-12)

where \( C \) is the concentration at elevation, \( z \); \( C_0 \) is the reference concentration, a concentration at the reference height \( z_0 \); \( H \) is the water depth; \( \omega \) is the settling velocity of sediment grains; \( K_x \) is the eddy diffusivity and \( \beta_i \) is a coefficient. It was
found above that \( u_* \) varied during the tidal cycle, whereas \( z_0 \) was relatively constant.

The reference height, thus, can be treated as fixed and assumed to be equal to a roughness length. Once \( z_0 \) is determined, \( C_0 \) and \( P \), can be obtained by means of two-levels of \textit{in situ} measured concentration data, using the following relationships:

\[
P = \log\left(\frac{H - z_2}{z_2} \cdot \frac{z_1}{H - z_1}\right) / \log\left(\frac{C_2}{C_1}\right),
\]

\[
C_0 = C_1 \left(\frac{H - z_1}{z_1} \cdot \frac{z_0}{H - z_0}\right)^{-P}.
\]

For abnormal concentration profiles, i.e., positive concentration gradients, the mean sediment concentrations were taken as the computational concentrations at all levels because their profiles are rather uniform compared to the normal sediment concentration profile.

\subsection*{3.4.3 Calculation of sediment transport rate}

Sediment transport rate, \( q_s \), can be defined as follows:

\[
q_s(t) = \int \int u \delta \sigma, \quad (3-15)
\]

where \( \delta \sigma \) is the vertical elemental area across which sediment transport occurred. For the mudflat channel, the problem can be treated as one dimensional, and sediment transport rate per unit width, thus, becomes

\[
q_s(t) = \int_{z_i}^{z_f} u \delta z = \sum_{i=1}^{n} u(z_i) c(z_i) \delta z_i. \quad (3-16)
\]
In the meanwhile, the cumulative sediment transport rate which is most concerned can be expressed as:

\[ Q_s = \int_{t_0}^{t_f} q_s(t) \, dt = \sum_{j=1}^{n} \sum_{i=1}^{n} u(z_i) \, c(z_i) \, dz_i \, dt. \] (3-17)

Equation (3-17) in finite-difference form was used to examine variations of sediment transport rate with the tide cycle. The following comments pertain to the difference technique used.

a. The vertical grid spacing, \( \Delta z \), was adopted as 10 centimeters, and time interval \( \Delta t \) was 20 minutes.

b. Data were discarded for periods when all instruments were exposed to the air. Although there exists sediment transport in the lower part of channel during those periods, it is quantitatively negligible because of the low current velocities.

c. Sediment transport rate is positive when seaward (northward) and negative when landward (southward).

3.4.4 Factors affecting sediment transport rate

Sediment transport rate depends on three factors: 1) current velocity, 2) sediment concentration, and 3) water depth. Sometimes, high current velocity and concentration may produce a low sediment transport rate because of low tides and water depths. Under the action of wind stress and high air pressure, water level variation may not always vary regularly, which can create large differences between the amounts of sediment transported during ebb and flood periods. A typical example
is the last tide cycle (n=617-713) of this data set (see Fig. 3-14). During the early measuring period (n<666), sediment concentration peaks occurred at low water level. Therefore these periods contribute little to the sediment transport rate because of the low water depth. During the last measurement period (n>666), peak sediment concentration occurred before the tide level reached its lowest level. In other words, peak sediment concentrations matched relatively large water depths. This difference is the main reason for the increase of sediment transport rate during the last sampling period. Although the mechanics of this phase shift are not clear, a possible reason is that wave shear stress caused by surface wind added to the total bottom shear stress because of the onset of the summer monsoon season during the last sampling period.

Finally, the results of calculations show that the net sediment transport rate is seaward (see Fig. 3-15). This net seaward sediment transport can be attributed to the following factors:

1. Flank slumping produces the main component of channel sediment transport rate owing to groundwater levels falling in mudflats during ebb.

2. Mudflat channels and their tributaries are main conduits for seaward sediment movement during ebb periods, but sediment can be carried onto the mudflat surface from all directions during flood periods.

3. Wind-wave associated shear stress suspends additional sediment on the mudflat surface and this additional material also contributes to ebb sediment transport rate.
Fig. 3-14 Comparison of sediment concentration with tidal level and current velocity at station 2 during the last time period (from 17:00 6/6 (n=620) to 15:00 6/9 (n=712)).
Fig. 3-15 Temporal variation of the accumulative sediment transport rate, $Q_a$, (upper figure) and the sediment transport rate, $q$, (lower figure) at station 2 during 0:00 5/11 - 15:00 6/9 (total 712 hours). Positive values are seaward and negative ones are landward.
§3.5 A proposed conceptual model of sediment transport on the mudflat

As mentioned above, tidal flats of the west coast of South Korea receive little or no local river-derived sediment. Any sediment added to the system, thus, must come from the Yellow sea. The accumulation rate on the mudflat is about 0.15–2.0 cm yr\(^{-1}\) (Wells et al., 1990). Our observations from a mudflat channel, however, suggest that the direction of the net sediment transport is seaward. These two phenomena appear to be contradictory.

There is a widely accepted conceptual model called the "settling-scour lag" model (Postma, 1954; Van Straaten & Kuenen, 1958). The core of the model is that because of flocculation of sediment, both critical initial velocity and settling velocity of the sediment will be altered. The \textit{in situ} settling velocities of fine-grained sediment increase with the suspended sediment concentration due to the formation of sediment flocs (Owen, 1971). Therefore, net suspended sediment transport is seaward during periods of low suspended sediment concentration levels and landward during periods of high suspended sediment concentration levels (Pejrup, 1982). This model can explain suspended sediment transport processes on the mudflat surface well. However it may not explain the processes in the mudflat channel. Data presented in this study indicate that net sediment transport in a seaward direction occurs during the period of high suspended sediment concentration levels (see Fig. 3-15). Through analyzing the measured data, a new conceptual model of sediment transport processes on mudflats is proposed; it can be described as follows.
At low tide, the mudflat and part of the distributary channel beds are out of the water. As the tide floods, the sediment carried by the current from the bay will flow onto the mudflat. At high tide when velocities are small, sediment will be deposited on the mudflat surface. As the tide begins to ebb, the ebb tidal current will carry part of the sediment on the mudflat into the nearest channel tributary. With the falling of the tide, the mudflat surface is exposed first. Under the effects of sunlight and winds, the mudflat surface may consolidate and become more difficult to erode. So sediment can continue to accumulate on the mudflat surface. Combined with tidal level variation, sediment movement on the mudflat can be divided into three periods.

1. Consolidation/exposure period

When the mudflat and/or channel bed are exposed, the fine sediment, especially cohesive sediment, will consolidate, which means its water content will decrease and dry density will increase. Thus, erosion-resistance strength of the sediment will increase; mudflat surface will become less erodible. Krone (1972) found that the initial resuspension velocity of this flocculated, consolidated sediment is one order of magnitude higher than its deposition velocity, which is the cause of "scour-lag". For the mudflat, this corresponds to period I and II (or VI, VII), and for the channel bed, it corresponds to period I (or VII) only (Fig. 3-16). So the mudflat has a longer exposure period than the channel bed during each tidal cycle.
Fig. 3-16 Sediment movement on the mudflat can be divided into three periods: 1) Consolidation/exposure period---Low water level (mudflat: period I(VII) and II(VII); channel bed: period I (VII)). 2) Sediment carrying/erosion period---Maximum flow velocity (mudflat: period III and V; channel bed: period II, III, V & VI). 3) Depositional period---Slack water and high water level (mudflat: period IV; channel bed: period IV).
2. Sediment carrying/erosion period

With increasing tide levels, the current velocity will increase, which leads to transport of sediment onto the mudflat. A similar situation occurs when tide level falls from its peak level with net seaward sediment transport. The period of rising/falling tidal level is a period of large sediment transport. From the in situ measurements, it has been found that sediment transport is remarkably different during the rising and falling periods although flow velocities may be nearly the same. The characteristics of these periods mainly include:

a. Large sediment concentration peaks occur after the velocity peaks of ebb flow.

b. Net sediment transport rate during ebb is much greater than during the flood period.

The erosion period of the channel (period II, III, V & VI) is longer than that of the mudflat (period III and V).

3. Deposition period

When an incoming tide reaches its highest level, tidal velocity approaches a minimum. During this period, suspended sediment will be deposited. Both mudflat and channel, have approximately the same duration of deposition (period IV). From the above description of the sediment transport process model, it can be seen that, although mudflat and channel experience the same three periods of sediment transport, mudflats have a longer consolidation/exposed period which may suggest that sediment
on the mudflat is more difficult to erode than that on the channel. Sediments on the mudflat are much finer than those in the channel bed. For the cohesive sediment, this consolidation effect is more obvious.

The core of this conceptual model can be summarized as follows: Marine sediment comes onto the mudflats from all directions, then it is deposited on the mudflat surface. Through ebb erosion (wind wave, upstream erosion and slumping), sediment flows out mainly through mudflat channels and their tributaries, which results in an ebb sediment transport rate larger than flood sediment transport rate within the channels. Settling and scour lag play important roles in the sediment transport processes on the mudflat surface. However, sediment size difference between mudflat surface and channel bed, upstream erosion and slumping are the key features for sediment transport processes on mudflats.

The conceptual model presented here explores the sediment transport process on the mudflat. It can well explain the sediment accumulation on the mudflat surface with the net seaward sediment transport in the mudflat channel. But this model is only based on one-month of measured data. Long time-series data (> 1 year) are necessary to verify this model. Thus, the seasonal variation of sediment transport processes on the mudflat can be explored.

By examining sediment transport processes on the mudflat, it has been found that mudflat channels play key roles in the sediment transport budget. In order to clarify this sediment transport model for the mudflat, a knowledge of the relationship
of sediment resuspension, bottom shear stress and vertical velocity profile is crucial. As we know, all three factors are interactive. In the next chapter, we will focus our attention on these three aspects.

§3.6 Summary

In this chapter, sediment transport and its related features are the major topics, which can be summarized as follows:

1. The broad mudflats in Namyang Bay were formed under conditions of high tidal range and small fluvial sediment input. The sediment transport process was characterized by high accumulation rates on the mudflats with net seaward sediment transport rate. Sixteen in situ surface sediment samples indicated that the channel bed consists of sediment with a larger grain size than that on the mudflat surface, and grain size decreased in a downstream direction. Sediment transport on the mudflat, which is subjected to high tidal range variation, was strongly related not only to tidal current features but also to erosion and deposition characteristics of the sediment. High tidal range created the steep channel slope, relatively deep channel and finer grain sediment on the mudflat surface.

2. The reference concentration, a factor that characterizes sediment transport and the vertical concentration profile, varies with the bottom shear stress. For this study which mainly represents fair weather conditions, the resuspension coefficient was found to be, $\gamma_0 = 3.65 \times 10^4$. When the wind wave shear stress is significant, this kind of linear relation between reference concentration and relative shear stress is not valid.
3. Because the bottom shear stress in the channel is much higher than the critical shear stress of bed sediment, little suspended sediment is deposited on the channel bed. In other words, suspended sediment does not exchange with the bed sediment significantly.

4. During the neap tidal period, unstable stratification was frequently observed. This observation suggests that the fine-grained suspended sediment contributed to the surface layer occurred by slumping of sediment at the channel banks.

5. There exist three types of roughness length variations with bottom shear stress: proportional, constant, and reverse proportional. In this study, roughness length generally remained constant. A comparison with other data suggests that sediment compositional changes could exert an important influence on roughness length. Therefore, the sediment composition variation is another key factor in the determination of the roughness length besides the bottom shear stress.

6. In this study, the drag coefficient was relatively constant ($C_d=4.215\times10^{-3}$), which may be a consequence of the relatively invariant nature of sediment composition and bed forms. Thus the bottom shear stress will be better correlated with flow velocity, which has great practical importance for the determination of bottom shear stress.

7. Velocity profiles within the mudflat channel vary with the tidal cycle. Well-fit logarithmic velocity profiles usually were observed during the peak flood velocity period. Generally the fit worsened as the flow decelerated.
8. During the fair-weather season, the sediment transport rate in the mudflat channel correlates well with the residual current. During strong wind period, however, the relationship is less obvious. Because of wind-induced wave shear stress, the sediment transport rate will increase dramatically although residual current changes little.

9. Net sediment transport rate during ebb is much greater than during flood. Upstream erosion of the channel bed, flank slumping, and inflow of sediment from the mudflat surface provide sources of sediment for seaward sediment transport.

10. Sediment transport on the mudflat can be divided into three periods during a tidal cycle: 1) a consolidation/exposure period, 2) a sediment carrying/erosion period, and 3) a deposition period. The final results of these three periods are the formation of a fine grain mudflat surface, an erosion-resistant surface owing to long exposure time, and coarser material in the channel bed. The erosion of the mudflats mainly consists of upstream erosion in the channel and flank slumping at channel banks. The meandering tributary channels are the ultimate result of the channel erosion and mudflat deposition.
Chapter 4

Effect on the bottom turbulent boundary layer of stratification caused by sediment resuspension
§ 4.1 Introduction

In Chapter 2, three methods for estimating the bottom shear stress were summarized. The "inertial dissipation" method is relatively efficient and stable. Through the work in Chapter 3, it has been realized that the sediment resuspension can influence the bottom shear stress. Based on the above observations, in this chapter, the "inertial dissipation" method will be developed further to include the effect of sediment resuspension on the bottom shear stress.

The "inertial dissipation" method is based on three assumptions (e.g., Huntley, 1988a; Green, 1992). The first is that only a neutrally stratified turbulent flow exists in the wall layer. Thus the dissipation of turbulent kinetic energy by viscosity is balanced by energy (shear) production with the assumptions of local isotropic, horizontally homogenous, and stationary surface-layer turbulence conditions. When bed sediment is resuspended, and a vertical sediment concentration gradient develops, buoyancy becomes important and the assumption may not be valid. The second assumption is that the "law of the wall" was introduced into the derivation of shear production. This law only holds for a neutrally stratified flow. For a non-neutrally stratified flow, an universal function related to buoyancy production has to be added (Panofsky, 1963; Businger et al., 1971). The third assumption is that turbulent velocities have to be measured within the constant stress region of the logarithmic layer in order to satisfy the relation $u_r^2 = \langle w' u' \rangle$. Moreover, the Reynolds number has to be large enough for the existence of an inertial subrange (T&L, p266). The "inertial
dissipation" method was established on the above assumptions. The question is how to modify the method when the buoyancy term can not be neglected.

Of the three methods for estimating time-averaged bed shear stress, the velocity measurements have to be made within the constant stress layer for the "inertial dissipation" method and the "eddy correlation" technique. For the "shear velocity" method, the sensors must be within the logarithmic layer which is approximately twice the thickness of the constant stress layer (e.g., Huntley, 1988a). Thus the determination of constant stress layer structure will be crucial. The results from this study, introduced in Chapter 2, indicated that the thickness of the constant stress layer was overestimated by the present calculation for the current in the inter-tidal channel.

Mean shear, Reynolds stress and rate of energy dissipation are related (Gross and Nowell, 1985). During sediment resuspension, buoyancy production may be another important factor (West et al., 1990). The transport of fluctuating turbulent energy, turbulent momentum flux, and temperature variance are usually dominated by buoyant transport (Lumley & Zeman, 1978). Thus, the evaluation of buoyancy flux is important for studying turbulent energy transport.

This chapter is based on analyses of the high frequency velocity, temperature and sediment concentration data collected at station 3 and the long-time-series mean flow velocity and sediment concentration data at station 2. Delineation of a constant stress layer requires knowledge of the vertical distribution of momentum flux, eddy viscosity and mixing length. The in situ measured data will be used for this delineation. The vertical and streamwise velocity fluctuation data and their covariance,
namely momentum flux, are necessary to determine all three terms. Multi-level mean velocities are of importance for obtaining the vertical velocity profile. Our seven-levels of velocity fluctuation data satisfy these requirements.

A modified "inertial dissipation" method will be developed based on considerations of buoyancy production in the kinetic energy budget and non-neutral conditions in the "law of the wall". The new method will be reflected in the form of the flux Richardson number, which represents the ratio of buoyancy to shear production. The two-level sediment concentration data at Station 2 will be used for estimating the flux Richardson number.

The objectives of this chapter are to gain insight into 1) the nature of the constant stress layer and verification of its thickness; 2) the role of buoyancy in turbulent energy transport; and 3) the modification of the present "inertial dissipation" method under influence of the non-neutral stratification induced by suspended sediment.

§ 4.2 Determination of constant stress layer

Since data measurements have to be taken within the constant stress layer for the "inertial dissipation" method to be valid, a full understanding of the constant stress layer will be crucial for estimating bottom shear stresses. In Chapter 2, the uniform Reynolds stress distribution in the lower part of the BBL was presented. Here, the vertical distribution of mixing length and eddy viscosity based on in situ measurements will further indicate the existence of the constant stress layer and characterize its nature. Before doing that, the presence of turbulent isotropic characteristics will be verified because this is a very primary assumption for turbulence data analyses.
4.2.1 The examination of isotropicity of turbulence

If the turbulence is isotropic, the relationship, \( u'^2 = v'^2 = w'^2 \), has to be satisfied (Cebeci, 1974, p13), which can be used to indicate turbulence isotropy. By examining data from a typical tidal-cycle (see Fig. 2-2), the relationships between the variances of \( u' \) and \( w' \) (\( \sigma_u \) and \( \sigma_w \)) have been determined (see Tab. 4-1). An examination of Table 4-1 reveals the following features:

1. For the original data, \( \sigma_u \) is larger than \( \sigma_w \) by a ratio from 1.2 to 5.4. The data suggest that the difference depends mainly on the width of the turbulent frequency band (or wave number band). The dominant frequency band of streamwise velocity fluctuations is much lower than that of vertical velocity fluctuations. Removal of the lowest frequencies would significantly reduce the differences. Now the question is what is the width of the turbulent frequency (or wave number) band. If these boundaries can be set, the lower frequency part of the raw data can be filtered out. Table 4-2 shows the comparison of \( w' \) and \( u' \) variances after high-pass filtering (\( f > 1 \) Hz). The differences between them can be seen well.

2. For smaller mean velocities (\( u_{\text{mean}} < 20 \text{ cm/s} \); see Tab. 2-4 for corresponding mean velocities), \( \sigma_u \) and \( \sigma_w \) are relatively close (\( \sigma_u/\sigma_w = 1.2-2.8 \)). For larger mean velocities, the ratio of \( \sigma_u \) to \( \sigma_w \) is larger (\( \sigma_u/\sigma_w = 1.7-5.4 \)). This indicates that higher mean velocities may contribute selectively to the high frequency components of current velocities.

3. Because of sensor problems, little lateral velocity data were acquired. The data sets k1514 and k1515 (refer to 5/15 14:00 & 15:00), have been chosen for
Table 4-1 Comparison of variances of \( u' \) and \( w' \) in a tidal cycle

(5/15 12:00-19:00, 1992)

<table>
<thead>
<tr>
<th>Time</th>
<th>( \sigma_{u4} )</th>
<th>( \sigma_{w4} )</th>
<th>( \sigma_{u5} )</th>
<th>( \sigma_{w5} )</th>
<th>( \sigma_{u7} )</th>
<th>( \sigma_{w7} )</th>
</tr>
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<tr>
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<td>1.35</td>
<td>6.38</td>
<td>1.19</td>
<td>3.30</td>
<td>1.59</td>
</tr>
<tr>
<td>13:00</td>
<td>5.46</td>
<td>2.10</td>
<td>5.28</td>
<td>2.30</td>
<td>4.40</td>
<td>2.18</td>
</tr>
<tr>
<td>14:00</td>
<td>3.20</td>
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<td>2.78</td>
<td>1.16</td>
<td>1.78</td>
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</tr>
<tr>
<td>15:00</td>
<td>2.97</td>
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<td>2.78</td>
<td>1.02</td>
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<tr>
<td>16:00</td>
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<td>0.88</td>
<td>0.41</td>
<td>0.62</td>
<td>0.39</td>
</tr>
<tr>
<td>17:00</td>
<td>2.08</td>
<td>1.27</td>
<td>2.02</td>
<td>1.26</td>
<td>1.40</td>
<td>1.17</td>
</tr>
<tr>
<td>18:00</td>
<td>3.00</td>
<td>1.15</td>
<td>2.90</td>
<td>1.16</td>
<td>2.66</td>
<td>1.12</td>
</tr>
<tr>
<td>19:00</td>
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<td>1.37</td>
<td>3.14</td>
<td>1.10</td>
<td>2.80</td>
<td>1.61</td>
</tr>
</tbody>
</table>
Table 4-2 Comparison of variances of $u'$ and $w'$ with high-pass filtering 
($f > 1$ Hz) (5/15 12:00-19:00, 1992)

<table>
<thead>
<tr>
<th>Time</th>
<th>$\sigma_{u4}$</th>
<th>$\sigma_{w4}$</th>
<th>$\sigma_{u5}$</th>
<th>$\sigma_{w5}$</th>
<th>$\sigma_{u7}$</th>
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</thead>
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<td>0.89</td>
<td>0.37</td>
<td>0.41</td>
<td>0.36</td>
</tr>
<tr>
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<td>0.61</td>
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<td>0.39</td>
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</tr>
<tr>
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<td>0.17</td>
<td>0.23</td>
</tr>
<tr>
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<td>0.27</td>
<td>0.22</td>
<td>0.16</td>
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<tr>
<td>16:00</td>
<td>0.18</td>
<td>0.17</td>
<td>0.41</td>
<td>0.20</td>
<td>0.16</td>
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<tr>
<td>17:00</td>
<td>0.21</td>
<td>0.23</td>
<td>0.30</td>
<td>0.30</td>
<td>0.18</td>
<td>0.22</td>
</tr>
<tr>
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<td>0.28</td>
<td>0.33</td>
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<td>0.23</td>
<td>0.27</td>
</tr>
<tr>
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<td>0.48</td>
<td>0.38</td>
<td>0.29</td>
<td>0.25</td>
</tr>
</tbody>
</table>
113

comparison ($\sigma_x = 0.41$ and 0.32). The spectral distribution of lateral velocities are consistent with the other two velocity components (Fig. 4-1 (1),(2)).

4.2.2. Vertical distribution of mixing length and eddy viscosity

In order to describe the mean-velocity distribution in the bottom boundary layer, the concepts of mixing length and eddy viscosity are widely used to relate the Reynolds stress to the local mean-velocity gradient (Cebeci & Smith, 1974). Based on the mixing length theory of Prandtl (1933), the mixing length, $l_m$, can be expressed as:

$$l_m = \sqrt{\langle w' u' \rangle} \left( \frac{\partial u}{\partial z} \right) = \sqrt{\langle w' u' \rangle} \left( \frac{k_z}{u_*} \right). \quad (4-1)$$

Using dimensional analysis, Tjemstrom (1993) set a length scale form for momentum mixing as $l_m = u_* \partial u/\partial z^{-1}$, which is the same as Eq. (4-1) when $\sqrt{\langle w' u' \rangle}$ is associated with $u_*$. With momentum flux data, the vertical distribution of mixing length can be obtained. This length scale is a characteristic length associated with the large energetic eddies of turbulence (Lesieur, 1987).

According to the concept derived by Prandtl, the turbulent stresses are proportional to the velocity gradients. The coefficient of proportionality was called the "eddy viscosity" because it is similar to the molecular viscosity. The eddy viscosity is defined as:

$$A_z = \langle w' u' \rangle \frac{\partial u}{\partial z} = \langle w' u' \rangle \left( \frac{k_z}{u_*} \right). \quad (4-2)$$
Fig. 4-1(1) Temporal variation of three components of flow velocity fluctuation, $u'$, $v'$, $w'$, selected from the data of station 3. The sampling frequency is 4 Hz. The measuring time is 512 seconds (corresponding to 2048 data).
Fig. 4-1(2) Comparison of spectrum for the three components of flow velocity fluctuation: $u'$, $v'$, $w'$, measured at station 3.
While the two expressions yield different absolute values, the vertical distributions are similar within the constant stress layer. Actually, the relationship between eddy viscosity and mixing length is 

\[ A_z = l_m(u_z/kz) \]

Applying measured data to the above equations, the vertical profiles of both mixing length and eddy viscosity can be seen (Fig. 4-2). Some interesting features of the \( l_m \) and \( A_z \) profiles are summarized as follows:

1. Linear distributions exist only in the bottom region \((z<150 \text{ cm})\) where both parameters increase with elevation. For the smaller values of \( l_m \) and \( A_z \), the distributions are relatively uniform.

2. The relationship between eddy viscosity and mean current velocity can be classified into three regions: 1) Eddy viscosity is low and varies directly with flow velocity when velocity is less than 10 cm/s; 2) For 10 cm/s < \( U < 50 \) cm/s, eddy viscosity decreases as flow velocity increases; 3) When velocity exceeds 50 cm/s, a direct proportionality relation exists between eddy viscosity and mean flow velocity (see Fig. 4-3). Analysis of mixing length variations yields similar results.

3. Both mixing length and eddy viscosity depend strongly on momentum flux. Accurate measurements are essential to obtain better vertical profiles of these parameters. In this study, the lower frequency wave peaks are found frequently in the velocity spectrum, thereby limiting data usefulness.
Fig. 4-2 Vertical distribution of mixing length, $l_m$ (left figure) and eddy viscosity, $A_z$ (right figure) in the lower portion of bottom boundary layer ($z < 200$ cm), at station 3 during 12:00-19:00 5/15, 1992 (total 8 data).
Fig. 4-3 The relation of eddy viscosity with current velocity at station 3 during 12:00-19:00 5/15, 1992. The values: u3, u4, u5, and u6, refer to the mean flow velocities at different levels (z= 45, 70, 100, and 150 cm above bottom).
4.2.3 Determination of the constant stress layer

Three vertical length scales are associated with a turbulent bottom boundary layer without wave effects: an Ekman layer, a logarithmic layer, and a constant stress layer. The determination of the bottom shear stress depends strongly on the constant stress layer. But the constant stress layer is closely related to the whole bottom boundary layer. Previous studies have indicated that the thickness of a logarithmic layer is about 10% of the whole boundary layer thickness (Hinze, 1975; Businger & Arya, 1974). The constant stress layer is widely held to be thinner than the logarithmic layer. Tennekes (1973) and Huntley (1988a) have presented information to show that the thickness of a constant stress layer is about one-half that of the logarithmic layer. Therefore, the thickness of the constant stress layer can be estimated as

\[ Z_{cs} = \frac{1}{2} Z_{log} = \frac{1}{2} \times 0.1\delta = 0.05\delta \] (4-3)

where \( \delta \) is the thickness of the bottom boundary layer.

Blackadar and Tennekes (1968) developed a method for the determination of the thickness of Ekman layer in a steady-state horizontal flow. One can write:

\[ f(v - v_e) + \frac{d (\tau) }{dz} = 0 \] (4-4)

\[ f(u - u_e) + \frac{d (\tau) }{dz} = 0 \] (4-5)
where \( u_g \) and \( v_g \) are the longitudinal and lateral geostrophic flow velocities; \( \tau_x \) and \( \tau_y \) are the longitudinal and lateral shear stresses. The above equations are actually for flow in an Ekman layer. The total depth of this layer can be estimated as

\[
h = u_o^2 / f W \quad (4-6)
\]

where \( W \) is the velocity scale, the magnitude of which is between \( u_o \) and the free stream geostrophic velocity. In order to determine \( W \), a non-dimensional coefficient, \( A_o = u_o / W \) is set. Thus, equation (4-6) becomes:

\[
h = A_o \frac{u_o}{f} \quad (4-7)
\]

where \( h = \delta \). In practice, \( \delta \) is hard to obtain. The results from this study indicate that the constant stress layer is generally within the region: \( 100 \text{ cm} < z < 150 \text{ cm} \). Using the mean friction velocity \( (u_o = 1.47 \text{ cm/s}) \) and local Coriolis parameter \( (f = 8.77 \times 10^{-5} \text{ s}^{-1}) \) obtained above and equating Eqs. (4-3) and (4-7), the thickness of boundary layer can be obtained as

\[
\delta = (0.12 - 0.18) \frac{u_o}{f} \quad (4-8)
\]

In the real world, a pure Ekman layer is seldom observed. The concept of the Ekman layer, however, does reflect the existence of a bottom boundary layer. Therefore the thickness of the Ekman layer is approximately equal to the thickness of the bottom boundary layer. In previous studies, this thickness constant, \( A_o \), was widely taken to lie within the range from 0.20 to 0.40 (Panofsky, 1984; Huntley, 1988; Weatherly and
Weatherly and Martin (1978) found that the oceanic BBL thickness should be identified with the height at which the turbulence generated in the BBL goes to zero, and the BBL thickness for the stratified flow case \((N^2>0)\) is considerably less than 0.4 \(u/f\). In this study, the value is taken as 0.12-0.18. The lower value, however, does not mean that it is more accurate than the previous one because they have been obtained from radically different environments. The method presented in this study is a rough estimation, but results agree well with the lower thickness constant. Since the form of Eq. (4-8) is widely used, this relation is retained to indicate that the boundary layer thickness will be reduced in the stratified flow. Of course, additional studies are required to verify the lower value of the thickness constant.

§ 4.3 Turbulent kinetic energy transport balance in bottom boundary layer

4.3.1 Temperature data and sediment concentration data

The acquisition of temperature data and sediment concentration data has been described in chapter 2. Here the temperature and concentration data will be used to obtain the buoyancy production.

The temperature data were measured at five different levels (35, 70, 125, 175, 200 cm above bottom). The temperature and flow velocity were sampled simultaneously (12:00 5/15–3:00 5/17, 1992). The time series of the five-level temperature data are plotted in Fig. 4-4. By comparing temperature with the mean flow velocity, some relationships can be found:

1. Unlike the turbulent velocity data, the high frequency temperature data have obvious trends strongly related to the direction of the tidal current. When the flood
Fig. 4-4 Temporal variation of 5-level water temperature data (z=10, 25, 45, 70, 125, 175, 200 cm above bottom) compared with corresponding mean flow velocity $u_2$ (z=25 cm above bottom) at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (total 40 hours)
tidal current velocity reaches its maximum value, water temperature is falling; when the ebb current velocity reaches its maximum value, water temperature approaches a maximum value too. This indicates that water temperature in the bay is higher than the offshore water temperature during the sampling period.

2. The relation of water temperature and tidal current velocity can be summarized as: the maximum flood velocity coincides with the smaller peak of water temperature; and maximum ebb velocity matches the larger peak of water temperature.

3. The water temperature is the lowest during both high and low slack waters.

The above relationships are only based upon our one-month-long data. Longer time-series data are required to verify their seasonal variations.

The sediment concentration data were originally from the two-levels of transparency data. The relationship between transparency, Tr, and suspended sediment concentration, C, has been established (Wells et al., 1992)

\[ C = 10^{2.31 - 0.017r} \tag{4-9} \]

where the unit of C is mg/l. Thus the temporal and vertical variations of sediment concentration with mean flow can be seen (Fig. 4-5). Some basic relationships are summarized as follows:

1. There is no significant difference between the sediment concentration at the two levels, which shows the vertical sediment concentration is nearly uniform within this region.
Fig. 4-5 Temporal variation of 2-level sediment concentrations, C1 (z= 100 cm) and C2 (z=200 cm) compared with corresponding mean flow velocity u5 (z= 100 cm) at station 3 during 12:00 5/15 - 3:00 5/17, 1992 (total 40 hours).
2. There exists a clear temporal variation in sediment concentration between the flood period and ebb period. The sediment concentration peak during the ebb period (C= 200 mg/l) is 30% larger than that in the flood period (C=150 mg/l). This agrees well with the results from the data of station 2. During the ebb period, the channel current carries a large amount of sediment eroded from the mudflat.

4.3.2 Examination of the kinetic energy balance equation using in-situ measured data

Under horizontally homogeneous and stable stratified conditions, the steady-state energy budget can be simplified into (Wyngaard & Cote, 1971):

\[ -\langle u'w' \rangle \frac{\partial U}{\partial z} + B - \varepsilon + \lambda_n = 0 \]  \hspace{1cm} (4-10)

where B is the buoyancy production, which could be a source or a sink depending on the sign of the vertical density flux; \( \lambda_n \) is called the imbalance term which contains turbulent transport, pressure transport and all possible inhomogeneity effects.

In order to evaluate each term in equation (4-10) using the measured data, the above equation can be further modified. The dissipation term, \( \varepsilon \), can be expressed either by relationships associated with the "law of the wall" or by the turbulence scale. For the former, it can be expressed as \( \varepsilon = u_z^3/\kappa z \); for the latter, as \( \varepsilon = u_z^3/l \) (Chapter 2). In the inter-tidal channel, buoyancy production was mainly associated with temperature and sediment concentration gradients. For temperature, the buoyancy term can be expressed as \( B_t = g \beta \langle \theta' w' \rangle \), where \( \beta \) is the coefficient of thermal expansion; \( \theta' \) is the temperature fluctuation. In a fluid-sediment mixture, the density of the mixture can
be expressed as $p_a = p + (p_s - p)c'$. Thus, the component of buoyancy production associated with sediment concentration gradients will be $B_c = g<\rho_m w'>/\rho$. Substituting the above two relations into the buoyancy term, total buoyancy production, $B$, can be expressed as

$$B = B_s + B_c = -g[\beta<\theta' w'> - \frac{(p_s - p)}{\rho} <c' w'>]. \quad (4-11)$$

Thus the kinetic energy budget equation can be rewritten as

$$-\langle u' w' \rangle \frac{\partial U}{\partial z} - \frac{u_+^3}{kz} \left[ -g[\beta<\theta' w'> - \frac{(p_s - p)}{\rho} <c' w'>] + f_n \right] = 0, \quad (4-12)$$

or

$$-\frac{\langle u' w' \rangle^{3/2}}{kz} \frac{u_+}{kz} \left[ -g[\beta<\theta' w'> - \frac{(p_s - p)}{\rho} <c' w'>] + f_n \right] = 0. \quad (4-13)$$

Based on the above equation, estimates of the kinetic energy budget are shown in Fig. 4-6. Several important points are summarized as follows:

1. The shear production term has the same order of magnitude as the dissipation rate term. Buoyancy production contributed by the sediment concentration gradient is surprisingly small, and has almost the same order as the contribution by the temperature gradient. A more detailed discussion is given in the next section.

2. Shear production can be divided into three regions (see Fig. 4-6). In the lower portion of the boundary layer ($z<0.3 \text{ m}$), shear production is maximum and decreases upward dramatically. In the middle part of the boundary layer ($0.3 \text{ m} < z < 0.7 \text{ m}$), the vertical gradient of shear production decreases gradually. In the upper
Fig. 4-6 Vertical distribution of shear productions (left figure) and energy dissipation (right figure) at station 3 during 12:00 - 19:00 5/15, 1992.
region (z > 0.7 m), the shear production is relatively constant and close to zero. Compared with shear production, the dissipation rate is relatively uniform. With respect to the mean flow velocity, shear production is dominant in the lower part, and dissipation rate is dominant in the upper part of the boundary layer (Fig. 4-7). It is found that turbulent energy production is greatest in the high shear region where a large velocity gradient exists. This relation is also valid for the meteorological case although the mean velocity profile may be totally different (Smedman et al., 1994).

4.3.3 On the order of the buoyancy term

According to the above analysis, the buoyancy term contributes only a small part of the total energy balance. Most of the buoyancy term is associated with temperature gradients and sediment concentration gradients (Eq. (4-11)). Now let us analyze the scale of these two terms. For the temperature gradient-induced buoyancy term, \( g = 981 \text{ cm s}^{-2} \), and the coefficient of thermal expansion, \( \beta = 2.31 \times 10^{-4} \text{ °C}^{-1} \) at the mean temperature of 17.4°C and salinity of 32 % (Newman, 1966, Tab. 3.1). Thus

\[
B_t = g\beta \langle \theta' w' \rangle = 981 \times 2.31 \times 10^{-4} \langle \theta' w' \rangle = 2.27 \times 10^{-1} \langle \theta' w' \rangle
\]

For the sediment concentration gradient-induced buoyancy term, \( g[(\rho_s - \rho)/\rho] = g' = 981 \times (2.65-1.0)/1.0 = 1618 \text{ cm s}^{-2} \). In order to keep dimensional agreement, the unit of concentration should be changed from mg/l to vol/vol. A multiplicative conversion coefficient, \((2.65 \times 10^6)^{-1}\), is used. Therefore,

\[
B_c = (g'/2.65 \times 10^6) \langle c' w' \rangle = (1618/2.65 \times 10^6) \langle c' w' \rangle = 6.1 \times 10^{-4} \langle c' w' \rangle.
\]

The data presented in this study indicate that the flux \( \langle c' w' \rangle \) is 10–100 times larger than \( \langle \theta' w' \rangle \). Thus \( B_c \) has approximately the same order as \( B_t \) or less. This analysis
Fig. 4-7 The detailed comparison of shear productions (Solid lines: P5, P6, P7, and P8) and energy dissipations (Broken lines: E5, E6, E7, and E8) with the corresponding mean flow velocity environment at station 3.
indicates that suspended sediment did not contribute dominantly to the buoyancy term.

There are three reasons for this. First, during the sampling period, there were no significant wind-induced waves acting on the sea bed. The long time series of sediment concentration data from station 2 showed that large sediment suspensions first appeared in June, the beginning of summer monsoon season. Second, the vertical sediment concentration profile is relatively uniform, which makes the mass flux, \( <c'w'> \), small. The third is that the transparency sensors were far from the bottom (lowest sensor was 100 cm above bottom). The dominant concentration gradient mainly exists near the bottom. Away from the bottom, the concentration gradient will dramatically decrease.

4.3.4 Validation of Kolmogorov law

A necessary condition for the existence of an "inertial subrange" is that the turbulent velocity wave number, \( K_\omega \), must be within the region:

\[
K_i \ll K_\omega \ll K_d
\]

(Lesieur, 1987), where \( K_i \) is the characteristic wave number of the "large energy containing eddies", and \( K_d \) is the wave number where viscous effects become important. When \( K_i < K_\omega \), the spectrum shape will be influenced by non-local external forces. When \( K_i > K_d \), the lower portion of the spectrum can be neglected (Lesieur, 1987). George (1994) set the critical frequency boundary between the orbital wave regime and the inertial subrange as \( f_c = (gh)^{1/5} l \), where \( l \) is the turbulent length scale described in chapter 2. Based on the field measurements, the lower boundary of the inertial subrange can also be set.
Under condition of non-neutral stratification, flow dynamics are governed mainly by three forces, i.e., buoyancy, inertia and viscosity. Buoyancy creates an upper ceiling for the growth of turbulent eddies, while viscous forces limit the size of the smallest overturning eddies (Itsweire & Helland, 1989). The range of possible overturning turbulent scales is described by the inequality: \(1.2 \, L_i \geq \lambda \geq 15 \, L_i\), where \(L_i = (\varepsilon/N_i)^{1/2}\) is the Ozmidov scale (Dougherty 1961; Ozmidov 1965) which scales the action of buoyant forces, and \(L_o = (v^3/\varepsilon)^{1/4}\) is the Kolmogorov scale. With the relation \(K = 2\pi/L\), thus, the wave number extent of the inertial subrange can be set as:

\[
K_d = (2\pi/15)(v^3/\varepsilon)^{1/4} \quad \text{and} \quad K_i = (2\pi/1.2)(\varepsilon/N_i)^{1/2}
\]

Now let us look into the above two values based on our field data. The value of the kinematic viscosity, \(v\), is \(1.08 \times 10^{-5} \, \text{cm}^2 \, \text{s}^{-1}\) at the average temperature of \(17.46^\circ\text{C}\); The energy dissipation rate, \(\varepsilon\), is taken as its average value of \(0.37 \, \text{cm}^2 \, \text{s}^{-3}\). As a rough estimation, the Brunt-Väisälä frequency \(N\) can be obtained from the relation: \(N^2 = B/K_z\), where \(B\) is buoyancy production with the average value of \(0.0034 \, \text{cm}^2 \, \text{s}^{-3}\); The eddy diffusivity, \(K_z\), could be assumed to be the same value as eddy viscosity, namely, \(A_z/K_z = 1\). Lumley & Panofsky (1964, p65) believed that this is likely to be true in a region of large production and dissipation of fluctuations such as a boundary layer. Based on the calculation of \(A_z\), \(K_z\) was taken as \(30 \, \text{cm}^2 \, \text{s}^{-1}\). Finally the two values can be obtained as \(K_d = 9.74 \, \text{cm}^{-1}\) and \(K_i = 0.01 \, \text{cm}^{-1}\). The spectral plots presented in chapter 2 indicated that the "inertial subrange" always fell within this region and frequently was within the range \([0.05, 2.0] \, (1/\text{cm})\).
§ 4.4 The effect of sediment resuspension on the bottom shear stress

4.4.1 Derivation of a basic model based on the modification of the "inertial dissipation" method

As mentioned previously, the present "inertial dissipation" method is only valid when the buoyancy production can be neglected, namely under neutral stratification. In non-neutral stratification, the effect of buoyancy production should be considered. To incorporate buoyancy, the following method is proposed:

1. When buoyancy production is not negligible, the steady, horizontally homogeneous kinetic energy balance equation can be written as

$$-\langle u'w' \rangle \frac{\partial U}{\partial z} - e - \frac{g}{\rho} \langle p'w' \rangle = 0 \quad (4-14)$$

In terms of a flux Richardson number, buoyancy production can be expressed as

$$\frac{g}{\rho} \langle p'w' \rangle = R_f \langle w'u' \rangle \frac{\partial U}{\partial z} \quad (4-15)$$

Substituting Eq. (4-15) into Eq. (4-14), the kinetic energy balance equation becomes

$$-\langle w'u' \rangle \frac{\partial U}{\partial z} (1 + R_f) - e = 0 \quad (4-16)$$

2. If the turbulent velocity was measured within the constant stress region of the logarithmic layer of an unstratified fluid, then the "law of the wall"

$$\frac{\partial U}{\partial z} \kappa z = 1 \quad (4-17)$$

is valid. When buoyancy forces act in this region, the velocity profile will be a function of Richardson number (Panofsky, 1984), namely,
where \( \Phi(R_f) \) is an universal function of Richardson number.

3. By substituting the mean current scale, Eq. (4-18), to Eq. (4-16), the new relation becomes

\[
\frac{u_* \Phi(R_f)}{\kappa z} = \frac{\varepsilon}{u_*^3(1 + R_f)}. \tag{4-19}
\]

or

\[
\varepsilon = \frac{u_*^3}{\kappa z} \Phi(R_f)(1 + R_f). \tag{4-20}
\]

Here \( \Phi(R_f) \) can be expressed by

\[
\Phi(R_f) = \begin{cases} 
1 + 4.7R_f & \text{in stable situations} \\
(1 - 15R_f)^{1/4} & \text{in unstable situations}
\end{cases} \tag{Businger et al., 1971}
\]

(Businger et al., 1971). The two relations, however, could vary from case to case (Carl, 1973). Adams and Weatherly (1981) verified this expression in stable situations using oceanic bottom boundary layer data. Their result is \( \Phi(R_f) = 1 + 5.5R_f \), which is very close to the result derived by Businger et al. (1971) using meteorological data.

With the above derivation, the friction velocity in non-neutral stratification can finally be determined from

\[
u_* = (\kappa z \varepsilon)^{1/3} \Phi(R_f)(1 + R_f)^{-1/3}. \tag{4-21}
\]
From the above equation, it can be found that if \( R = 0 \), then \( u = \kappa z e^{1/3} \), which is the original relation between friction velocity and dissipation rate for neutral stratification.

Now the key problem is how to obtain the Richardson number, \( R^* \), which indicates the ratio of buoyancy production to shear production. There are several ways to obtain the Richardson number:

1) The momentum flux, \( \langle w'u' \rangle \), heat flux, \( \langle \theta w' \rangle \), and density flux, \( \langle \rho w' \rangle \), due to sediment can be measured directly. When the flux is known, the Richardson number can be obtained directly from its definition.

2) The Richardson number can also be calculated from the vertical profiles of mean velocity, temperature and sediment concentration. The procedure for this calculation can be summarized as:

The flux Richardson number can be expressed as a function of the Monin-Obukhov length, \( L \), as indicated by Adams & Weatherly (1981)

\[
L = \frac{u_*^3}{\kappa g \langle \rho' w' \rangle / \rho} = \frac{u_*^2}{\kappa^2 g z (\rho_s - \rho) \frac{\partial \delta_c}{\rho} \frac{\partial z}{\rho}} \tag{4-22}
\]

Following Green (1992), the sediment concentration gradient in the above equation can be determined from the vertical distribution of sediment concentration (Eq. 3-12). The Richardson number can then be determined from \( R^* = z/L \).

4.4.2 Application of the method

Unlike temperature gradient dominated stratification in the meteorological case, most of the flow stratification in an oceanic boundary layer is stable, especially for
sediment concentration gradient-induced stratification. For stable stratification, the expression for friction velocity, Eq. (4-21), will become

$$u_\ast = \left( \frac{\kappa z \varepsilon}{\lambda} \right)^{1/3} \left[ (1 + AR_f)(1 + R_f) \right]^{-1/3} \tag{4-23}$$

where the constant $A$ will be taken as 5.5; $\left[ (1 + AR_f)(1 + R_f) \right]^{-1/3}$ is the stratification coefficient, $K_s$. From Eq. (4-23), the effect of flow stratification on friction velocity can be seen. In the previous section, energy dissipation rate and Richardson number, calculated in several different ways, have been introduced. Thus, the calculation of this method is straightforward. Since ample sediment concentration profile data are available for this study, the Richardson number can be calculated directly. The calculation procedure can be divided into the three steps:

1) Calculation of the friction velocity at neutral stratification using the relation,

$$u_\ast = \left( \frac{\kappa z \varepsilon}{\lambda} \right)^{1/3}.$$

2) Calculation of the Richardson number by either Eq. (3-4) or $R_f = z/L$ and Eq. (4-22), where sediment concentration profiles can be determined by the Rouse formula, Eq. (3-12).

3) Calculation of friction velocity at non-neutral stratification using Eq. (4-23).

Because this calculation starts from the neutral stratification condition, in order to obtain accurate results, an iterative computation is recommended. The results for the Richardson number and stratification coefficient, $K_s$, are shown in Fig. 4-8. As mentioned in Chapter 2, when non-neutral stratification was considered, the vertical
Fig. 4-8 The vertical distribution of Richardson number (upper figure) and stratification coefficient $K_s$ (lower figure) at station 3 during 12:00 5/15 - 3:00 5/17 (total 40 hours).
friction velocity profiles were relatively uniform (Fig. 4-9). Finally, Some noteworthy points of this modification model should be mentioned:

1. The presence of a sediment concentration gradient will reduce the bottom shear stress. Table 4-3 lists the comparison of mean shear stress with and without the consideration of stratification. When 0.02 < Rf < 0.35, the bottom shear stress can be reduced by 6-35% (see Fig. 4-10). This result is very close to the results obtained by Adams and Weatherly (1981), using the turbulence closure model of Mellor-Yamada (1974).

2. A large mean current velocity is associated with a small reduction of shear stress (6-17%); a small mean current velocity is associated with high shear reduction (27-35%). This relation is supported by in situ measured shear stress data which were introduced in chapter 2.

4.4.3 Discussions

1. In the near bottom region (z < 56 cm, for this case), this method will underestimate the shear stress because its Reynolds number is less than the critical Reynolds number. When this situation occurs, the results derived by Gross and Nowell (1985) can be used to modify the shear stress in the region where the Reynolds stress is lower than the critical Reynolds stress. This modification was expressed as:

\[ u_\ast = \bar{u}_\ast (z/z_c)^{1/3} \]  

(4-24)

where \( \bar{u}_\ast \) and \( u_\ast \) are the original and modified friction velocities respectively; \( z_c \) is the critical height, which can be determined from the critical Reynolds number (Fig. 2-19).
Fig. 4-9 The vertical distribution of friction velocity, $u_\ast$, calculated from the "inertial dissipation" method with neutral stratification (upper figure) and non-neutral stratification (lower figure) at station 3 during 12:00-19:00 5/15, 1992 (total 8 hours).
Table 4-3 Shear stress variation estimated using two methods

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<th>$u_{x2}$</th>
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<th>$\tau_2$</th>
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<th>mean</th>
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</tr>
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</table>

Notes: subscripts 1 and 2 refer to the neutral and stable cases, respectively.

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Fig. 4-10 Comparison of computational results of bottom shear stress between the "inertial dissipation" method with neutral stratification (original method) and non-neutral stratification (new method) at station 3 during 12:00-19:00 5/15, 1992 (total 8 hours).
2. Using the modified "inertial dissipation" method proposed in this study, the results indicate that mean current velocity is inversely related to shear reduction when considering the stratification induced by sediment resuspension. The in situ measured data in this study indicate that the resuspension coefficient is extremely low, which means that increasing bottom shear stress did not significantly increase sediment resuspension. However, the bottom shear stress will affect both shear production and buoyancy production. The results presented here imply that higher mean velocity causes higher shear production than buoyancy production, which leads to relatively low Richardson numbers. This inverse relation between mean current velocity and shear reduction is due to the low resuspension coefficient, i.e., the result is site specific. More studies are required to determine a general relation.

§ 4.5 Summary

In summary, the following points are concluded:

1. Linear distributions of both mixing length, \( l_m \), and eddy viscosity, \( A_e \), exist only in the bottom region \( (z<150 \text{ cm}) \) where the values increase with elevation. For smaller values of \( l_m \) and \( A_e \), the profiles are relatively uniform.

2. Eddy viscosity and mixing length vary with mean current velocity. This study indicates that 1) within the region of \(-10 \text{ cm/s} < u < -50 \text{ cm/s}\), eddy viscosity and mixing length decrease as the mean flow velocity increases; 2) when velocity exceeds \(-50 \text{ cm/s}\), the direct proportionality between them is attained. More studies are required to verify this relationship.
3. A new boundary layer thickness coefficient was obtained, $\delta=0.12-0.18u_*/f$, which is lower than all previously reported values. With this parameter, the calculation of the constant stress layer thickness has the best match with the measured results from this study.

4. The temperature and sediment concentration data in this study show the following features:

Because the bay water temperature is higher than the nearshore water temperature, the maximum flood velocity matches low peak temperature values; and maximum ebb velocity coincides with high peak temperature. The sediment concentration during the ebb period is thirty percent larger than that during the flood period. The vertical concentration distribution in the mid-depth region ($100 \text{ cm} < z < 200 \text{ cm}$) is relatively uniform.

5. The kinetic energy budget indicated that shear production has the same order of magnitude as the dissipation term. In the vertical, shear production can be divided into three regions: a low maximum-value region ($z < 0.3 \text{ m}$), a steeply changing mid-region ($0.3 \text{ m} < z < 0.75 \text{ m}$), and a upper uniform region ($z > 0.75 \text{ m}$). Compared with shear production, the vertical distribution of dissipation rate is relatively uniform.

6. The necessary condition for the existence of an "inertial subrange" is that the turbulent velocity wave number $K_i$ has to satisfy the condition, $K_i << K_i << K_d$. Based on the in situ measured data, the two boundary values are $K_i=9.74 \text{ cm}^{-1}$ and $K_i=0.01 \text{ cm}^{-1}$. Most inertial subrange bands varied within the range $K_i \in [0.05, 2.0] (1/\text{cm})$.
7. A modified "inertial-dissipation" method was developed from the consideration of the stratification induced by sediment resuspension. With this new method, the effect of sediment resuspension on bottom shear stress can be evaluated. The results indicated that when the Richardson number is within the region [0.02, 0.35], the bottom shear stress could be reduced by about 6-35% (mean value is 19.4%). This method has significant potential for both further research and practical use.
Chapter 5

Conclusions and recommendations for future studies
This dissertation is the first comprehensive study of sediment and flow dynamics within the bottom boundary layer of a high-tide-range inter-tidal mudflat channel environment. So both basic characteristics and unique phenomena will lay a foundation for future high-tide-range mudflat study. This study is focused on: 1) estimating bottom shear stress, 2) sediment transport processes over an inter-tidal mudflat, and 3) the effect of sediment resuspension on bottom shear stress. Seven levels of 3-D velocity fluctuation, two levels of sediment concentration and five levels of temperature data were collected in a selected distributary channel of the mudflat. Time series analysis techniques were adopted to support this study.

From the results of this study, the following conclusions can be drawn:

1. The dynamic features of flow in a bottom boundary layer are important to provide an insight into bottom shear stress. Based on seven levels of high frequency velocity data, vertical distributions of Reynolds stress (momentum flux), eddy viscosity and mixing length were found to be reasonably consistent with previous theories. The Reynolds stress profile is relatively uniform near bottom (z < 100 - 150 cm). The mixing length and eddy viscosity have linear distributions in the same region. Above that region all three parameters decrease with elevation. These features strongly signal the existence of a constant stress layer within the bottom boundary layer.

2. Comparison of three methods for estimating bottom shear stress indicates that

1) The "eddy correlation" technique gives the smallest value of shear stress. 2) Bottom shear stresses estimated from the "velocity shear" method are strongly related to mean...
current velocities, thus a drag coefficient can be used for establishing a relation between bottom shear stress and mean current velocity. 3) The "inertial dissipation" method gives satisfactory results when the turbulent fluctuations are measured within the constant stress layer (lower 100-150 cm) and their corresponding Reynolds numbers are greater than the critical Reynolds number ($Re_c=3000$). When the dissipation rate is calculated from a turbulent scale relationship, i.e., $\varepsilon = u'^2/l$, the estimated bottom shear stress shows good agreement with that from the "shear velocity" method. Using this turbulent scale relationship, the "inertial dissipation" method could be simplified. All three methods for estimating bottom shear stress are closely associated with the constant stress layer. In this study with non-neutral stratification, it is found that the thickness of the constant stress layer is less than indicated by previous studies. A new boundary layer thickness coefficient, $K=0.12-0.18$ is obtained. With this coefficient, the calculated constant stress layer thickness has the best match with measurements in this study. More analyses are required to verify this coefficient.

3. Sediment transport processes on a high-tide-range mudflat are strongly related not only to tidal current features, but also to erosion and deposition characteristics of the sediment. A high tide range creates steep channel slopes, relatively deep channels and finer grained sediment on the mudflat surface. Sediment transport processes on the mudflats can be divided into three time periods: 1) a consolidation/ exposure period, 2) a sediment carrying/ erosion period, and 3) a deposition period. The resultant features of the three periods are the formation of a
fine grained, erosion-resistant mudflat surface with coarser material in the channel bed. Mudflat erosion takes place mainly through upstream erosion in the channel and flank slumping at the channel bank. The meandering tributary channels result from channel erosion and mudflat deposition. Longer time-series data are required to establish the seasonal variation of sediment transport processes on the mudflat.

4. There exist three types of roughness length variations with bottom shear stress: proportional, constant, and reverse proportional. In this study, roughness length generally remained constant. A comparison with other data (Adams et al., 1996) suggests that in addition to bottom shear stress, compositional variation of channel bed sediment is relevant to the variation of roughness length. In this study, suspended sediment from the mudflat surface is unlikely to exchange with bed sediment under the high bottom shear stress conditions encountered. Because of the large difference in sediment size between the mudflat surface and channel bed, bed sediment maintains its relative steady size composition and bed form, which are the key factors producing constant roughness length.

5. A modified "inertial-dissipation" method incorporating stratification induced by suspended sediment is developed. By this method, the effect of sediment resuspension on bottom shear stress can be evaluated by the relation:

\[ \tau_o = \rho (k \varepsilon) ^ {2/3} \left[ \left( 1 + AR_x \right) \left( 1 + R_x \right) \right] ^ {-2/3} . \]

The results indicate that when the Richardson number ranges from 0.02 to 0.35, the bottom shear stress is correspondingly reduced 6-35% (mean value is 19.4%) because
of the stratification. This method has significant importance for both further research and practical use. It suggests that stratification induced by sediment resuspension should play an important role in estimating bottom shear stress, although more studies are needed to verify this method.

Recommendations for future study include: 1) The determination of energy dissipation rate is crucial for the use of the "inertial dissipation" method. This study indicated that the dissipation rate calculated from the "turbulent scale" method, $\varepsilon = u^3/l$, has satisfactory results for estimating the bottom shear stress. However, the determination of the length scale, $l$, in this method is, thus far, subjective. A theoretical derivation is required. 2) In the shallow inter-tidal channel, strong surface wind stress could influence the distribution of flow shear stress. The surface boundary layer may then overlap the bottom boundary layer. Thus, the determination of bottom shear stress will become more complicated. Owing to the lack of relevant data, this phenomenon was not included in the present study. 3) Also, there is a scarcity of in situ measured information on near-bottom ($z < 10$ cm) shear stress owing to sensor size and installation problems over a movable bed. The improvement of measurement instrumentation is always our expectation. 4) The three types of variations of roughness length, summarized here, have potential importance for the study of sediment transport and vertical velocity profiles. Further study of their mechanisms of formation is required. The direct measurement of bed form size is relatively easy in the inter-tidal channel. Thus, relationships of bed form dimension to sediment size, composition and flow velocity can be established.
References


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Appendix A  High/low pass—filtering of high/low frequency component in the spectrum

When the turbulent energy dissipation was studied, the energy spectrum was one of the most powerful research approaches. It was necessary to filter low frequency components from the turbulent data before the energy spectrum was estimated. The main procedures are summarized as follows:

1. Remove the mean of the original measured data, then do a Fast Fourier Transform (FFT) retaining the real part, Xr, and Imaginary part, Yi.

2. A box filter can now be created in the frequency domain. Its size will depend on the breadth of the low frequency band which will be removed. Here, A is supposed to be the box filter

3. In order to remove the low frequency part of the FFT, the Xr and Yi are multiplied by A. That is
   \[ X_{r2} = X_r \cdot A \]
   \[ Y_{i2} = Y_i \cdot A \]
   This sets the Fourier coefficients of the unwanted frequencies to zero and leaves the other coefficients unchanged.

4. \[ X_{r2}^2 + Y_{i2}^2 \] are then the periodogram estimates of the high-pass filtered signal.

5. Do an inverse Fast Fourier Transform

6. The data from step 5, are the high-passed time domain signals

For the low-pass analysis, I used the same principles as for the high-pass filter described above.
Appendix B Calculation of confidence interval

The confidence interval for an estimate is defined as the probability that a sample mean lies within the designated margin for error about the population mean. Here the confidence interval was introduced to indicate the representativeness of the spectral density. The procedures for calculation are summarized as follows:

1. Degrees of freedom

The degrees of freedom represent the number of independent variables in the estimate. In the spectral density function as estimated by the periodogram, there are only two independent variables, which are the squares of the real part and imaginary part of the finite Fourier transform, namely $X_r^2$ and $X_i^2$ in the $X = X_r + X_i i$. So its degrees of freedom is TWO. Sometimes the spectral density data will be averaged in order to make the spectral density function curve smoother. In this case, the degrees of freedom should be

$$k = 2m$$

where $m$ is the number of periodogram estimates averaged.

2. The determination of the parameter by $\chi^2$-tables (Chi-Square)

Suppose you would like to set 95% confidence intervals, which means there is a 5% chance of the true mean lying outside the interval. From the $\chi^2$-table, $\chi^2$ values can be found based on the degrees of freedom and distribution percentages. If we average eight periodogram estimates ($m=8$), $k=2m=2\times8=16$. Thus, $\chi^2$ values are 29.077 and 6.839 referring to the 2.5% and 97.5% distribution points with the mean value $E[\chi^2] = k = 16$.

Now suppose the true mean value is $M$, for any estimated value $S$, it has 95% probability of lying within the band

$$\frac{6.839}{16} \left( \frac{S}{M} \right) \leq \frac{29.077}{16}$$

which means $0.55 S \leq M \leq 2.34 S$. On logarithmic axes, no matter the value of $S$, this is always a region of the same length. So, finally, the 95% confidence interval region was set at $(0.55, 2.34)$.
### Appendix C Momentum flux \(<w' \cdot u'>\) at station 3

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Lun Xu was born on January 27, 1958, in Nanjing City, JiangSu province, China, the son of Zhenhua Xu and Xiang Yi. The hospital he was born has ever been used by the U.S. Embassy in China. He attended Chongming Laboratory Elementary School, Chongming Middle School in Shanghai, Gongnong High School in Hebei, graduating from the latter in January 1975. He became a mechanical worker at a military plant in August 1975. In 1978, he realized his first dream by entering North China Institute of Water Resources and Hydro-Power. He joined the faculty in the same university after he received his bachelor degree in hydraulic engineering, in July 1982. In September 1983, he realized his second dream by entering graduate program in the China Institute of Water Resources and Hydro-Power Research. He received his master of engineering degree in April 1986. Soon he became a registered engineer. In August 1991, Dr. Charles E. Adams Jr. supported him to realize his third dream by entering the doctoral program in oceanography and coastal sciences at Louisiana State University. Mr. Xu will receive his doctoral of philosophy degree in August 1996.
DOCTORAL EXAMINATION AND DISSERTATION REPORT

Candidate: Lun Xu

Major Field: Oceanography & Coastal Sciences

Title of Dissertation: Boundary Layer Turbulence and Sediment Transport in an Inter-tidal Mudflat Channel

Approved:

[Signatures]

Major Professor and Chairman

Dean of the Graduate School

EXAMINING COMMITTEE:

[Signatures]

Date of Examination:

10/5/95