

LARGE SCALE FLOW FEATURES IN SOME GRAVEL BED RIVERS

by

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Large Scale Flow Features in Some Gravel Bed Rivers

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## ABSTRACT

This thesis concerns itself with the nature and investigation of large scale flow features. A large portion of the thesis is devoted to review, criticism and organization of the relevant literature, most of which occurs in disciplines outside of fluvial geomorphology. Due to the type of data collected, 'large scale flow features' were defined operationally to include scales occurring between channel depth and a few multiples of channel width.

Lengthy speed records (varying from 20 minutes to one hour), sampled once per second, were collected in two local gravel bed rivers, the North Alouette and Squamish. All series were collected with an Ott dynamo current meter. Records were later subjected to probability, spectral and rescaled range analysis.

The analytic procedures chosen were not conducive to investigating the internal behavior of the series. As a result, emphasis was placed on using the literature review to generate hypotheses concerning gross series behavior under different channel conditions. The hypotheses were tested by comparing results obtained under differing channel size, discharge and local bed topography conditions.

Problems in adequately resolving the lowest and highest frequency parts of the series limit the confidence placed in the

conclusions. However, the results of analysis suggest that spectral behavior at large scales is fairly invariant under the range of channel sizes considered. Additionally, the behavior of some series measures suggest that large scale flow behavior is influenced by local channel bed topography.

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## I. INTRODUCTION

### Introduction

In any general approach to turbulent flows (which includes all open channel flows) the flow may be considered to be composed of two independent parts; a mean part and a fluctuating part (this is the Reynolds decomposition; see Tennekes and Lumley, 1972).

Traditional engineering approaches to the problems of fluvial processes (see Chow, 1959; Henderson, 1966) concentrate on problem solving based on 'mean' flow quantities. Hence in Henderson (1966) the bulk of the book is devoted to the energy and momentum equations of the mean flow and to general formulations of empirical resistance equations. This approach takes little or no account of the fluctuating flow component.

However, the solution of certain problems appear to be fairly intractable to this type of approach; notably initiation of sediment transport (particularly vibration of gravel particles; see Sutherland, 1967) and subsequent transport and bed deformation. In these general cases investigation of the interaction of fluctuating components and bed materials appears

to hold more promise of solution than mean flow approaches. Similarly, an argument may be made for the importance of fluctuating quantities at the largest deformation scales<sup>1</sup>; that is the semi-coherent bed or planform fluctuations of meandering channels. By simple analogy, it may be worthwhile to undertake investigation of fluctuating flow components at similar scales.

This idea of fluctuating energy at a variety of scales being important in fluvial processes is hardly original with this thesis. In fact, investigations of velocity fluctuations have been carried out at these disparate scales (those related to bed material sizes and those related to planform variation) for a number of years, though it is not unfair to say, investigations in natural channels have been few in number compared to those in laboratory flows. Similarly, investigations of small scale velocity fluctuations have dominated those at larger scales.

Fluctuations at small scales are generally referred to as 'turbulence' and there exists a well-defined experimental and theoretical body of literature. Some experimental papers describing the larger scale components in natural rivers (McLean and Smith, 1979) and atmospheric flows (eg., Kaimal et al, 1972) exist but no general body of theory is presently available.

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<sup>1</sup>Scale is used here in the general meteorological sense (Fiedler and Panofsky, 1970), a time scale is an average period and a spatial scale is an average wavelength

Several problems immediately appear when attempting to view the relationship between these fluctuating flow and bedform scales. First, the relationship between flow and bedform elements is complex at any scale and certainly is not anywhere near understood at present. In many cases, as in separation eddies, the relation between the bedform element and the fluctuating flow component appears to be mediated by variations in the mean flow structure related to the presence of the bed element. Additionally, alterations in the organization of bed elements are accomplished through increases in flow rate, which affect the structure of the mean and (possibly) the fluctuating components. These comments indicate that it will be extremely difficult to simply describe flow and bed element relations.

There are also problems that are unique to investigation of large scale components. In the case of relatively small scale fluctuating components generated by bed roughness elements, once a sensor is a few roughness 'wavelengths' above the bed the source appears to be continuous and the statistical properties of the fluctuating component vary only at right angles to the bed (Rouse, 1963).<sup>2</sup> Similar comments may not be true for the large scale structure and results may be influenced by the relative positions along the channel of the sensor and the serially isolated generating elements (this is to say, that

-----  
<sup>2</sup>This excludes downstream variation due to the larger superimposed bed elements. Equilibration of the small scale fluctuating components to these variations is fairly rapid.

flows are seldom deep compared to macroscale wavelengths). This will be discussed in more detail in Chapter VII.

Notwithstanding the above comments, one of the objectives of this thesis remains to attempt to investigate the effect of large scale roughness elements on the large scale velocity fluctuations. However, as pointed out previously, there is no body of 'theory' for these fluctuations in rivers, and to my knowledge, even no summary of the relevant literature. A necessary first step becomes comparison and evaluation of expressed ideas about the nature of these large scale fluctuating velocity components. A second step, then, is an attempt to integrate these ideas into the measurement and analytic context of this thesis. Since investigation for this thesis has proceeded by collecting data suitable for time series analysis these ideas must be evaluated in the context of serial analytic techniques such as spectral or rescaled range analysis. This evaluation forms the bulk of the material of Chapter I.

This can be put into a somewhat more specific form. Statements concerning these features often invoke 'quasi-periodic', 'periodic' or 'intermittent' flow processes. This chapter reviews what these terms mean for the types of measurements collected and their implications for the type of analysis carried out in later chapters.

Also, since the scales considered in this thesis and those of turbulence overlap, it is not only necessary to attempt to



define boundaries to the term 'large scale', but also to integrate ideas in the area of overlap. Consequently, concepts and analytic techniques from turbulence literature will be summarized. Much of this will be accomplished in Chapter II.

### Rationale and Context

This thesis grew out of interest in turbulence structure in rivers and particularly out of an interest in the relation between flow structure and bedforms. The whole of this topic (flow structure and bedforms) can not be studied in the context of a Masters thesis and this thesis represents an investigation of one part of one approach to this topic.

One possible approach to the study of flow structure and bedforms is to develop a hierarchial model of bedforms and associated flow processes. (notably Popov, 1964; Jackson, 1975, 1977; Nowell, 1975). This discussion follows Jackson (1977). For sand bed channels, the hierarchy usually relates the smallest forms, microforms (ripples and parting lineations) to inner layer structure in the flow (such as streak spacing; see Kim et al, 1971). The second level of the hierarchy, fluvial mesoforms, including dunes and large-scale lineations, are related to quasi-deterministic processes such as those associated with bursting (see Kim et al, 1971; this thesis, pp. 8). These fluvial mesoforms generally scale with channel depth or boundary layer thickness, or in a more dynamic sense, macroscale length

(Nowell, 1975). The mean period between bursts also scales with gross flow parameters (Rao et al, 1971; Laufer and Narayanan, 1971).

The largest bedforms, fluvial macroforms, include point bars, alternating bars, and the pool and riffle sequence. Jackson's comments on the relation of these features to macroscale flow features are somewhat confusing. He states that these features tend to be in response to long term hydrological conditions and scale with channel width (see Leopold Wolman and Miller, 1964). He further goes on to state that the flow features associated with these macroscale forms include helicoidal flow and to quote (pp. 16):

... "obscure long-period oscillations in flow velocity"...

The implications of this differentiation between associated and causative flow processes are interesting. The first implication is that this implies a viewpoint which accepts that flow processes measured over relatively long time periods (i.e. of the order of an hour) can be accepted as being produced by the largest scale roughness in the river, since the periods of change of these largest elements are extremely long compared to macroscale time scales. Secondly, this view suggests that flow processes that occur when these largest elements are being formed or altered (i.e., when bedform and flow are interacting) may be very different from those measured at lower channel

stages, indicating that most long term velocity records that have been collected to date may provide no information on the relation between macroscale form and flow processes. At present the effect of changing discharge is unknown. This will be discussed further in Chapter VII.

This three-stage classification where inner layer thickness, channel depth and width are the major scaling parameters may not be appropriate for all bedforms. As Jackson (1977) points out many bar types associated with gravel bed channels (braid, spool, linguoid, etc.) show both mesoformal and macroformal traits.

Perhaps more important, though is the fact that the same litany of processes and forms does not occur in gravel bed rivers. While lip service is usually paid to this fact, serious consideration is often not.

On the microscale, ripples and other small scale features do not appear and instead microscale roughness is expressed by the distribution of dominant gravel particles. Similarly, flow processes differ. The inner layer does not exist as an unbroken sheet and flow processes near the boundary are dominated by wake shedding around dominant three dimensional particles and vorticity amplification and wake interaction due to these same particles (Nowell, 1975).

Similarly differences appear to exist in the region of the mesoscale. While some mesoscale bed features such as dunes

appear to exist in gravel rivers (Galay, 1967) they are by no means as common nor do they appear to behave in a similar manner. The limited information available seems to suggest that dune wavelength does not increase with increasing depth indicating differing relations between flow features and forms than occur in sand bed channels.

Bursting<sup>3</sup> is generally associated only with smooth boundaries (Jackson, 1975; which is to say that roughness heights are less than inner layer thickness or of the order of a few millimetres for geophysical flows) and the relevance of bursting phenomena over rough boundaries and at large Reynold's numbers has not been established (Laufer, 1975).

Bullock, Cooper and Abernathy (1978) suggest that the viscous sublayer eruption model is inadequate to explain large eddies observed in pipe flow at high Reynold's numbers. Similarly, investigation of low frequency velocity records (Reynold's stress component) by McLean and Smith (1980) collected in the Columbia River showed no evidence for

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<sup>3</sup>Bursting is recognized as a quasi-deterministic sequence of events that reoccur randomly in space with some observable mean period, in boundary layers (Laufer, 1975). The process includes several events: namely, lifting of a low speed streak from the wall, oscillatory growth of the structure and chaotic fluctuation called breakup (Kim et al, 1971). Other studies (Corino and Brodkey, 1969; Nychas et al, 1973) describe different and non-compatible series of events. Bursting is a process of intermittent turbulent production and in geophysical flows is recognized by large intermittent excursions of the Reynolds' stress time series above the mean or indicated by large kurtosis of the normalized  $u'w'$  distribution.

intermittent turbulent production.

While the existence of bursting at large Reynold's numbers is unclear, the rough boundary case is equally so. Investigations of bursting over rough boundaries in laboratory flumes consist of one paper; Grass (1971). His results do not settle the issue since only two of the flow structures normally associated with bursting (ejection and inrush) were identified. Grass further suggests that (pp. 252)

"...different dominant modes of instability might prevail for different boundary roughness conditions."

No evidence is available for the existence of bursting phenomena at high Reynolds numbers over rough boundaries.

The foregoing suggests that bursting is not ubiquitous in geophysical flows (McLean and Smith, 1979) and that the conclusions of visual research are not extendable to geophysical flows over rough boundaries at large measurement scales and large  $y^+$  distances from the boundary (Ecklemann, 1974; intermittency is no longer observable past  $y^+ = u^*y/v > 200$ , where  $v/u^*$  is a length scale characteristic of the inner layer). Other intermittent processes may exist in the mesoscale range in gravel bed rivers and these will be discussed later.

Little is known (excluding bar types) of the different macroforms occurring in sand and gravel bed rivers. One form that may occur only in gravel bed rivers is the scour hole (Neill, 1969) though its existence may be more of a result of

interaction between planform river behavior and valley properties than due to the type of bed roughness. Observations of this feature are limited (Neill, 1969; this thesis, Chapter V).

With the above differences in mind it is perhaps easiest to think of the flow/bedform hierarchy in gravel channels as being composed of two parts; a microscale part including dominant roughness elements and associated flow processes and a macroscale part including large and intermediate scale bedforms and flow processes. \* The micro and macroscales can probably be separated somewhere near the lower limit of the turbulent range.

Jackson (1977) does not suggest any connection between these flow processes at different scales; however, it seems likely from the superposition of bedforms that these various identified scales exist in the flow at the same time. Nowell (1975) has suggested a hierarchy of interacting flow instabilities for gravel bed rivers. In this hierarchy, it is suggested that 'internal waves' in the flow may trigger instability at local scales (where this instability may persist through a complex interaction between the local and larger scales; see Mollo-Christensen, 1971) which produces flow regularities. Adjustment of the bed to these flow regularities may produce bedform periodicities which in turn sustain the

-----  
\* Macroscale will now be used to refer simply to scales larger than channel depth rather than Jackson's sense of the term.

'internal waves' and produce bed regularities over wider areas.

Very little is known of these non-linear interactions between widely different scales in the flow. In fact, investigation has generally proceeded by examining the elements of the bedform/flow process hierarchy separately. At this point relatively detailed knowledge is available about meso and microscale flow processes (for sand bed channels and laboratory flumes) and about relations between meso and microscale structure and some bed elements. However, little investigation has been undertaken of flow processes classifiable as macroscale. This thesis, then, intends to experimentally investigate these macroscale flow phenomena in gravel bed rivers.

The literature concerned directly with this topic is very small and that concerned with the investigation of the probabilistic or frequency structure of macroscale flow phenomena in gravel bed rivers is virtually non-existent. However, this general type of investigation is fairly common under other conditions.

A large body of work collected in atmospheric flows concerning 'turbulence' beyond the low frequency limit of the inertial subrange (see Monin and Yaqlom, 1975b for a summary; also Panofsky and Busch, 1969; and Kaimal et al, 1971 for experimental results) is useful for comparative purposes and general information on analytic and data collection techniques.

Similarly, a body of work exists concerning investigation of coherent flow structures and universal spectral scaling for laboratory flows under the general heading of fluid mechanics. Again, some of these results are useful for the purposes of comparison.

Additionally, some techniques used for the study of velocity records in this thesis (notably rescaled range analysis) are used extensively in unrelated fields such as hydrology. While the specific results are of no potential use, the general conclusions on the nature of the probability structure of time series are.

Finally, results of spectral investigations of time series collected in sand bed rivers are available (notably McLean and Smith, 1977; Grinval'd, 1972; Nordin et al, 1972; and Yokosi, 1967). These results are extremely useful for comparison with this thesis and will form part of the basis of the conclusions.

Other approaches to the determination of frequency structure of macroscale flow phenomena (notably, moving average techniques) will be considered and results will be reported here and contrasted with the results of spectral investigations.

As has been pointed out in preceding paragraphs, the information pertaining to this topic is scattered over a wide range of disciplines; in fact, all disciplines that consider the nature of fluid flow. Consequently, terminology is not standardized across these disciplines or even within them.



Additionally, general words in usage often have a very different meaning attached in the context of some discipline.

As a result, it is necessary in this thesis to decide on an appropriate term and description for what up until now has been called 'macroscale flow phenomena'. It is also necessary to provide operational definitions and boundaries for the topic under consideration; not only to limit the amount of material considered in this thesis, but also to provide a guide to field work and velocity sampling.

There are obviously numerous ways to develop terminology and definitions. This thesis proceeds through a subchapter called 'Some Ideas Related to Macroscale Flow Features'. In this subchapter a review is provided of some ideas about the nature of macroscale flow phenomena with some explanation and comparisons of terminology. This subchapter is not designed to be exhaustive of information about these flow phenomena. However, it is meant to introduce some ideas about the topic that will be useful in this introductory chapter.

One additional point needs to be mentioned before proceeding to the body of this chapter; this is the problem of using terminology and concepts that are unfamiliar to fluvial geomorphologists despite being common currency in other disciplines.

This problem has been dealt with in several ways. First, Chapter II provides a brief and simplistic review of some

classic turbulence parameters that are used in this thesis. Second, where it does not interrupt narrative flow terminology is explained when it is first encountered or an attempt is made to refer to an appropriate section of this thesis for more detail. Footnotes have been used extensively to provide tangential material.

This thesis does not attempt to fully explain or derive material borrowed from the literature. Hence, for instance, there is no theory or derivation of spectral analysis since this technique is so common that providing a full explanation is as redundant as explaining standard regression analysis.

### Objectives

This thesis is exploratory in nature. Consequently, a great deal of emphasis is placed on review, criticism and organization of relevant literature. However, the detailed objectives of the thesis are:

1. to measure macroscale flow structure in two gravel bed rivers and investigate properties of the velocity records collected using frequency and probabilistic statistical techniques. Cross-comparison of results from these two channels will form the bulk of the conclusions.
2. to compare results of measurements made in this thesis with other measurements collected under different boundary and flow conditions. Since other results are often presented in

different forms and detailed information on channel and flow characteristics are not available, these comparisons must of necessity be of the grossest kind. However, it is felt, that this type of gross comparison is acceptable for a first order look at this material.

3. to outline research methodologies useful for further investigation of these phenomena

#### Some Ideas Related to Macroscale Flow Phenomena

In an invited address to the First Midwestern Conference on Fluid Dynamics (May, 1950), Hugh L. Dryden commented briefly on the problem of the low frequency end of the velocity spectrum.

To quote:

There is, however, another limit when the wavelength  $2\pi/k$  approaches the dimensions of the air stream. We are then dealing with observations of phenomena which hardly meet the description of turbulence and are better described as pulsations of the entire flow. The modern theory shows that the spectral distribution of energy at small values of the wavenumber is not a statistical distribution but one determined by the initial geometry of the system originating the turbulence.

This quotation states two ideas that are of importance in investigating these phenomena. First, Dryden suggests that the geometry and dimensions of the channel containing the flow are important in limiting the type of features that may be observed in a given wavelength range and possibly in determining an upper limit to the size of various flow features, and since the distribution of energy is dependent on flow geometry, channel

dimensions may be used as suitable scaling parameters to produce a universal low frequency spectrum. Second, since these low frequency phenomena involve 'the entire flow' it may be possible to describe the low frequency behavior of the flow from a few velocity measurements (what this suggests is that transverse correlations of downstream velocity components tend to be large).

Similar ideas occur elsewhere. In Russian hydrological literature the idea of the occurrence of large scale 'pulsations' was introduced by Velikanov (1936; as summarized in Raudkivi, 1967). These features were experimentally investigated as velocity pulsations by Demente'ev (1963).<sup>5</sup> To quote Velikanov (from Demente'ev, 1963);

"...the entire stream can be conceived to be consisting of almost periodically repeating macroforms the dimensions of which are of the order of the transverse dimensions of the stream itself."

The term 'pulsation' (as translated by the AGU) seems to be generally reserved to refer to these large scale quasi-periodic macroforms. This quotation expands the definition put forward by Dryden to describe the features as longitudinal pulsations occupying the stream cross-section (evidence for the transverse extent is contained in a few correlations of synchronous

-----  
<sup>5</sup>As Hall and Johnston (1974) note, Demente'ev states (pp. 599) that "...the distribution of values of pulsation point velocities has a random character and follows Gauss' normal distribution of error"--a rather ambiguous experimental approach in light of later comments.

velocity records; see Demente'ev, 1963) and characterized by a dominant quasi-periodicity. It is not indicated whether this quasi-periodicity is related to channel dimensions.

Terms other than pulsation (macroturbulence, Korchokha, 1968; low frequency turbulence, Lyapin and Chebotarev, 1976) are used to refer to local phenomena possessing large size, continuity or regularity of appearance. Korchokha (1968) further suggests that in the case of the regular appearance of eddies shed from the lee of dune, that this regularity is controlled by the passage of pulsation phenomena. This suggests that the pulsation is a non-stationarity of the mean flow and controls the generation of other macroscale flow phenomena. It should be noted that the above distinction is not explicitly made in the literature surveyed, but rather is inferred from the ways the terms are used.

In America, the seminal paper regarding non-turbulence <sup>6</sup> is by Matthes (1947). In this paper, the term macroturbulence is used as an umbrella term for "...large scale turbulence phenomena", though large scale is left undefined. Returning to Dryden (1950) for a moment, it seems obvious that if the turbulence/non-turbulence distinction is based on the size of the phenomena relative to some limiting dimension of the

-----  
<sup>6</sup>a convenient definition of turbulence is "... turbulent fluid motion is an irregular condition of flow in which the various quantities show a random variation with time and space coordinates so that statistically distinct average values can be discerned." (Hinze, 1959, pp. 5)

channel, then for natural rivers these dimensions are width and depth. Generally, since  $w/d \gg 1$ , channel depth can be chosen to represent the scale that is 'large scale'.

Despite separating turbulence/non-turbulence on the basis of scale, the first level of Matthes' classification schemata is based on other observable properties of the features (i.e., types of motion, the second level involves the orientation of the principal axis). Matthes identifies six classes of macroturbulence phenomena which are

"caused by channel roughness but do not appear to be related to microturbulence."

Several of these classes will be discussed below.

The first class, Rhythmic and Cyclic Surges, includes velocity pulsations. Matthes comments that these features are irregular in magnitude and frequency. The second class, Rotary, Continuous, includes several features that qualify as separation eddies. As Matthes notes, these features are a product of a flow structure that is a result of flow around or over an obstruction. Interestingly enough, the shedding of features (called 'kolks') from the lee of ripples (Lane, 1944; described as an 'upward moving turbulent current') is included under Vortex action upward, intermittent rather than periodic. Matthes does not connect these features with separation eddies shed from the lee of the ripple as Korchokha (1968) does.

Boils or kolks have been recognized as intermittent

processes of some importance in velocity records by various authors (Nordin et al, 1971; associated with  $1/n$  spectrum and hence Hurst phenomena: Jackson, 1976; Grass, 1971; boils at surface connected with ejection phase near boundary: McLean and Smith, 1979).

Since these phenomena are generally associated with flows over two dimensional (and perhaps three)<sup>7</sup> roughness elements, this type of intermittency may be applicable to gravel bed rivers. While dunes are hardly ubiquitous in gravel rivers, structures or elements of an appropriate size or shape to produce this intermittency associated with separated flow may exist. Additionally, boils were observed to be common features during field work on the Squamish River. The following discussion concentrates on intermittency associated with dunes since only these types of investigations have been done.

Russian research on flow over sand dunes describes a structure which seems to be equivalent to the turbulent current originating on the face of ripples (Lane, 1944). Lyapin and Chebotarev (1976) suggest that in the troughs between individual bedforms there is a continuous alternation of instantaneous

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<sup>7</sup>Comparison of the spectral structure of flow upstream and downstream of a three-dimensional block roughness indicates that the block plays a role in redistributing turbulent energy. For frequencies above some neutral frequency, which appear to be related to the size of the block, dissipation becomes more rapid; for lower frequencies amplification occurs. Measurements in the wake of a single bar show a small increase of energy content at low frequencies (Nowell, 1975).

streamlines, from almost straight to convex upwards, and a corresponding fluctuation in the free surface. It is claimed that the succession of these currents produces a secondary hydraulic structure which seems to be a bottom roller shed from the lee of the dune. These low frequency structures were detected in a flume by Znamenskaya and Klaven (1967) with a period of 0.5 seconds.

Additional confirmation is available in two papers.

Znamenskaya (1963) has presented results describing in great detail the variation of macroturbulence structures with Froude number and the height to wavelength ratio of the bed feature. The typical form of macroturbulence associated with dunes was a large roller eddy contained in a zone in the lee of the feature where this zone often extended up to 50% of the depth of flow. These roller eddies were shed at regular intervals and moved up the pressure slope of the next downstream dune to the surface of the channel.

Korchokha (1968) summarized observations of these features in natural channels during experimental research on dune movement in the Polomet River. His visual observations indicated that boils observed on the surface of the channel were related to eddies shed from the trough of an upstream dune. While the size of the eddies does not remain constant during a single discharge and an average size appeared to increase with increasing discharge it was noted that the period of appearance



remained constant, independent of stream depth. Data from Korchokha (1968) indicated a period of approximately 6-10 seconds and he further noted that Grinval'd obtained a similar period on the Turumchuk River.

Since general evidence indicates that dune parameters (Church and Gilbert, 1975) are controlled by flow depth, an appropriate non-dimensional form of the period between successive rollers might be based on outer layer parameters as follows

$$TU / D \sim \text{constant}$$

where this is also the appropriate Reynolds number independent scaling for the mean period between bursts (Laufer and Narayanan, 1971; Rao et al, 1971; where the constant value is 5). The inverse of this scaling is plotted by Korchokha (pp. 552) against a parameter called 'eddy dimension' (which from the previous comments appears to be discharge related). The general form of his figure shows an approach of the non-dimensional period between rollers to a constant value of '5' with increasing 'eddy dimension'.

While the similarity between the scaling for the mean period between burst and between successive rollers is intriguing (also mentioned in McLean and Smith, 1979), numerous objections and qualifiers must be mentioned. First, the results from Korchokha are from one river and a very limited range of discharges. Additionally, the above-mentioned tendency to a

constant value is estimated from a few points on one graph.

Second, the relationship is obviously not universal over a wide range of discharges. The existence of the dunes, themselves, is limited to a relatively narrow range of flow conditions. Third, it is entirely unclear how this scaling is affected for flow over non-deformable elements of different spacings, shapes and dimensionality.

It is worthwhile at this point to attempt to summarize some of the identified features of the macroscale flow range in gravel bed rivers. These include a non-local, ubiquitous 'pulsation' component, a localized, (i.e., dependent on the position of roughness elements of an appropriate size and shape) relatively large scale component that may be intermittent, and additionally, a background of features varying in scale, that may be ubiquitous (e.g., boils in Matthes' sense) or extremely localized (e.g., bank eddies). It will likely be impossible to separate these components on the basis of energy contributions at specific scales.

Secondly, it is worth discussing the effect of channel dimension (where this may be a vertical, transverse or longitudinal scale) on energy distribution in the macroscale range. The simplest possible expectation is to assume that energy is skewed to lower frequencies as the "size" of the channel increases. This approach is consistent with the idea of channel "size" measures providing limitations on observed flow

scales (discussed in next section).

This is not to imply that the shape of the energy distribution is invariant over the range of scales from depth to a few multiples of width. Local effects may dominate the shape under different conditions.

### Definition of The Topic

In any experimental approach to investigation of these phenomena it is necessary to define practical upper and lower frequency limits to macroscale flow phenomena; that is to say, macroscale flow phenomena should be separated from turbulence on one end and hydrological events on the other.

As suggested earlier (Dryden, 1950) separation of macro and microscale components probably occurs at scales related to channel dimensions. While this is a reasonable basis of separation, it is more satisfying to separate the two phenomena on a dynamic basis. In line with the previous discussion of hierarchial instabilities (see pg. 4, this thesis), Nowell (1975) further states:

"The turbulence macroscale may also act as a lower limit for these coherent forces for the random vortex stretching which occurs at smaller scales would likely lead to the destruction of regular local bed features".

The macroscale<sup>6</sup> is that scale associated with the

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<sup>6</sup>The existence of this well-defined term is one reason why macroscale flow phenomena is an inappropriate term for the topic under consideration.

extraction of energy from the mean flow, and usually is some fraction of flow depth (see Chap. II). This extracted energy is cascaded to higher frequencies and the macroscale represents an approximate low bound for turbulence phenomena.

However, by accepting depth as a lower limit for these macroscale flow phenomena many local effects as identified by Matthes (1947) will be included in any records collected. These effects appear to vary greatly in size and intensity. Matthes suggests that the surface expression of some of these features (i.e., boils) can be very large (up to 300 feet in transverse and streamwise directions in the Mississippi River) and involve relatively large amounts of energy. These features appear to fall into that part of the scale continuum between depth and width values.

It is difficult to suggest a lower bound for these phenomena. Drainage basins for the two rivers are relatively large (Squamish Basin--4800 sq. km.; N. Alouette Basin--37 sq. km.; Inland Waters, 1976) and periods for flood events are probably measured in days and hours, respectively. Sampling at an interval related to depth (about once per second) for a period of a day or more not only provides more data than can be easily handled but may introduce complex non-stationarities in the mean velocity, which are unrelated to macroscale flow phenomena.

Consequently, it is easier to attempt to define a lower bound for these macroscale phenomena in terms of channel scales. Ishihara and Yokosi (1967) indicate that eddies larger than  $10w$  in the streamwise direction are unstable due to asymmetry, where this result is based on periods determined from the application of many term moving averages to velocity records from rivers and flumes. This result is not supported by spectral analysis of the velocity records. Many of the spectra produced by Ishihara and Yokosi (1967) and Yokosi (1967) cut off at frequencies higher than this limit based on  $10w$ , though spectra that extend down to this frequency show no indication of peaking.

Nowell and Church (1971) have related characteristic river dimensions and eddy types. Their approach involves identifying channel dimensions as limiting the size of certain flow features. The largest channel scale identified is a 'reach length' where this is some channel length within which no severe flow distortion occurs, implying, as they note, restriction to relatively straight sections between major bars or bends. Reach length distances will generally be of the order of a few channel widths ( $<10$ ). Nowell and Church (1971) further suggest that very low frequency 'pulsations' are resonance effects since the irregular channel geometry should not allow flow structures to exist over very large distances.

There is some evidence to indicate that relatively short records sample the bulk of the macroscale energy. Comparison of

lengthy speed records (approx. 3 hours) with shorter (30 minutes) longitudinal velocity records (McLean and Smith, 1979) indicate that the additional energy resolved by extending the record length was only 5-10% of that recorded by the shorter series. Again, this result indicates that most of the macroscale flow energy can be resolved by looking to frequencies corresponding to scales of a few multiples of channel width.

The second question involves choosing a convenient term to refer to the phenomena occurring between scales equivalent to the macroscale and those of the order of several channel widths.

There are numerous objections to most terms previously used in the literature and some of these will be outlined here. Many of the terms such as macroturbulence or macroscale flow phenomena suffer from similarity to other well defined terms to which they bear no relation. Other terms such as pulsation (Dement'ev, 1963) and eddy structure (Nowell and Church, 1971; Grass, 1971) suffer from implying certain ideas about the nature of the topic considered, whereas other terms such as velocity fluctuations (Savini and Bodhaine, 1971) and low frequency turbulence<sup>9</sup> (Lyapin and Chebotarev, 1976) suffer from a complete lack of specificity.

There appears to be no need to introduce another term into the literature and, for convenience, these phenomena will be

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<sup>9</sup>In atmospheric research low frequency turbulence refers to those scales larger than the inertial subrange and extending up to the macroscale (Monin and Yaglom, 1975a).

referred to as large scale flow features. This definition is operational and large scale refers to those scales occurring between channel depth and several multiples of channel width.

This term is appropriate only for the spectral investigation carried out in this thesis, and is not meant to coincide with any of the various hierarchies and classifications previous discussed.

### Approaches to Measurement

To this point no account has been taken of the context of measurement and in fact, results have been presented without reference to types of measurement systems.

While it seems that visual studies (i.e., Kim et al, 1971) and the type of investigation undertaken in this thesis are not directly comparable, it is important to note the limitations inherent in comparing results from similar measurement systems collected at different measurement scales and from different flow system scales.

The scale of the system affects particularly measurement of low frequency velocity components. As Nowell and Church (1971) note, high wavenumber (or frequency) components are independent of initial and boundary conditions and as such are modellable in flumes. Nowell (1975) has demonstrated that some of the properties of these high frequency components of flows over gravel beds are reproducible by flume experiments over discrete

roughness elements.

Nowell and Church further go on to state that large scale motion is not reproducible at different system scales since (pp. 26)

The low wavenumber range of the motion is neither completely de-coupled from the mean motion, nor is it independent of the boundary conditions, or of the size of the apparatus ... or channel.

However, as mentioned earlier channel dimensions may provide important scaling parameters for breaks or changes in the nature of the frequency distribution of energy. Second, the reproducibility of the low frequency end of the spectrum can be tested under different boundary conditions through some universal spectral scaling.

The scale of measurement affects the comparability of several parameters. Many low frequency studies (as this thesis) sample velocities in the order of .1 to 10 times per second while usual sampling rates in turbulent research are of the order of 200-400 times per second. The comparability of summary quantities such as turbulent intensities measured at these different scales is strongly dependent on whether both series sample to scales that include the major energy containing eddies. Other quantities such as cross-correlations are affected by the scale of measurement and results vary with different frequency cut-off points (Bullock et al, 1978), while accurate measurement of quantities such as Reynolds stresses depends on



sampling that resolves the major portion of the correlated energy (McLean and Smith, 1979).

Quantities such as moments of the velocity distribution and the shape of the velocity distribution are also expected to be dependent on measurement scale. Which is to say that velocity distributions may be different for different flow processes or ranges (i.e., dissipation, isotropic, energy containing) and the total superimposed velocity distribution may be affected by which particular ranges or processes are sampled.

Results are more comparable in the variance-frequency domain since the independence of variance estimates is controlled by the resolution bandwidth (estimates more than one resolution bandwidth apart are independent). Consequently, results from different record lengths and measurement scales may be comparable in some frequency ranges.

Additional to considering the measurement scale the actual method of measurement needs to be discussed. Traditional systems for measuring low frequency turbulence sample the longitudinal velocity (i.e. in the direction of the mean velocity) with a fixed current meter, though occasionally all three Cartesian components are sampled by rotating three orthogonal meters 45 degrees into the mean velocity (McLean and Smith, 1979). The measurement system in this thesis differed in that the signal was collected by a meter that was free to swing in the transverse and vertical dimensions (often referred to as speed

measurements since they are non-directional; see Chapter IV).

Comparability of results from this thesis with other low frequency time series depends on the relative energy contents of the three velocity components. As noted earlier, area preserving plots of the spectra of long measurement period free-swinging records and longitudinal velocity records matched fairly well at low frequencies (McLean and Smith, 1980; overlap was in the range of  $10^3$  to  $10^2$  seconds) which is encouraging.

Additionally inspection of spectra of the three velocity components indicates anisotropy at relatively small periods. The transverse component begins to fall below the longitudinal component by periods of two seconds while the vertical component starts to fall off at periods around 12 seconds. These results indicate that energy at large scales in the transverse and vertical components is constrained by the channel geometry and that results from this thesis are comparable with other low frequency spectra over most of the range of data. Problems of comparability appear to occur only in the highest frequency range.

### Spectral Analysis

This section is not an explanation of spectral techniques but rather attempts to deal with certain limitations to what can be discovered with spectral analysis and also with certain problems posed by the specific type of investigation chosen.

In general, spectral analysis operates on an ordered series of numbers (in this case, speeds) sampled at a constant time interval. This technique, through a Fourier transform, assigns the variance of the series to the frequency interval ranging from zero to the Nyquist frequency (which is related to the sampling interval), Since the transform fits sine and cosine terms to the original series, the variance assigned to a given frequency is the amplitude of a sine wave of that frequency.<sup>10</sup>

However, when a given process is referred to as low frequency (and again this is relative to the sampling interval and the length of the data series) it can refer to two different behaviors (Mandlebrot and Wallis, 1968). This is to say, the term low frequency can refer to either a long period oscillation of flow velocity or it can refer to events that occur relatively infrequently.

Bursting phenomena as pointed out previously are an example of events that occur intermittently with some statistical mean period. It is an arguable question whether these type of intermittent events and their associated large eddies can be detected by spectral analysis of fixed point velocity measurements. Part of the following discussion is a critique of

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<sup>10</sup>As Tennekes and Lumley (1972) point out, and as is mentioned above, each wave in a velocity field is associated with a single Fourier coefficient. An "eddy", on the other hand, has many Fourier coefficients associated with it (otherwise, it will not decay in the autocorrelation function). Consequently, the spectrum does not represent a decomposition into "eddies" but rather into time scales.

spectral analysis from the point of view of detecting intermittent events.

Base and Davies (1967) have demonstrated that a synthesized flow consisting of identical randomly spaced vortices can duplicate the traditional spectral measures of turbulence. Additionally, these structures may be regarded as three-dimensional vorticity lumps (final stages of bursting are characterized by multiple scales of motion; Kim et al, 1971) which has additional implications for spectral interpretation. The classic view of the spectrum is that the energy contribution at some wavenumber band  $k$  to  $k+dk$  belongs to an eddy corresponding to that scale. Davies and Yule (1975) point out that a region of vorticity (an approximation of a coherent structure) will distribute energy over a broad wavenumber band, where the largest scales are related to the size of the region and the smallest scales are related to the size of the steepest velocity gradient in the region. Spectral analysis will provide a reasonable estimate of the relative energy content of these real scales; however, the existence of the coherent structure will probably not be interpretable from the spectral measurement.

The foregoing has meant to point out the difficulty of detecting a flow structure based on intermittently occurring events. Additionally, it has been suggested that large scale flow structures may involve a single frequency pulsation or a

quasi-periodicity (Demente'ev, 1963; Dryden, 1950); that is to say, a long period oscillating behavior.

Again, some comment must be made on the ability of spectral analysis to detect these type of events. First, the question of evaluating periodicities. Detection of periodicities is generally based on inspection of the probability, autocorrelation and spectral functions (Bendat and Piersol, 1968). Techniques based on the probability density and autocorrelation functions are weak and can only detect very obvious periodicities so only detection through the spectral function will be described here (one of the basic functions of spectral analysis is to separate a signal from background distortion).

Although periodic components theoretically appear as a delta function of infinite variance at some frequency, spectral analysis spreads this power over the resolution bandwidth. Decreasing the resolution bandwidth should produce a peak whose spectral density increases in proportion to the decrease in resolution bandwidth and whose width is equal to the resolution bandwidth. Bendat and Piersol (1968) indicate that it is possible to recognize a sinusoidal component whose variance is equal to  $1/20$  of the variance of the random portion of the record. However, a sinusoid may be more difficult to recognize against a non-random background and can be indistinguishable from narrow band random noise at some resolution bandwidths.

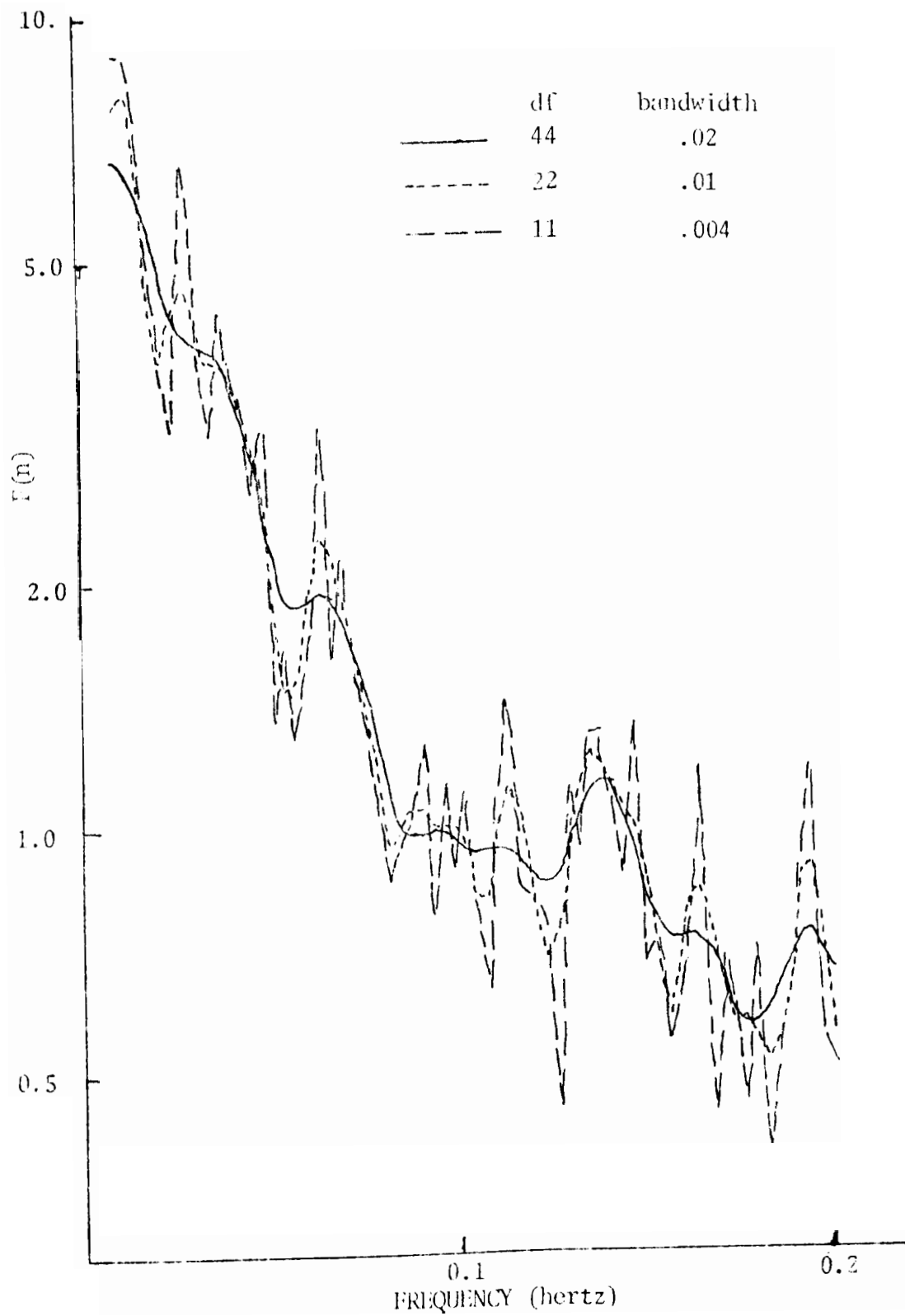


Figure 1.1: Spectral behavior under varying resolution bandwidths

Detection of these periodicities involves record lengths sufficient to keep the degrees of freedom<sup>11</sup> at a large enough value to minimize spurious peaks in the random portion of the data which occur when bandwidth decreases. Figure 1.1 illustrates the effect of decreasing bandwidth on the spectrum over the range .01 to .2 Hertz (data was collected in the Squamish River, at  $y/D=0.8$ ). Results from this five-fold change in resolution bandwidth (based on truncation point) indicate that local peaks observed at small bandwidths are the result of uncertainties in spectral estimation rather than the occurrence of periodicities in the record. This is not to imply that energy concentration in broad frequency bands is not useful for interpretation; merely that periodicities (or sinusoidal 'pulsations') are not detectable.

The question of the ability to detect quasi-periodicities

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<sup>11</sup>the comments are related to an autocorrelation approach to estimation of the spectrum (Jenkins and Watts, 1968). In this system, if  $N$  (record length) and  $df$  (degrees of freedom, representing the variance of the estimate, and usually set  $15 < df < 30$ ) and  $h$  (the digitizing interval) are fixed then

$$a = df/2Nh$$

where  $a$  is an estimate of the minimum bandwidth detectable. The bandwidth for a particular window used to filter the data can generally be written as a function of  $m$  (truncation) which is what is controllable in the spectral analysis package. Obviously, to view to data at smaller bandwidths requires decreasing the degrees of freedom and increasing the variance of the estimator. Similar results are also specifiable for the FFT approach.

is more difficult to assess.<sup>12</sup> As 'quasi-periodicity' is seldom considered in the analysis of velocity records, the brief discussion that follows is taken from techniques used in the analysis of nearly periodic, wind generated water waves, as discussed by Plate (1967).

This phenomenon can be investigated through either the autocorrelation or spectral functions. As Plate notes, autocorrelation functions of wave records show near periodicity as expressed by oscillation around the zero value with a well defined mean period and a tendency of the function to decrease over long time delays (Note--the autocorrelation of a periodic function does not decay over long time delays). The Fourier transform of this particular type of record produces a continuous spectrum with a broad peak around a frequency value equivalent to the average period obtained from the autocorrelation. Plate goes on to suggest that small phase shifts between adjacent waves tend to spread the variance over a broad range of frequencies. This is equivalent to suggesting varying time intervals between the passage of identical sinusoidal phenomena. Whether these near periodic (i.e.,

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<sup>12</sup>particularly since the distinction between quasi-periodicity and randomness depends on definition (though randomness may be defined as a flat spectrum). Generally, as Plate (1967, p. 314) points out, the criteria for distinguishing the two depend on

"the intended application of the record, and the knowledge of the processes that generate the record".



phase-shifted) phenomena are detectable against a random background of higher relative variance is unknown but it is expected that information obtainable from the autocorrelation will be obscured.

As a minor sidelight, it is worth discussing other techniques that are more capable of extracting the type of information that may be needed from velocity records. One possible technique for investigating changes in phase and variance is complex demodulation, which is in effect, a local application of spectral analysis. This technique looks at the local area of some point in the velocity record and produces phase and variance values for some pre-specified frequency,  $n$ . This technique produces  $N$  (where  $N$  is record length) estimates of the phase and variance, hence it is in no sense a 'summary' statistic. However, the technique is obviously useful for investigating phase-shifted waveforms as discussed previously.

The general usefulness of the technique though is severely dependent on a priori knowledge of some frequency value,  $n$ , of particular importance. Since this knowledge is not available the usefulness of the technique must await further developments in the understanding of large scale flow features.

A final consideration is necessary when using spectral techniques and this concerns the question of non-sinusoidal waveforms in the velocity record.

As Klemes (1974) points out it is difficult to detect anything except very obvious periodicities in natural records. Detection of non-stationarity in the mean is limited to either linear trends or simple sinusoidal components of a sufficient variance compared to more random parts of the record. Results from digital pattern recognition techniques applied to inner layer velocity records (Wallace; in Davies and Yule, 1975) recognized a basic pattern in the flow involving a slow deceleration in the longitudinal velocity component followed by a more rapid acceleration (corresponds to an ejection and sweep phase). This pattern was recognized as occurring in approximately 50% of the time history. While the above results are not directly relevant to investigation of large scale components, it is also possible to occasionally recognize (visually) non-sinusoidal repeating patterns in some of the data collected on the North Alouette and Squamish Rivers. It is not known whether these patterns have any significance since no theoretical basis is available to predict their form or relative importance.

The preceding discussion has emphasized the negative results of spectral techniques as these apply to empirically verifying the type of ideas about large scale flow features presented in earlier sections. More positive results of spectral analysis will be presented in Chapter II; particularly with reference to applying the successes of a spectral approach to

the turbulence problem to these larger scales.

### Rescaled Range Analysis

Rescaled range analysis is a probability measure that may be applied to the original velocity series. The technique is a serial one and results are independent of the marginal distribution of the data series (Mandelbrot and Wallis, 1969). The technique was originally developed by Hurst (1951) as a consequence of attempting to define the dam storage necessary to yield the average flow based on long term discharge records where the storage was obtained by computing cumulative sums of departures from the long term mean.

The technique, as used recently, examines the effect of increasing sample size (S) on the rescaled range (R/s, where R is the range and s is the sample standard deviation) and generally

$$R/s \sim S^{**H}$$

For a Gaussian series, H equals 0.5, however, investigation of numerous lengthy records by numerous authors (Hurst, 1953; Mandelbrot and Wallis, 1969; Nordin et al, 1972) indicates that H is greater than 0.5 for natural records. This phenomenon is generally referred to as the Hurst phenomenon.

Before discussion of the implications of the Hurst Phenomenon it is necessary to define an operational method for derivation of the rescaled range. One form used is:

$$R(t, S) = \max(0 < t' < S) \left[ \int_0^{t'} y(t) dt - t'/S \int_0^S y(t) dt \right] \\ - \min(0 < t' < S) \left[ \text{same quantity} \right]$$

and

$$s^2(t, S) = 1/S \int_0^S \left[ y(t) - 1/S \int_0^S y(t) dt \right]^2 dt$$

more general form is given by Mandelbrot and Wallis (1969) to include random starting points in the time series  $y(t)$  (i.e.,  $t=0$  does not mean  $y(t=0)$ )<sup>13</sup>

Unlike other techniques, the actual value of  $H$  has been of little importance in this type of investigation (other than  $H$  greater than 0.5) and discussion has revolved around several questions. Two of these are

4. which types of models that produce  $H > 0.5$  behavior are suitable as process description
5. whether the rescaled range behavior represents a transient; which is to suggest that present geophysical series are simply not long enough to approach  $H = 0.5$ .

These two points will be discussed in turn.

One model for generating time series data with appropriate  $H$  values is a class of processes referred to as fractional

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<sup>13</sup>Nordin et al's (1972) form is used here for ease of understanding. The program used determines the  $R/s$  values for a fixed set of sample sizes. Then for a given record length starting points are chosen so that a maximum number of sample sizes can have 10 independent (non-overlapping) estimates. The above equation is then applied for these sample sizes at the chosen starting points, and the cumulative sum of the series at each starting point is set to zero. The calculated  $R/s$  values are averaged which helps stabilize estimates at small sample sizes.

Brownian noise (Mandelbrot, 1965; Mandelbrot and Van Ness, 1968). Fractional Brownian noises arise from a process of infinite memory; which is to say that each increment is a weighted average of all past increments. The application of this model based on persistence<sup>1\*</sup> to velocity fluctuation implies that integral time scales do not exist (Nordin et al, 1972). Additionally, it is extremely difficult to integrate known flow behavior with this class of processes.

One additional feature of fractional Gaussian noises is that the spectral density function  $G(n)$  varies as (Mandelbrot and Van Ness, 1968);

$$G(n) \sim n^{1-2H}$$

and since most  $H$  values are near one, this implies a  $-1$  slope over most of the spectrum. For velocity data, the  $-1$  spectral slope is also associated with the energy containing eddy range (the production range). Nordin et al (1972) found this relationship to hold for much of the frequency range of turbulence scale measurements in the Mississippi and Missouri Rivers. However, the frequency range of their data was not extensive and the spectral slope appeared to drop at a larger rate than  $-1$  for the high frequency portions of their data.

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<sup>1\*</sup>'Persistence' in the sense of cyclic components (for instance, the sun spot cycle; see Mandelbrot and Wallis, 1969) manifests itself as a break in the R/S diagram which shifts the curve to the right at sample values near the period of the phenomenon. Why a quasi-periodic element does not provide infinite memory is unknown.

Similarly, their frequency data were not extended to larger scales that might lie outside the production range.

Fortunately, other approaches are equally successful for producing  $H > 0.5$  behavior. A more appropriate modelling scheme for the Hurst phenomenon in river flows is presented by the broken line process (Nordin et al, 1972; Mejia et al, 1974). This process results from linear interpolation between equally spaced independent random variables and random displacement of the starting point. The broken line process has several advantages in that it can preserve short term (i.e., autocorrelation structure and consequently integral scales) and long term ( $H$  values) properties of the series. These  $H$  values may be preserved by adding broken line processes and it is extremely intuitively appealing to see the Hurst phenomenon as a result of several superimposed intermittent processes of some finite persistence. At turbulent scales this is reasonable as dissipation, wake shedding and [bursting?] are demonstrably intermittent (Monin and Yaglom, 1975b).

The question of the appropriateness of this type of model becomes more complicated for large scale investigations, which effectively do not include these known processes. In the case where process knowledge is not available, several approaches may be taken. Following Mandelbrot and Wallis (1968) it could be assumed that the low frequency case is a simple one. As they state for precipitation fluctuations:

"Whereas precipitation fluctuations of wavelength near one day or one year may participate in several physical mechanisms, fluctuations of precipitation near one lifetime are likely to participate in one mechanism only. Thus, the latter are likely to be simpler than the former."

Similar comments could be applied to the low frequency part of the continuous velocity spectrum.

However, it is probably more worthwhile to assume models that are compatible with previously discussed ideas about large scale flow features. In this regard, it is worth looking at whether these ideas are compatible with other modelling schemes for series exhibiting the Hurst phenomenon.

To this point suggested mechanisms at large scales have involved an intermittent and/or a non-stationary mean component. As has been suggested several times these two components may be intimately connected. It is worthwhile to attempt to delve into the differences between these two mechanisms and attempt to investigate what differences exist operationally and conceptually between these two terms and, indeed, whether any differences exist.

Intermittency is described by Monin and Yaglom (1975b) in discussing the character of turbulent dissipation.<sup>15</sup> They state (pg 600);

"The word is meant to denote the tendency of small-scale

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<sup>15</sup>It is worth noting that Monin and Yaglom additionally suggest that as Reynolds number increases the range of scales involved in intermittency increases (up to the energy range) and the degree of intermittency increases.

turbulence to concentrate into individual "bunches" surrounded by extensive flow regions in which there are only much smoother large scale disturbances (or perhaps no disturbances at all)."

On the simplest level, non-stationarity refers to the tendency of the probability structure (i.e. statistical properties) of the series to change over time (Klemes, 1974). Nonstationarity is usually attributed to either transient operating conditions or long range changes in operating conditions, where these conditions refer to the 'processes' that generate the series. The three basic representations of nonstationary data are (Bendat and Piersol, 1968);

1. time varying mean quantities.
2. time varying mean square values (this is the second moment around the origin and is equal to the variance plus the square of the mean value)
3. some combination of the above.

where the time variation may be of numerous types (such as sinusoidal). The mean value and the mean square value are the two properties usually tested to determine (weak) stationarity (see Chapter IV).

Intermittency as expressed by Monin and Yaqlom appears to be an example of time varying mean square value, though additional complication is added by the variance in the two distinct record types being associated with extremely different



scales.<sup>16</sup> It is reasonable to assume that a similar type of intermittency is associated with shedding of a roller behind obstacles.

Since the other flow component is also a form of nonstationarity, in this case a time varying mean value, the data can be seen as an expression of nonstationarity of type (3).

The question then turns to detection of nonstationarity in velocity records. As Mollo-Christensen (1971) demonstrated the fine structure of the spanwise correlation of the longitudinal velocity component is extremely sensitive to averaging time. Similarly, evidence of nonstationarity at small scales is easily removed by averaging or summary techniques. As is mentioned in Chapter IV all data series are found to be (weakly) stationary, however, this result is based on the behavior of mean and variance values averaged over 150 second lengths of record (this value is chosen deliberately to be greater than the limits of autocorrelation). Since the interval related to the mean square nonstationarity is of the order of a few seconds for the rivers investigated in this thesis, these nonstationarities will be undetectable. This is a fundamental problem of all summary statistics and a more suitable approach to local nonstationarity may be embodied by complex demodulation (Bloomfield, 1976; see -----

<sup>16</sup> The mean interval between phenomena is not descriptive of the scales and amplitudes operative during the passage of the phenomenon.

previous section).

It is worthwhile to return briefly to modelling schemes of Hurst behavior. There has been no attempt to my knowledge to model the Hurst phenomenon with series exhibiting mean square non-stationarity.<sup>17</sup> Klemes (1974) has had numerical success in reproducing Hurst behavior with a model based on mean nonstationarity. Essentially, the model depended on the serial organization of epochs. Each epoch (where these are subsets of the data series of varying lengths) was a Gaussian random series with unit variance and different mean values. These epochs were combined for varying lengths and mean value distributions and  $H > 0.5$  behavior was found for certain combinations.

While this is just another modelling scheme its importance lies in demonstrating the multiplicity of processes that can be used to model the behavior of statistical parameter such as the rescaled range behavior, and hence the basic inability of many statistical techniques to distinguish or 'evaluate' process models.

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<sup>17</sup>Serial statistical techniques such as spectral analysis and autocorrelation analysis are inappropriate or at least extremely difficult to interpret for nonstationary data. As Klemes (1974) points out the differences between his nonstationary model and a stationary fBn model is that some of the variation in the series is attributed to the 'mean' rather than the fluctuations (i.e. such features as pseudocyclicity, quasi-periodicity, 'swings' or the 'Joseph effect'). In terms of series measures this distinction makes no difference and these distinctions are not inherent in the processes generating the series but in the arbitrary model.

The other major area of concern about rescaled range analysis (transient behavior) involves discussion of whether natural time series are simply not long enough to show  $H = 0.5$  behavior. While various types of models show asymptotic behavior of  $H = 0.5$ , few natural series have been found which do, though various data transformations show slope breaks for large  $S$  less than sample size (Hansen and Rosbjerg, 1974).

These internal slope breaks may be of use in determining important data scales. In this light, recent observations by Van Atta and Helland (1977) are interesting. Their data was collected for grid generated turbulence, and the rescaled range diagram showed an initial transient ( $H > 0.5$ ) which they related to the correlation structure of the turbulence, then an  $H = 0.5$  range, which extended for one decade, and  $H > 0.5$  behavior for large  $T$  values. The large  $T$  behavior was related to a quasi-periodicity in the data due to the circulation time of the flume.

This behavior suggests two independent and separate (i.e. in scale) levels of energy insertion. Insertion occurs at scales related to the grid spacing and scales related to the travel distance around the flume circuit. Since the grid determines the upper size limit of the 'turbulence' features, behavior between these two scales is random. It is also interesting to note that when the grid was removed the transient  $H = 0.5$  range disappeared from the rescaled range diagram. In terms of the

previous argument this implies that boundary roughness eddy generation is not controlled at the upper limit as grid turbulence is.

A suitable non-dimensional form of the rescaled range diagram, suggested by Van Atta and Helland is;

$$RU/sD \sim US/D$$

where  $D/U$  is a characteristic time scale. This scaling is effective in collapsing rescaled range data from several sources (see their Figure 2). The characteristic time scales used involved, mesh length and mean velocity for the grid turbulence data, channel depth and mean velocity for Nordin et al's (1972) turbulence data and an estimate of 10 years for the Nileometer data. This particular choice of scalings gives an  $H = 0.5$  range for  $10^2 < US/D < 10^3$ ; however, only the grid turbulence data series is sufficiently long to reach these nondimensional time values (note--this random range occurs between 100 and 1000 times the integral time scale of the grid turbulence).

Van Atta and Helland go on to suggest that by analogy with the grid turbulence behavior there should exist an  $H = 0.5$  range somewhere between a multiple of the scales of energy insertion due to bottom roughness (the characteristic time of the mean shear) and some large scale energy insertion scale. For rivers, an appropriate large scale flow through time may be based on the mean velocity and a 'meander length', though perhaps better longitudinal scales may be defined by bed spectral analysis.

It should be barely possible to test for this change to a  $H = 0.5$  slope with the data collected in this study. Data from the North Alouette and Squamish Rivers extend to US/D values of approximately 500. It is possible to observe an approach to an  $H = 0.5$  slope in rescaled range plots from Nowell (1975; his figure 44a) at the farthest end of the data (above UT/D = 500 approximately). However, as Mandelbrot and Wallis (1969) note it is foolhardy to place much reliance on results at either extremely small or large sample values.

#### Summary and New Directions

While this chapter has been lengthy for an introductory chapter, this length has been necessary. As was pointed out in the introduction there were two broad areas of concern that needed discussion. These were;

1. A comparative approach to general ideas about processes occurring at low frequencies, independent of the experimental context.
2. An attempt to integrate these ideas into the measurement and analytic context of this thesis--i.e., which ideas are testable.

Process oriented ideas have been defined operationally as nonstationarities of the mean or mean square values of the data series. It has been demonstrated that the serial statistical techniques commonly used in the analysis of velocity records are

insufficient to conclusively evaluate these models of process behavior. This may possibly be a result of a poor specification of the implications of these process models. In other words, the general process models as expressed in Chapter I, are not sufficiently well-developed to be described as 'testable theory' for the analytic techniques used in this thesis.

This obviously introduces problems into the interpretation of these statistical results. As Plate (1967, pp. 311) states with regard to the spectral analysis of dunes;

"As in the case of turbulence, an understanding of the spectrum shape will come only from a physical understanding of the interaction processes between frequency components of the basically uncorrelated superimposed sinusoids that make up a time function."

While this is discouraging in terms of solving the 'low frequency problem' in any simple way, advances can be made by a re-orientation of the overall approach.

Consequently the next steps involve

1. overview of classic turbulent spectral interpretation--that is, the relations governing neighbouring scales (Chapter II).
2. development of a descriptive and comparative approach to the available low frequency spectra. This is to say, discussion of spectral variation in terms of channel scales and general flow conditions (Chapter VI and VII).

## II. A BRIEF REVIEW OF TURBULENCE PARAMETERS

### Introduction

As has been mentioned previously there is some knowledge of flow processes at small scales; that is to say, some understanding of spectral shape. Consequently, it is necessary to discuss such concepts as local isotropy since this concept in particular has been used in several papers concerned with low frequency spectra in rivers. Additionally it is important to evaluate whether other 'turbulent scale' concepts can be extended to the range of scales considered in this paper.

Also this Chapter is intended to provide a brief overview and explanation of concepts and terms that have been used in this thesis. For this, the author has used Tennekes and Lumley (1972) and Nowell (1975) extensively.

### The Boundary Layer

Open channel flows are a subclass of boundary layer flows. One of the important characteristics of flow over a solid boundary is the presence of two length scales in the flow. The presence of the wall enforces a no-slip condition and gives rise to a viscous scale  $\nu/u^*$  (where  $\nu$  is viscosity: this is the basis

of the nondimensional distance,  $y^+$ , which also appears in the logarithmic velocity profile for smooth boundaries) over smooth boundaries, and additionally a characteristic roughness,  $k$ , over rough boundaries. The other length scale is  $\delta$ , the boundary layer thickness. There is a definitional problem with this scale in open channels since the boundary layer may intersect the surface before the thickness is fully developed; which suggests the boundary layer may be suppressed (Nowell and Church, 1979). However, channel depth is generally used in place of thickness.

Due to the presence of these two scales the boundary layer is most conveniently viewed as being composed of two parts. One part is an inner or 'wall' layer dominated by  $k$  or  $v/u^*$  and the other an outer layer where channel depth is the important scale. Development of the logarithmic outer layer velocity profile and the law of the 'wall' from this scaling is well described by Tennekes and Lumley (1972).

Additionally, for large roughness elements a wake zone may form at the top of the roughness elements, characterized by a uniform layer of velocity and turbulence intensity (Nowell, 1975), where the thickness of the layer corresponds to obstacle dimensions. Hence, in the wake region the size and spacing of roughness elements become important length scales determining the size of eddies. In general, the spacing of roughness elements determines the flow-boundary interaction and resistance to flow (Morris, 1955). For isolated roughness elements the



nature of the intervening boundary (smooth or rough) becomes important. Wake interaction occurs when spacing of elements allows wakes to interact and increase the inviscid resistance, and skimming occurs for densely packed elements that 'lift' the flow above them.

### Turbulent Scales in Open Channel Flows

Several scales are important in defining turbulence. As mentioned previously, the macroscale is the largest turbulent scale and is that scale that interacts with the mean shear to extract energy. The macroscale is usually some fraction of the channel depth and tends to increase with increasing distance from the boundary.

Operationally, this scale is given as the area under the autocorrelation function (also called the Eulerian Integral scale). The macroscale may also be defined from the spectrum (i.e., Raichlen, 1967). Another scale may be defined from the autocorrelation function (Rouse, 1963). This scale is based on the limit of positive correlation and represents the size of the largest "eddy". The maximum time lag of positive correlation represents this largest "eddy" since dependence of velocity fluctuations should only exist within an eddy, hence the limit of correlation represents the passage time of this feature.

The additional scales used in turbulence research are associated with dissipation (the smallest scales) and are the

Taylor microscale and the Kolmogorov microscale. Dissipation in turbulent flows is associated with the effect of viscosity. The Taylor microscale is associated with the peak of the dissipation spectrum (the derivative spectrum) and the Kolmogorov microscale is approximately a decade smaller and represents the beginning of viscous effects on turbulent flows. Both of these scales are much smaller than the velocity sampling intervals used in this thesis.

### The Frozen Turbulence Hypothesis

The classic approach to turbulence is to measure over time at a fixed point (Eulerian approach). Then, the turbulence structure is regarded as being advected past the sensor by the mean flow and appears as a continuous, ergodic random variable.

This type of measurement produces statistical measures in the time domain whereas, in general, turbulence is scaled in terms of lengths. Transformation from temporal to spatial scales is possible if the rate of flow is large relative to a characteristic turbulence velocity, i.e., if

$$\sqrt{u'^2} / U < 1$$

(this does not appear to be true for the North Alouette data)

then

$$u'(t) \sim u'(x/U) \text{ or } d/dx = -(1/U) (d/dt)$$

The frozen turbulence hypothesis assumes that the turbulence structure does not evolve relative to a reference frame

travelling with the mean flow. However, this assumption is only valid under limited frequency conditions. Lumley(1965) discusses several reasons why this approximation breaks down at low frequencies, among them

3. eddies in the production range or eddies of the same size as the macroscale are being stretched as they pass the instruments.
4. smaller eddies are imbedded in larger eddies producing a fluctuating convection velocity.

These effects are dependent on the relative turbulent intensity and transformation to a wavenumber space for scales much above the inertial subrange becomes very inexact. Hence, data in this paper is expressed in either frequency or non-dimensional frequency form. All scales ( $x$ ) calculated from

$$x = U/n$$

are necessarily inexact.

### Spectrum/Spectral Presentation

The fluctuating velocity may be decomposed via a fourier transform to yield the turbulent kinetic energy per unit mass associated with the various frequency components.

There are several ways of scaling these spectral amplitudes derived from the fourier analysis to produce a spectrum and terminology varies widely. Jenkins (1961) will be followed in this thesis. The power spectrum function (or energy spectrum;

Harris, 1974) is

$$\int_0^{n(c)} f(n) dn = \text{variance}$$

where  $n(c)$  is the maximum or Nyquist frequency based on the sampling interval ( $h$ ) and  $n(c) = 1/2h$ . In this thesis the spectral density function is used where this is

$$\int_0^{n(c)} P(n) dn = 1$$

where obviously  $P(n) = f(n)/\text{variance}$ .<sup>1</sup> This format has an advantage for comparing variance accumulation with frequency and also spectral shape for several spectra.

Other similar terms are also in relatively common usage. The energy density spectrum (Harris, 1974) is the squared amplitudes from the fourier transform divided by the averaging interval applied to the raw spectrum estimates (in the FFT procedure this averaging interval is the resolution bandwidth). This type of energy density spectrum (which obviously differs dramatically from Jenkins, 1963) was also used by Mclean and Smith (1979) for the spectrum of their transformation of time data into wavenumber space. Another term used is the power spectral density function (Bendat and Piersol, 1968), where this function sums to the mean square value. This converges to the spectral density definition used in this paper when the mean

<sup>1</sup>Operationally this does not always appear to be true. The spectral density function calculated in this thesis integrates to one-half of the variance, while for many FFT programs (Harris, 1974; Mclean and Smith, 1980; also TROLL econometrics package, SFU) the squared amplitudes sum to twice the variance of the series. This can create problems in comparing spectra from different sources.

value of the series is zero. (means are usually removed before spectral analysis).

Similarly, there are a variety of graphic styles for presenting spectra. The discussion follows Harris (1974). Three of the forms described involve manipulation of the graph axes; while the fourth involves manipulation of the spectral density values.

A seldom used graphic form is spectral density versus frequency. The only useful feature of this presentation is that it is area-preserving, however, information tends to be squeezed into a small part of the graph. Similarly the logarithm of spectral density as a function of frequency is seldom used. This form tends to highlight spectrum irregularities; hence, it was used for Figure 1.1.

The most common graphic device is the logarithm of spectral density against the logarithm of frequency. The form is used principally because it is appropriate for looking for power law relationships between density and frequency. The form is also useful for a detailed view of density variation at low frequencies. The disadvantage of this form is that a unit area under one part of the graph is not equivalent to a unit area in other parts (i.e. it is not area-preserving); the form concentrates area in the lowest bandwidths and leads to visual overestimation of the variance contribution of these bandwidths.

One other plot that is fairly common is area-preserving; frequency times density as a function of the logarithm of frequency (McLean and Smith, 1980). A similar form that is not area-preserving is used in this thesis, that is the logarithm of frequency times density against the logarithm of (non-dimensional) frequency (see Chapter VII). Both of these plots emphasize high frequency components of the spectrum.

### Spectral Slopes and Processes

The primary role of turbulence is as an energy dissipator. To this point, two scales have been identified, a macroscale where energy is inserted and a microscale where viscous dissipation occurs. Energy is passed between these two scales through a cascade process initiated by vortex stretching. One of the implications of this cascade process is that "eddies" of a certain scale receive the bulk of their energy from slightly larger scales and transfer energy to slightly smaller scales. Consequently, widely different scales are independent in terms of energy interaction.

Classic theories (Kolmogorov, 1941) of the turbulent energy spectrum indicate that for sufficiently large Reynolds numbers (usually  $>10^5$  for rivers, indicating sufficient space between production and dissipation scales to include a range unconnected with either) there should exist a region in the spectrum called

the inertial subrange<sup>2</sup> where eddies are isotropic since they are independent of the mean shear and still unaffected by viscosity. These scales should only be affected by the rate of energy dissipation( $\epsilon$ ). In this region the universal spectral form is

$$E(k) = k' (\epsilon)^{2/3} (k)^{-5/3}$$

where  $E(k)$  is the three-dimensional energy spectrum. Spectral measurements indicate that the  $-5/3$  zone becomes more extensive with increasing distance from the boundary (Laufer, 1950; Nowell and Church, 1979), up to midchannel. With increasing roughness density or increasing  $u^*$  this range disappears. The existence of an inertial subrange is frequency limited and it is unlikely that sufficiently large frequencies are considered in this paper to observe this behavior.

In the energy containing region of the spectrum, which exists for eddies the size of the macroscale ( $L$ ), the energy density is distributed as (Hinze, 1959):

$$E(n) \sim (n)^{-1}$$

Measurements of the one-dimensional turbulent energy spectrum confirm the existence of this range (Klebanoff, 1955; Laufer, 1954; Nowell and Church, 1979). Opposite to the inertial subrange case the  $-1$  zone is most extensive near the wall.

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<sup>2</sup>Equally, there may exist an inertial subrange in the spatial structure of the boundary layer. This would be a region where the inner layer scale is too small to control the flow and the outer layer scale is too large. In this region  $y$  is the only relevant length scale (Tennekes and Lumley, 1972). See also Chapter VII.

Measurements in a gravel channel (Nowell, 1975) show an extensive -1 range at low wavenumbers, for measurements ranging from near the bed to approximately midchannel.

### The Continuous Frequency Distribution/Gaps and Two-Dimensionality

To this point spectral behavior has been discussed for what is essentially the microscale range ( $L < D$ ). Spectral behavior has been extended to much larger scales in atmospheric work (for a recent summary; see Monin and Yaqlom, 1975b) and these results are worth investigating despite the "unconfined" nature of atmospheric boundary layers.

First, since this research uses the terms 'macroscale', 'microscale' and 'mesoscale' these must be defined in an atmospheric context. These horizontal temporal and spatial scales then are (Fiedler and Panofsky, 1970);

1. microscale;  $< 1$  hour,  $< 20$  km; includes bouyant and mechanical production ranges and also convective systems.
2. mesoscale; 1 hour - 48 hours, 20 km to 500 km; includes diurnal breeze systems and squall lines.
3. macroscale;  $>> 48$  hours,  $>> 500$  km; weather map size features.



General results of patching<sup>3</sup> time spectra (see Van der Hoven, 1957; Panofsky, 1969; or for a summary, Monin and Yaqlom, 1975b) to cover the macroscale to microscale temporal range indicate a major energy gap in the range of the mesoscale where this gap is centered at around 1 cycle/hour (Van der Hoven, 1957). This gap is a general feature of long period longitudinal spectra, though under some atmospheric and boundary conditions this gap may disappear (Panofsky, 1969).

The explanation of this spectral gap is most simply seen in terms of preferred wavelengths of instability. Hence, the macroscale peak is associated with baroclinic and barotropic instabilities on synoptic scales whereas the microscale is associated with shear instabilities on the scale of a few kilometres or less. The spectral gap implies that these two different wavelength groups do not interact directly. However, energy exchange may proceed through cascading of macroscale energy to the mean flow (Fiedler and Panofsky, 1969), which then may be extracted at microscale wavelengths through vertical shear.

Additionally, macroscale and microscale motions are separable on the basis of their dimensionality. The larger scales are quasi-two-dimensional, since vertical development is small relative to horizontal scales. Spectra of vertical motions

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<sup>3</sup>i.e. spectral results from two or three records of different lengths and sampling intervals collected at one 'place' are matched (see Van der Hoven, 1957).

generally have large energy content only near the microscale.

The distinction between two-dimensional macroscale and three-dimensional microscale eddies is important for spectral dynamics in the atmosphere. Two-dimensional energy exchange is a reverse-cascade process; i.e., exchange occurs from smaller to larger scales (Kraichnan, 1967). Similar to three-dimensional turbulence there is a two-dimensional inertial range characterized by a  $-5/3$  spectral slope.

Some of the implications of these results are interesting for large scale spectral investigations in rivers. As Gage (1978) suggests microscale energy sources may be associated with an upscale two-dimensional cascade and a downscale three-dimensional cascade. The extension of the  $-5/3$  range of some low frequency spectra (Yokosi, 1967; see Chapter III) to much larger wavelengths than expected might be explained by this type of behavior. However, it is not known at what scales low frequency velocity fluctuations in rivers may be considered two-dimensional. Similarly, much of the data supporting these two-dimensional spectral ranges comes from upper air observations and extremely different results may occur near the ground due to local and mesoscale topography variations.

More importantly, for developing approaches to large scale features in open channel flows, the above material suggests that the macro/meso/micro scale distinctions made so far in this thesis may be rather meaningless in terms of separating distinct

process ranges.

Consequently, definitive separation between any independent macro/micro processes should be based on spectral gaps. Whether gaps will occur only between processes of energy insertion at hydrological and local roughness scales or whether a gap may occur between local roughness and large scale roughness 'processes' (if they exist) is unknown. Large scale spectral results from sand bed rivers suggest that a gap only occurs between hydrological and local roughness scales (see McLean and Smith, 1979). However, the one set of available results from a gravel bed river (Nowell, 1975) shows spectra peaking at relatively small wavelengths, which may imply a gap between local and large scale roughness 'processes' (no other example of peaking in turbulent scale  $\log f(n)$  vs.  $\log n$  spectra is known to the author; i.e., see Raichlen, 1967; Monin and Yaqlom, 1975b; Laufer, 1950; etc, whether this is due to the presence of the gravel bed or to relatively longer series lengths is unknown).

The search for 'gaps' between independent processes is not limited to spectral analysis techniques. A similar argument has been made concerning the meaning of "labels" (i.e., macrometeorology and climatology) dividing the study of the spectrum of climatic fluctuations by Mandelbrot and Wallis (1969). They suggest that a distinction occurs between independent processes if the rescaled range diagram shows an

intermediate region of slope  $H=0.5$ , or exhibits two distinct regions with an intermediate region of different character than both. This approach has already been discussed for Van Atta and Helland's (1978) results (see Chapter I, 'Rescaled Range Behavior').

### III. LITERATURE REVIEW

#### Introduction

This chapter looks at general results that have been obtained experimentally in natural channels, and is divided into two sections; one on the results from correlation and averaging, one describing the results of spectral analysis of velocity data. Results discussed in the spectral section will later be compared with data from this thesis.

#### Experimental Evidence--Correlation and Averaging

Most work has treated velocity pulsation measurement incidentally since most projects have been designed to determine a minimum time of velocity measurement necessary to reduce the relative error of the discharge measurement to less than some assigned value (Demente'ev, 1962; Savini and Bodhaine, 1971).

The structure of velocity pulsation in the vertical has been studied by many authors (summary in Demente'ev, 1962). It is generally found that pulsation intensity increases with depth and is greatest near the bottom. This conclusion was determined by establishing a mean velocity (generally for a period greater than twenty minutes) and expressing the mean or highest error of

shorter periods of measurement (generally thirty seconds and up) relative to the overall mean velocity as a percentage of the mean velocity. Results based on the standard deviation of the velocity record normalized by the mean velocity show similar results (Grinval'd, 1972) with the intensity of the large scale velocity fluctuations varying from 0.04 at the surface to 0.16 near the channel bed. Results based on hot film and propellor measurements in natural channels (McQuivey, 1973) over five minutes of record are also similar, though disagreement between the hot film and propellor measurements is large and dependent on the type of propellor meter used.

Demente'ev (1962) has presented results on the periods of longitudinal pulsation in mountain channels. The periods were determined by visual inspection of velocity records subjected to increasingly long averaging periods (from 30 seconds to 10 minutes).

On the Chirchik River periods identified in this manner range from 1.5-3.0 minutes to 30-40 minutes during high flow and up to a maximum of 25 minutes during low flows. Average amplitudes of these pulsations were also determined and found to increase from low to high water stages (from 47 cm/sec to 97 cm/sec). However the relative amplitude was approximately constant during different stages. The Pskem River showed periods of approximately 1.5 to 2.0, 3.0 to 5.0, 8.0 to 10.0 and 20 to 25 minutes, and a similar variation in amplitude to the Chirchik

River. Most of the 32 rivers investigated showed similar behavior with the length of the longest period appearing to be partly dependent on river size and discharge. Measurements by Tiffany (1950; 1967) on the Mississippi River indicated periods of 3-4 minutes, 20 minutes and 2.5 to 3.0 hours, based on a five minute travelling averaging interval.

Demente'ev's (1962) general conclusion on velocity pulsations and the factors affecting them were

1. Large scale pulsations occur in natural channels with periods ranging from 1-3 to 40-50 minutes or more. A definite relationship is observed between velocity fluctuations at different points in the cross-section.
2. The magnitude of pulsation increased with the velocity of the flow. However, the absolute amplitude may be greater during low flow on small streams due to the effect of large roughness elements.
3. The magnitude of pulsation increased with the roughness of the bed.

Data on periods associated with macroturbulence (Demente'ev, 1962; Tiffany, 1967) suggest discrete processes perhaps associated with a hierarchy of channel scales, however their data is not supported by more sophisticated analyses (see later section on spectral investigations). In both cases, the maximum periods determined by their analytic technique were of the same order as the length of record (Tiffany, 3 hours; Demente'ev, 30 to 60 minutes) and shorter periods identified by

these researchers are usually a small multiple of their shortest averaging interval.

In fact, the application of a moving average to the velocity data is a form of low-pass filter where the gain function is specifiable providing that the moving average has an odd number of terms (Koopmans, 1974). The discrete nature of the periods identified is related to the jumps in the number of terms in the moving average and better definition of the variance associated with these 'pulsations periods' is achieved through spectral analysis. The principal benefit of these types of moving average analyses is that they indicate that some proportion of the total variance may be assigned to relatively long wavelengths in the data set.

Evidence for the transverse extent of structures considered to be involved in velocity pulsation is derived from synchronous velocity measurements in the vertical and cross-section by several Russian investigators. Demente'ev (1962) measured velocities synchronously at two points in a vertical; at  $0.2d$  and  $0.6d$ , with a separation of  $0.48$  metres. The coefficient of correlation between these two measurements was observed to be  $0.89$ .<sup>1</sup>

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<sup>1</sup>this result is interesting in light of measurements by Bullock et al (1978). In their paper cross-correlations were determined for a variety of bandwidths. It was found that correlations remained high (in the vertical, for the downstream velocity) across the extent of the pipe for narrow bandwidths centered on one Hertz, despite the fact that the overall (i.e., over all bandwidths) cross-correlation showed typical behavior.



Demente'ev claims that this value is typical indicating the existence of structures in the flow whose transverse dimensions are of the order of the size of the stream channel. A more ambitious program of 17 synchronous measurements in a cross-section was also completed on the Varzob River. While no formal analysis of this data was undertaken, the various measurements showed a constant period of pulsation and appeared to be in phase.

Tiffany (1950) includes four synchronous velocity measurements from the Mississippi River, with a maximum separation of 15 feet. Velocity records smoothed with a thirty second averaging interval show excellent agreement. Agreement is not as good for the unsmoothed records. Differences are not unexpected since the maximum perceivable frequency (based on each meter revolution) will be large enough to include 'local' eddy environments for each record. In other words, the large separation of the meters implies that some small scale (i.e., less than 15 feet, which is of the order of the macroscale in the Mississippi River: McQuivey, 1973) features are sampled at each velocity recording position.

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1 (cont'd) Obviously, the current meters used by Demente'ev act as a low frequency filter and pass only correlated scales. Similarly, relatively high normalized cross-correlations (0.50) were observed by Mclean and Smith (1979) over two metres of vertical distance. Fluctuations at higher levels tended to lead those at lower levels due to the lag induced by the mean shear.

### Experimental Evidence--Spectral analysis

A limited number of longitudinal low frequency spectra from rivers are available. The bulk of the published spectra was collected in sand bed channels though some comparative spectra are available from atmospheric work over rough boundaries.

Some of the low-frequency spectral work has been entirely concerned with finding a  $-5/3$  region to conform with prediction by Kolmogorov (1941). Spectra from Grinval'd (1972) and Yokosi (1967) were both obtained under simple conditions in sand bed channels for this purpose. Both papers sampled data in straight sections of sand bed channels with no bed deformation eliminating production associated with those scales.

Both authors found  $-5/3$  slopes in their data though these were extended to scales where local isotropy wouldn't be expected on the basis of results for neutral conditions in the atmosphere (Kaimal et al, 1972). Results from Grinval'd (1972) suggest a broad flat region of power near the lower frequency limit ( $n=.03$  Hz) for surface measurements and changes in slope near the lower ( $-5/3$  range appears for at most a decade) frequency limit for other positions in the flow. Data from Ishihara and Yokosi (1967) and Yokosi (1967) suggest varying results for different channels, with a  $-5/3$  region extending in some cases only to  $n=0.1$  Hz or in others down to  $n=.001$  Hz (scales far larger than the macroscale).

Of course, the existence of a  $-5/3$  slope is only one of the criteria for local isotropy and its existence does not necessarily indicate isotropy. Other criteria include the  $4/3$  ratio between spectral densities of the transverse and longitudinal velocity components (Busch and Panofsky, 1968; Tennekes and Lumley, 1972). Hence the  $-5/3$  region found in these papers is perhaps more indicative of the types of channel conditions (straight, no bed deformation sand bed channels) where the measurements were taken than of local isotropy at these scales.

Several other spectra taken at turbulent measurement scales extend to low enough frequencies to overlap with data from this thesis and the previously mentioned papers. These spectra are available in Nowell (1975) and Nordin et al (1971).

The spectra presented in Nordin et al (1971) were collected in the Mississippi and Missouri Rivers. The principal feature of these spectra is an extensive  $-1$  range extending from approximately .01 to 0.2 Hertz, which in terms of channel scales implies a  $-1$  slope for scales greater than channel depth.

Spectra from Nowell are interesting since these measurements were taken in a gravel bed river (Cheekye River; spectra extend past  $n=0.1$ ). Results indicate that spectral shape is practically invariant with relative depth, though an extensive  $-5/3$  range was visible for measurements near to the surface (this is for scales smaller than those considered in

this thesis). The principal feature of the spectra are peaks near  $n=0.1$  Hertz, implying peaking at scales that are a few multiples of channel depth.

The most detailed set of low frequency measurements are presented by Mclean and Smith (1979). General features of these spectra include convergence to a  $-5/3$  slope at high wavenumbers and an extensive region of  $-1$  slope (approximately two decades) for the longitudinal velocity component. Again, despite the presence of a  $-5/3$  slope other tests indicate that local three-dimensional isotropy was not present at the scales of measurement.

More interesting results are based on the comparison of spectra collected at different relative depths and different positions along a sand wave.

Comparison of spectra at 35 cm ( $y/D=.02$ ) and 214 cm ( $y/D=.14$ ) demonstrate increasing power at high frequencies (principally for scales smaller than channel depth) as the bed is approached. This conclusion is further supported by the behavior of the co-spectra. The co-spectral maximum shifts towards higher frequencies as the bed is approached, indicating that distance from the boundary controls the size of the most energetic eddies (this result is important for universal scaling; Chapter VII).

Secondly, spectra collected at the 100 cm level ( $y/D=0.07$ ) over the crest and the trough of a sandwave are almost

indistinguishable, even at large wavenumbers ( $k=0.1$ , length scale is 0.5 metres). The implication of this result is that spectra in the outer layer are dominated by scales imported from upstream and are not particularly affected by changes in local bed conditions.

#### IV. EQUIPMENT AND DATA ANALYSIS

##### Introduction

In brief, the two problems connected with measuring velocity fluctuations in natural channels are; one, whether the instrument used is sufficiently sensitive to perceive the minimum scales of interest and two, how much energy is involved with scales smaller than those of interest and whether or not this energy is aliased into the data at frequencies of interest.

The periods under consideration in this paper are of a sufficient length to be perceived by a sophisticated current meter and the advantages of instrumentation such as hot film anemometers in perceiving very short period turbulent fluctuations would be of little use in this study. Current meters also have the advantages of stability of calibration over long periods and of robustness proven by countless applications in many channels.

### The Current Meter as a Turbulence Registering Device

In general, it is recognized that there are two effects which distort the output spectrum of turbulence that may be obtained from a current meter suitable for use in a natural channels. These two effects may be referred to as inertial dampening and spatial averaging (Plate and Bennett, 1969).

Inertial dampening is caused by the inertia of the propellor blades and dampens the meter response to high frequency velocity components. This effect makes a direct translation of measured fluctuations of the revolution of the propellor blades into velocity fluctuations untenable. Inertial dampening is probably of little importance for long period fluctuations since the response time of the meter to a step input is significantly less than the velocity gradients (over time) under consideration.

Spatial averaging is controlled by the diameter of the propellers. This effect occurs because the finite size of the propellor leads to an averaging of eddies of the same size or smaller than the propellor dimensions. In general, the effects of these small eddies on the individual propellor blades should cancel each other (Rabot, 1972). Spatial averaging can be important in determining the maximum frequency perceivable by the flow meter.

Limited information is available comparing simultaneous hot film and current meter measurements. Due to the much higher

maximum frequencies sensed by hot films it might be expected that estimates of turbulent energy content should be slightly higher for hot films, though most of the turbulent energy is contained in the frequency range detectable by current meters (<5 Hertz; Nowell, 1975). Comparative measurements from McQuivey (1973) based on five minutes of simultaneous measurement indicate that differences can be extreme and seem to depend on the type of current meter and channel size.

Data obtained in the Atrisco feeder canal and Rio Grande conveyance channel with an Ott minor propellor meter redesigned to produce 30 pulses per shaft revolution indicated that turbulent energy levels were vastly underestimated by the current meter. Data indicated that current meter estimates were approximately 50% smaller near the surface and often only one-third of the hot-film estimate near the bed in the Atrisco feeder canal. Data from the Rio Grande conveyance channel (some data collected with Price standard meter) was much smaller near the surface but often increased to 50% of the hot film value near the bed. Variation in between was extreme with current meter estimates occasionally exceeding the hot film results.

Comparative data collected in the Mississippi River used a standard Price meter with one pulse per shaft revolution. Current meter estimates of the turbulent energy obtained were of the order of 100% larger near the surface and 25% larger than hot-film results near the bed.



Data collected in this paper will be compared with McQuivey's (1973) hot film results in a later section. Though large differences exist in the length of record collected in this paper and McQuivey's it is felt that variance estimates should stabilize rapidly to the long term variance (Raichlen, 1967; approximately equal to long term variance after 1/10 of total record). Data from the Squamish River indicates that variance has approximately stabilized by 100 seconds of record (Figure 6.2).

#### The Current Meter

The current meter used to conduct measurements is an Ott Dynamo meter. This meter uses a precision DC measuring generator to produce a voltage that is proportional to the number of revolutions per minute of the propellor blades and hence to flow velocity. The particular meter under consideration uses a polystyrol propellor with two blades having a diameter of 125 millimetres and a pitch of 0.5 metres.

Minimum flow velocity for this particular propellor arrangement is 0.045 metres per second. This is a problem only at the station nearest to the bed in the Alouette River. Data transformation indicates minimum velocities measured were slightly below this value (calculated minimum was 0.034 metres/sec) and some near stalling of the propellor blades was visually noted. Hence the distribution of extreme values below

the mean is subject to doubt as to accuracy for this particular station.

The transformer constant for this meter was determined from the calibration curve. Regression gave a mean value for the constant of 2.469 volts/metres/second. The linear still water calibration indicates a maximum error of 1.5% in the measuring range between the minimum velocity and 0.2 metres/sec and 0.5% above 0.2 metres/second.

Estimates of the maximum measurable frequency are obtainable from the pitch (p) of the meter. The pitch is the length of water that passes through the meter blades in one revolution. If this is treated as a minimum detectable eddy size, the local mean velocity can be used to transform this to a frequency where

$$n(\text{max}) = U/P = U/.50 \text{ metres}$$

Since velocity records were eventually digitized at one second intervals giving a Nyquist frequency of 0.5 Hertz, this maximum detectable frequency estimate is only a problem when  $U < .25$  metres/second. Mean velocities were only less than this criterion at one point; near the bed of the Alouette River. This indicates that the energy distribution at high frequencies at this station may not reflect the energetics of high frequency eddies. However, the pitch may not provide the most appropriate estimate of minimum detectable eddy sizes for an analogue output current meter. A similar estimate could be based on the

propellor diameter, providing a higher maximum frequency.

It has also been established that the influence of banks and beds causes current meters to perform differently than their rating curve would suggest (Dickinson, 1967). However, the effect in terms of over-registration or under-registration varies with the type of meter. Variation primarily occurs in areas where the velocity gradient is steep over the integrating area of the propellor. The large diameter of the Ott propellor blades (12.5 centimetres) indicates that this effect might have serious consequences for velocity output very near a rough boundary or in a very shallow stream. This effect plus the minimum measurable velocity plus the large physical size of the current meter and weight indicates that sampling very close to the bed or surface is impractical for this system.

The meter suspension system used is different from that of in other studies of low frequencies in rivers. Here, the cable supporting the meter assembly is looped through a U-bolt at the top of a hanger bar which supports the Ott meter and Price weight. This arrangement allows the current meter to rotate in a horizontal plane and follow the direction of the 'mean' current. Since the meter is not rigidly attached to the hanger bar it may also move in a vertical direction. There seems to be two problems with this system; one is the problem of meter stability and the other is comparing 'speed' spectra produced by this system with longitudinal spectra produced by fixed current

meters.

Meter stability was particularly a problem at low local mean velocities. The system seemed to respond by a low frequency swinging back and forth or in and out of the mean velocity direction. Depending upon the response of the meter to non-axial components this could produce an artificial periodicity in the data. Swinging was eliminated as much as possible by quadrupling the fin area of the current meter.

At low frequencies the speed component measured by this system should be very similar to the longitudinal component since as was noted earlier most of the total energy will be contained in the longitudinal component due to production considerations and geometric constraints.

### Data Recording

Velocity data was recorded on a Chart recorder for later digitization. The model used was a Port-a-graph recorder manufactured by Esterline-Angus, division of Esterline Corporation.

During use the recorder was run at 360 mm/minute for ease of later digitization. The chart recorder had a range of voltage scales running from 2 millivolts to 50 volts though the ranges used in this study were the 5V and 10V ranges. The recorder provides an accuracy of plus or minus 0.5% of full scale. For the 10V range this implies a corresponding accuracy of velocity

determination of plus or minus 0.02 metres/second. The recorder also provides a balancing speed for the potentiometer of more than 300 millimetres per second or a time of less than 0.5 seconds for a full scale deflection. This speed is more than sufficient to follow velocity gradients sampled. Recorder drift is also minimized being less than 10 microvolts per hour; essentially no appreciable drift in terms of the accuracy of the recorder.

The velocity signal was low-pass filtered for most stations before being recorded. Data recording without the filter indicated that the trochoid wave output of the meter was appearing in the velocity record along with eddies smaller than the intended digitizing interval.

Roll-off characteristics of the this filter are presented in Figure 4.1. The filter included a X2 gain and a frequency cut-off at 2 Hertz, with power declining to 10% of the signal strength at 2 Hz by 10Hz. This filtering eliminated the signal arising from the meter and also eliminated the considerable energy available for aliasing from frequencies less than 0.5 Hz.

The record lengths used differed in the Squamish and Alouette Rivers. In the Squamish River record lengths used varied from 45-55 minutes. This choice of record length was based on the limited information available on the length of long periods in natural channels (Demente'ev, 1962; Tiffany, 1967) and also represents the length of chart paper available when

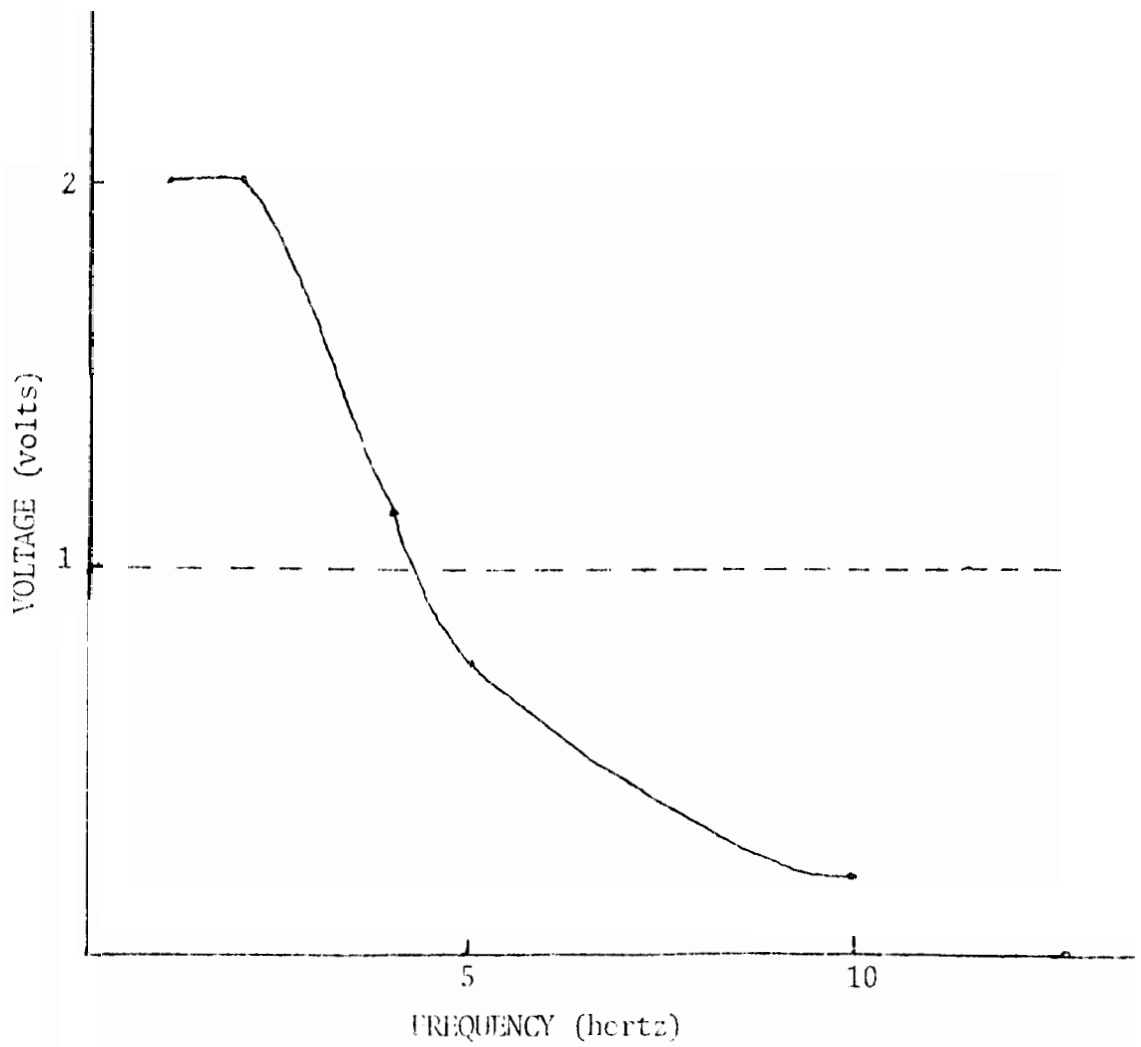


Figure 4.1: Low-pass filter characteristics

the recorder is run at 360 mm/minute.

Record lengths used at the Alouette River (approximately 20 minutes) represented a compromise between a reasonable minimum frequency (.001 Hz) and the desire to eliminate discharge change between the first measurement taken in the vertical and the last one.

### Bed Sampling

Sampling of depth variation along the thalweg was undertaken in both the Squamish and Alouette Rivers. Records in the Squamish River were collected with a Raytheon fathometer (Hickin, 1978). The length of channel covered is indicated in Figure 5.1.

The fathometer profile was divided in sections representing stations used in a study of mean and secondary current (Hickin, 1978). Distances between these stations were known and these distances were used in conjunction with measurements taken from the fathometer chart to produce approximate scales for each section of fathometer record. The computed scales varied between 1 inch to 31 metres and 1 inch to 135 metres. Part of this variation was due to changes in boat speed but these large differences also reflect the difficulty of locating stations accurately.

These calculated scales were then used to calculate a digitizing interval for each section that represented 10 metres

of actual distance. This 10 metre digitizing distance was chosen to keep the minimum digitizing interval of the order of one tenth of an inch.

There are two problems associated with this analysis. The large relative boat speed to chart speed principally determined the large digitizing distance used. Unfortunately, this provides a Nyquist wavelength of 20 metres. The limited data available on dunes in gravel bed channels (Galay, 1967) indicates gravel dune wavelengths of 12 and 18 metres for a comparable discharge in the North Saskatchewan River. These wavelengths would be undetectable by this analysis.

The small vertical scale used on the fathometer record (8.5 feet of relief/inch) and the thickness of the line representing the bed means that small variations (and short wavelength variations) in profile height are not represented very accurately.

Bed sampling in the Alouette River was a much simpler program. The thalweg was sampled by measuring the depth every fifteen feet with two range poles tied together with a fifteen foot length of rope.



### Digitization

Digitization of velocity records was carried out at two locations; the Department of Mechanical Engineering, UBC, and Tetrad Computer Applications, Ltd., Vancouver. All velocity records were digitized at an interval of 0.236 inches, corresponding to a one second digitizing interval. Both systems used an ISOTRONICS GRADICON digitizer to record x and y coordinate locations every 0.236 inches. Recorded data were output to an IBM keypunch, producing a card file.

### Data Analysis

Preprocessing of digitized data was limited to testing for stationarity in the mean and the variance. Testing for stationarity was accomplished with a runs test (Bendat and Piersol, 1968). The runs test used was a non-parametric test which compared the number of runs (crossings of some group median value) in a series of means and variances calculated from serial, shorter sequences of the overall data series. The length of these shorter series was designed to be greater than the limit of correlation in the data. All velocity and bed series were determined to be (weakly) stationary.

Spectral analysis was performed by a package called MIDAS, available on MTS. The handbook for this package (DOCUMENTATION FOR MIDAS) indicates that a Fast Fourier Transform is used to produce autocorrelation estimates which are then processed to

produce spectral estimates following the procedures outlined in Jenkins and Watts (1968).

In common with standard procedures for an autocorrelation approach, the program allows the operator to specify a truncation ( $m$ ) and to choose a window function. The combination of these two parameters determines the resolution bandwidth of the spectrum (resolution bandwidth is a function of  $1/m$ ; Jenkins, 1961) and the parameters may be adjusted in an attempt to compromise between fidelity and stability (referred to as window closing; Jenkins and Watt, 1968), keeping within the degrees of freedom imposed by the record length.

Large resolution bandwidths were used in this analysis (estimate based on the truncation is .02 Hertz), partly to produce reasonably smooth spectral curves with a limited number of points and partly to reduce spurious peaks in the high frequency range.

The principal problem with the package is related to the intervals on which frequency points can be chosen. Since the choice of frequency points is limited to some constant (i.e., over the frequency range of .001 to 0.5 Hertz) arithmetic interval, when data are plotted as the logarithm of the spectral density versus the logarithm of frequency very few points appear in the lowest frequency decade and points are clustered in the highest frequency decade (0.1 to 0.5 Hertz; generally two to three points are determined per resolution bandwidth implying

approximately 60 points over the frequency range) and there is an obvious definition problem for the lowest frequencies.

In order to produce a more constant density of spectral estimates over the logarithm of the frequency range, spectral estimates were calculated for three overlapping intervals (.001 to .01 Hz, .001 to .2 Hz, and .1 to .5 Hz). The problem with this approach is that spectral estimates calculated at points in the range .001 to .01 Hertz are not independent due to the large resolution bandwidth and decreasing this bandwidth de-stabilizes the high frequency estimates. If a different resolution bandwidth is used for each interval then the overlapping ranges do not match at the endpoints since the computation bandwidth extends varying distances around the end point of each range. Also, the bandwidths necessary to resolve the range between .001 and .01 Hertz are small enough that the degrees of freedom fall below the confidence level for the shortest series.

Consequently, large bandwidths were used for all the analysis and limited emphasis is placed on interpretation of spectral shape of the lowest frequencies. Resolving the lowest frequencies in the data accurately seems to be a common problem (McLean and Smith, 1979).

## V. THE FIELD SITES

### Introduction

This chapter is written in two parts; one dealing with the Squamish River, one describing the North Alouette River. The level of descriptive information available for the two rivers varies widely. Since measurement on the Squamish River was done in conjunction with a major project on mean flow structure in bends a wealth of published (see Hickin, 1978; 1979) and unpublished material is available. Conversely, the level of background material on the North Alouette is very limited. Each part is written in three sections; a section on the general area and discharge conditions, a section detailing upstream planform and channel bed conditions and a section on local information connected with the measurement site.

The purpose of this chapter is threefold. One purpose is to provide the necessary background information on measurement conditions. The second purpose of the chapter is to use spectral analysis in an attempt to identify longitudinal wavelengths that may be of use in collapsing low frequency velocity spectra. Thirdly, flow and channel behavior that may be of importance in providing local low frequency velocity spectrum interpretation

is detailed.

## PART I -- The Squamish River

### General Conditions

The Squamish River is a proglacial river which lies north of Vancouver, British Columbia and drains into Howe Sound. The area of the drainage basin is 2330 square kilometres (Inland Waters Directorate, 1977).

Mean monthly discharge (Inland Waters, 1977; based on 26 years of record near Brackendale) is 244 cumecs. Maximum monthly discharges occur during the summer months and the general recession from snowmelt continues until the minimum monthly discharges which occur during the late winter period (January to March). Maximum daily and instantaneous discharges occur during the fall period (though some of the annual flood population is associated with the peak discharges from the June snowmelt) and appear to be associated with rainfall on exposed ice in the upper parts of the basin.

Figure 5.1 shows 5 kms. of the planform behavior of the Squamish River (the measurement site is near station 27). The most obvious feature of this channel reach is the general

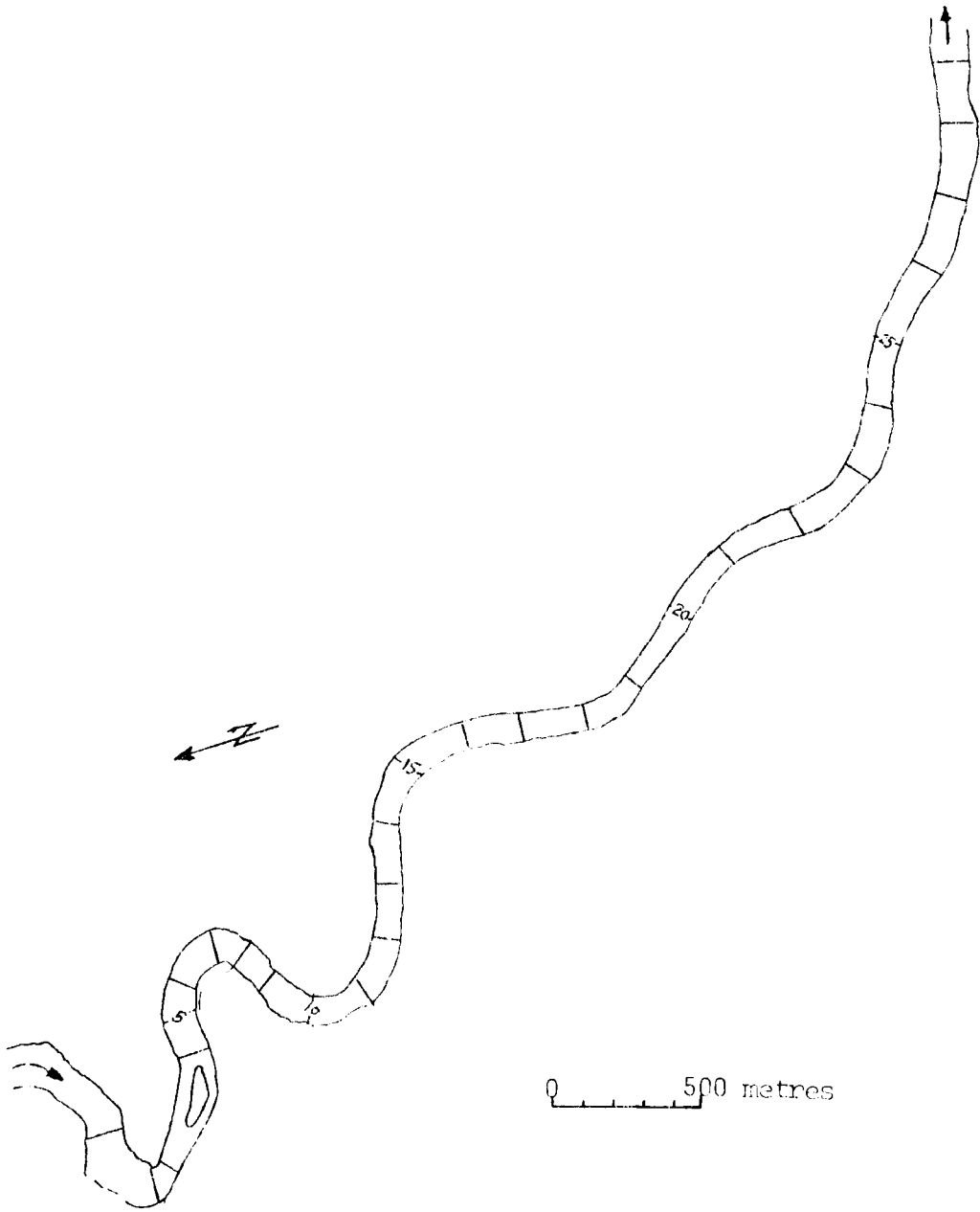


Figure 5.1: Planform behavior of the Squamish River reach.  
(adapted from Hickin, 1978)

decline of 'meander amplitude' and apparent increase of wavelength and radius of curvature to width values ( $r/w$  values vary from 1.2 at the most upstream bend to 8.7 in the neighborhood of the measuring site; Hickin, 1978).

The Squamish River is not a 'classic' channel meandering in its own alluvium. In fact, it is similar to the 'wandering' channel type described by Neill (1976). General characteristics of this type display neither clearly meandering nor braiding planform behavior, but rather alternating sequences of irregular loops and reaches divided by islands. Several of the upstream bends of the Squamish River abut against the granitic valley sidewall. In other areas (notably sections 19 and 24) the channel is narrowed on the western side by alluvial fans. The total influence of the granitic sidewall on the bed profile is unknown. Outcrops of granite project into the stream in several areas and it is possible that some of the large symmetrical features projecting above the surrounding channel bed are of granitic origin (see Figures 5.2 and 5.3).

Observations of delta sediments (Hoos and Vold, 1975) and observations in the test reach indicate that the Squamish River is a gravel bed channel. Estimates based on gravel material on point bars (Hickin, 1978) indicate sizes ranging from 5 to 50 mm. A similar range of particle sizes is expected for bed material. Deposits of sand size material appear to be limited to occurrences as veneers on some point bars.

### Reach Characteristics

Figures 5.2 and 5.3 present the bed elevation of the thalweg from station 1 (this is zero metres on Figure 5.2) to station 29 (Note--horizontal scale changes from Figure 5.2 to 5.3). Discharge when this trace was taken was around 425 cumecs--where this is the average discharge found by Hickin (1978) for the summer period. Sampling procedures for the bed elevation data are detailed in Chapter IV. This section will discuss general and statistical features of this bed profile.

The most obvious feature of the bed trace is the extremely deep pool occurring near the axis of the most upstream bend (station 2, Figure 5.1) This feature appears to be similar to 'scour holes' observed by Neill (1976) which were characterized by;

1. maximum depths generally 2.5 to 3.5 times the average depth under ten year flood conditions.
2. Location at bend apexes involving large angular deflections (at least 70 degrees, which corresponds to an r/w of 3.2; the minimum r/w value is 1.7).
3. most of the scour holes occur where the river impinges on the valley wall.
4. These features are persistent and change little from year to year.

The feature on the Squamish River fits these general conditions.



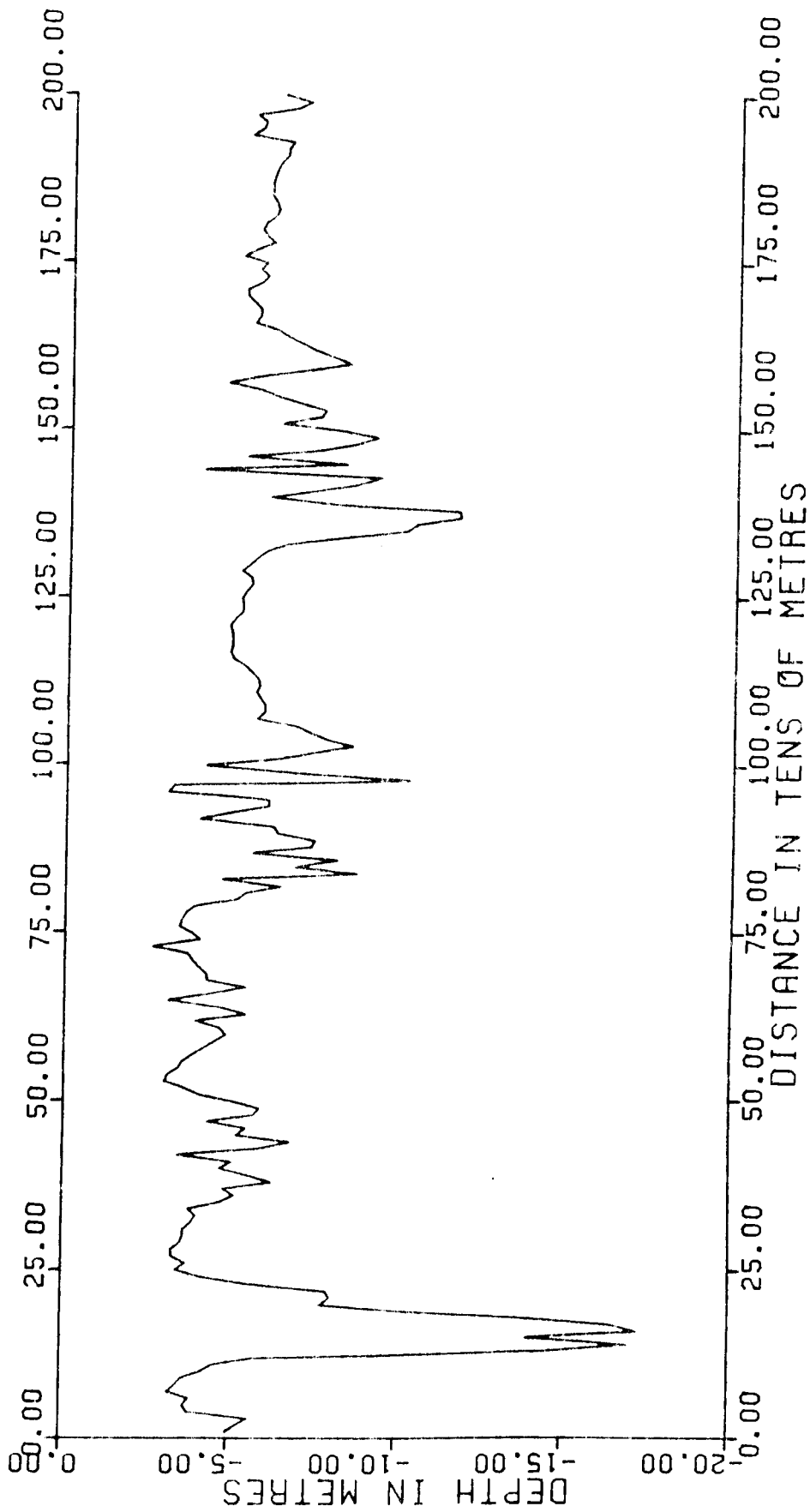


Figure 5.2: Squamish River bed elevation--Section I

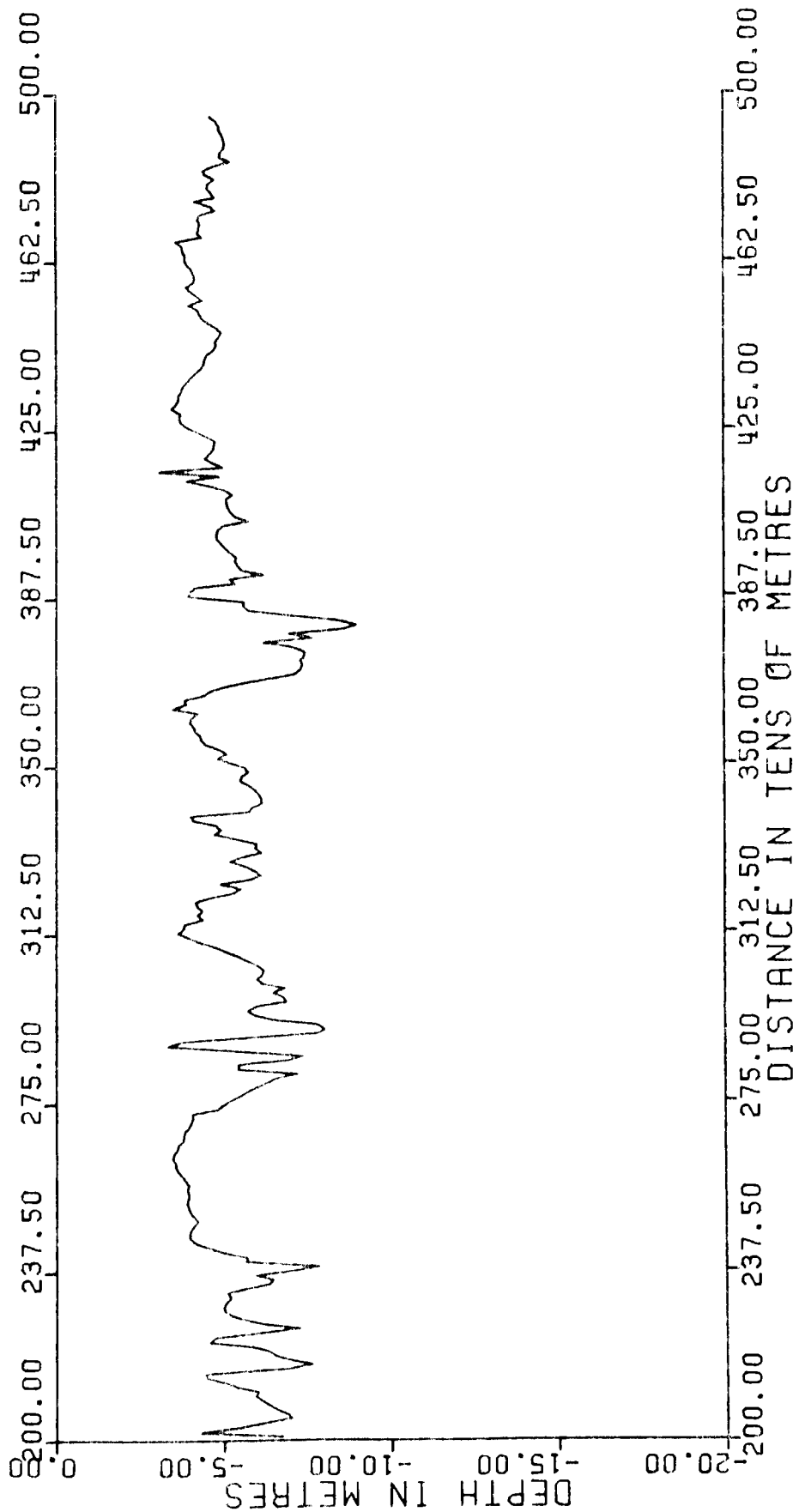


Figure 5.3: Squamish River bed elevation--Section II

The most interesting feature of these scour holes in terms of this thesis is their persistence over time. As Neill notes, their persistence, despite evidence of active gravel movement during the annual flood (Athabasca River), implies the presence of some stable eddy system. Neill associates one scour hole, which occurs in a straight reach, with a nearby horizontal eddy occupying a bank embayment (i.e., a separation eddy; equivalent to Matthes (1947) bank eddy, which are Class II, Rotary Continuous features).

General observation on the relation between flow structure and these type of large holes is limited to one paper (Moore, 1970). Observations by Moore were undertaken over a pit dredged into the bed of the Missouri River. This pit was approximately 450 metres long, 150 metres wide and on average 8 metres below the local river bottom. The feature persisted over three water years with peak discharges reaching 17000 cumecs over a sand bed.

Moore observed that above mean flow velocities of 0.5 metres per second, a large horizontal eddy formed over the pit. Surface observations showed that floats placed over the center of the pit were approximately stationary while floats near the bank moved upstream at velocities between 0.2 and 1 metres per second. This horizontal eddy appeared intermittently. Moore indicated that there was a period of approximately 6 and one-half minutes during which the eddy formed, built to a

maximum, and dissipated. Following this period was approximately 3 and one-quarter minutes of downstream flow. The period when the eddy was present was associated with vigorous kolk activity and deposition of sand size material between the dredge hole and the bank.

No similar flow structure was observed over the scour hole on the Squamish River. The mean flow structure was dominated by two elements; a large separation zone in an embayment on the concave bank and a mean flow with a strong helical component over the actual scour hole (see Hickin, 1979; Figure 1). Macroturbulent elements were generated in the shear zone at the margin of the mean flow. These features ranged up to 10 metres in diameter; however, observation indicated they remained identifiable only several diameters downstream.

The other interesting general feature of these bed profiles is what appears to be a tendency for the variance of the bed features to decline with decreasing amplitude and increasing wavelength through the reach. However, this tendency is not confirmed by a runs test (Bendat and Piersol, 1968).

#### Statistical Approaches--Longitudinal Scales

The bed elevation data were digitized at a constant interval of 10 metres. Common serial analytic techniques such as spectral analysis may then be applied to the spatial series.

The bed elevation data were analyzed, for the overall

series and two subseries; one including the tight bends of the upstream part of the reach (Section I, Table 5.I), the other including the relatively straighter section (Section II, Table 5.I) downstream. This was done for several reasons. It was felt that it was worthwhile to test for variation in longitudinal scales under different planform conditions. If the scales were different, then the results from the straight section would be more representative of local conditions at the measurement site. Secondly, the division of the overall series represents an attempt to characterize a visually obvious pool/riffle sequence spectrally.

Accordingly, the depth series was broken into two parts, Section I, the 'meandering' section running from stations 1 to 15 (Figure 5.2) and Section II, the 'straight' section running from 15 to 29 (figure 5.3). The division point, of course, is fairly arbitrary. However, it was chosen to include the three tightest bends into the meandering section and also to provide two series of fairly similar lengths, since the number of points in the data series is limiting for the resolution bandwidth.

Summary statistics for all three series are listed in Table I. Data are from the digitized bed profile and also from Hickin's (1978) field survey (mean widths and means of the average cross-section depths).

Most of the variation of these measures is obvious. The presence of the scour hole in Section I is reflected in the

TABLE 5.1--SQUAMISH RIVER SUMMARY STATISTICS

	MEAN X-SECTION DEPTH (metres)	MEAN WIDTH (metres)	MEAN THALWEG DEPTH (metres)	STANDARD DEVIATION (metres)	KURTOSIS	SKENNESS
SECTION I	4.5	94.2	5.9	2.4	7.2	-2.5
SECTION II	3.8	108.1	5.1	1.1	0.2	-0.8
OVERALL	4.2	101.4	5.4	1.8	12.7	-2.7

relatively larger standard deviations and negative skews. More interesting is the variation in the kurtosis values. The large kurtosis values of the meandering section reflect the large serially isolated variance contribution of the 'pools' and the relatively low variance contribution of the smaller bed elements in between. Removal of these large 'pool' contributions indicates that the marginal distribution (as in Section II) of the smaller scale bedforms is relatively more normal. The Section I distribution appears to represent two superimposed elements of the bedform hierarchy.

Spectra of these three series are presented in Figure 5.4.<sup>1</sup> Several features of these spectra are apparent:

1. a low wavenumber<sup>2</sup> peak at 0.03 cycles per 10 metres (x=340 metres) in the Section I spectra.
2. a broad high wavenumber peak near a wavenumber of 0.25 (x=45 metres) in the Section I and Overall Spectra.
3. a larger percentage of low wavenumber spectral density for the Section II spectra.

The peak at x=45 metres appears to be associated with the large narrow 'spikes' that occur in the depth series.

Discussion of the peak value observed for the Section II spectra is more interesting. It is worthwhile to compare this

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<sup>1</sup> The spectra of the two sections of record are significantly different at p=0.01 (test is from Bendat and Piersol, 1968).

<sup>2</sup>For these spectra wavenumber is just the reciprocal of wavelength.

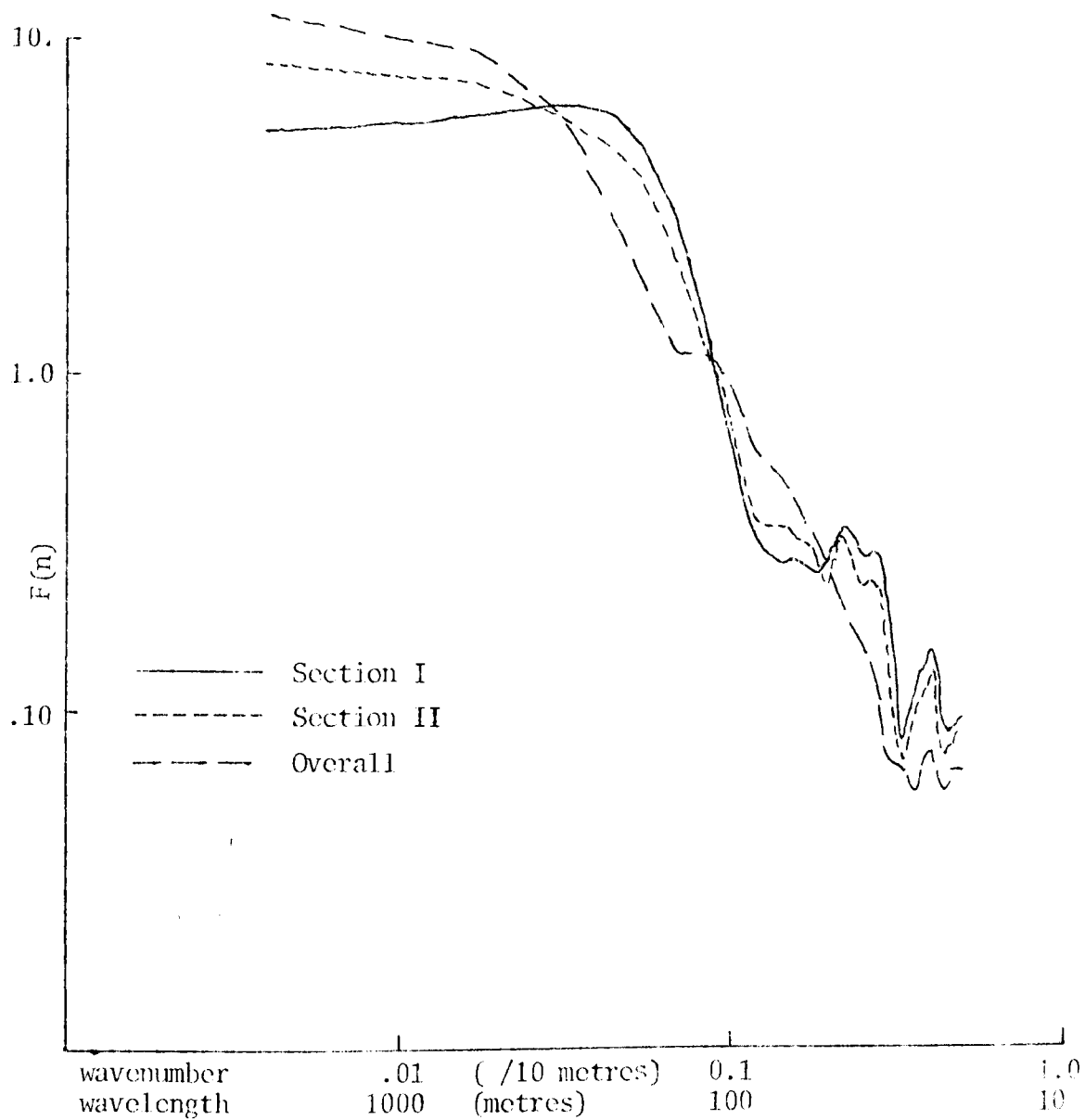


Figure 5.4: Squamish bed elevation spectra



value to other typical dominant bedform wavelengths and also to Squamish planform behavior.<sup>3</sup>

A path meander wavelength estimate for Section I is

$$x^* = (l/H/2) (l^*/l) = 1330 \text{ metres}$$

where  $l$  is the downvalley distance,  $H$  is the number of crossings by the channel of this downvalley line and  $l^*/l$  is the channel sinuosity. The downvalley meander wavelength is

$$x = x^*l/l^* = 952 \text{ metres}$$

and

$$x/w = 10.1$$

which corresponds to the traditional empirical equation (Leopold and Wolman, 1960).

The peak of the depth series, which corresponds to  $3.4w$  is small compared to most other estimates. Sperare (1974) found important depth spectral peaks in the range from  $5-20w$ ; results from a braided outwash channel (Church, 1972) were in the range of  $5-8w$ .

Second, this depth spectral peak is small relative to the planform pathlength meander wavelength. Sperare (1974) demonstrated that depth spectral peaks tended to occur at

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<sup>3</sup>Channel wavelengths can be related either to a pathlength ( $x^*$ , the distance along the channel, hence estimates from the depth series) or to the 'downvalley' or axial distance ( $x$ , as in visual studies; e.g. Leopold, Wolman and Miller, 1964). The two different measurements are related by the channel sinuosity ( $S_i$ ) as

$$x = x^*/S_i$$

one-half of the path meander wavelength.

While this particular wavelength ( $3.4w$ ) is visible in the depth series (see Figure 5.2), it is unreasonable to call it a 'pool and riffle' sequence. The wavelength is composed of a mix of bed elements. Starting at the zero mark on Figure 5.2, the bed elevation correspond to the following morphological elements. First a decline into the scour hole, second a rise over a small bar on the downstream lip of the scour hole, third a slight deepening, then a rise in the area of the midchannel bar marked on Figure 5.1, and then a general decline into the pool associated with the apex of the second downstream bend. In other words, the identified wavelength is associated with a fortuitous mix of bed hierarchy elements.

### The Local Site

Figure 5.5 presents a channel cross-section near to the measurement site. Measurements were taken from a river boat moored to the Squamish River cableway (Brackendale gauging station; Inland Waters Directorate).

The original measurement program involved collecting lengthy velocity records at approximately 10 points in the cross-section, at a near constant stage. However, due to equipment and access problems only two velocity records were collected during the relatively constant summer discharges (1 and 2, Figure 5.5). Both of these records were collected on

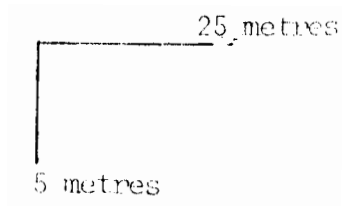
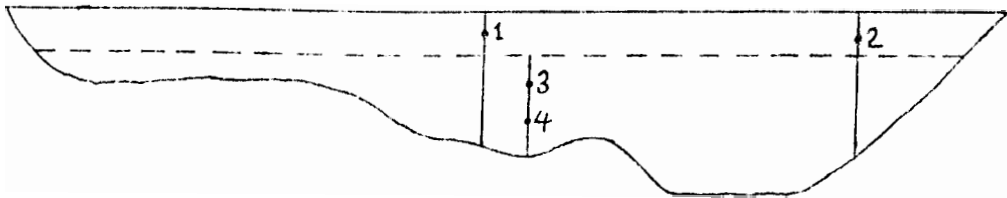


Figure 5.5: Cross-section near Squamish River measurement site.

August 25, 1977 at a daily discharge of 405 cumecs (Inland Waters Directorate, 1978). On a subsequent return to the Squamish River in late October, river levels had dropped considerably. Two records (numbers 3 and 4) were collected on October 3, 1977 at a discharge of 91 cumecs. At this point data collection on the Squamish River was abandoned and efforts switched to the North Alouette.

Observed macroturbulence during data collection was primarily of the 'kolk' type, expressed on the channel surface as non-rotating boils which decayed through radial expansion (Matthes, 1947). The appearance of the features in any region of surface was regular and rapid (i.e., a few seconds between) and the surface at any time tended to be covered with a dense network of these features. While the features were also apparent at lower discharges they did not appear to be as dense or regular. Dissipation of individual 'boils' was rapid and occurred within a few tens of feet.

Other macroturbulence features observed were also local in effect. Large suction eddies (similar to the classic Rankine vortex where there is a decline of water surface elevation over the center of the features; Rouse, 1963) were observed downstream of a rock ledge projecting into the river. These features are difficult to fit into Matthes classification since they should occur under Type II, Rotary, Continuous, which does not include features which are regularly shed. Again, despite

the fact that these features are relatively energetic, they dissipated within 100 feet downstream of where they were shed.

## PART II -- The North Alouette River

### General Conditions

The North Alouette River is an unregulated channel located to the east of Vancouver, British Columbia. This river joins the Alouette River and both flow into the Pitt. The basin area of the North Alouette River above the WSC gauge is 37.3 square kilometres. The mean monthly discharge is 3.1 cumecs (based on 19 years of record; Inland Waters Directorate, 1977). Low monthly discharges tend to occur in the late summer-early fall period (August to September) and maximum monthly discharges occur during the November to February period.

The channel pattern of the section of the North Alouette for which bed elevation records were collected can be classified as irregular (Kellerhals, Church and Bray, 1976). No repeatable pattern was observed and the channel tended to very broad bends with an occasional tight bend. One large, permanent island occurred in the reach (670 metres from start of trace; see Figure 5.6). Bed sampling continued up the right hand channel as

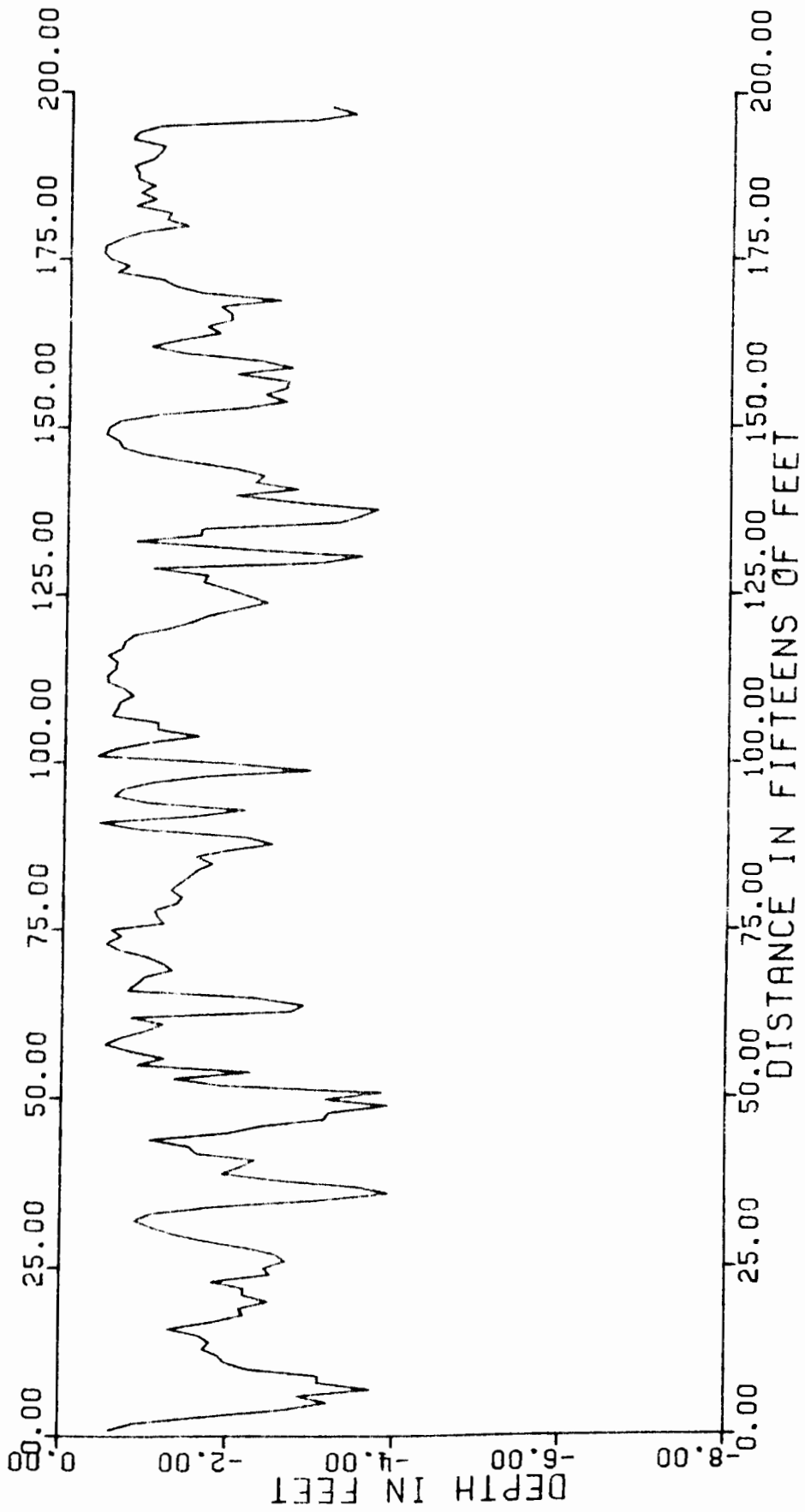


Figure 5.6: North Alouette bed elevation

this one appeared to be larger. The channel split extended for approximately one hundred metres.

The North Alouette is a gravel bed river. The bottom was visible during bed elevation sampling and material size appeared to range between 10 and 100 millimetres. No small scale forms (i.e. less than the sampling distance) were observed during wading.

### Reach Characteristics

Figure 5.6 presents the bed elevation of the thalweg of the North Alouette River upstream of the measurement site. Sampling procedures for these data are detailed in Chapter IV. Discharge on the day the bed elevation series was collected (June 24, 1978; discharge was 0.6 cumecs) was lower than when velocity records were collected (see footnote 1; Chapter VI). The trace represents 3000 feet of channel distance (approximately 80w based on the width during collection of velocity records). No obvious visual features occur in Figure 5.6 other than an irregular succession of 'pools' and shallows.

### Statistical Approaches -- Longitudinal Scales

The mean depth value for the channel reach was 0.50 metres with a standard deviation of 0.27 metres. The distribution of bed heights was non-normal with negative skew (-0.62) and negative kurtosis (-0.49).

The bed spectrum of the North Alouette River is plotted in Figure 5.7. Spectra in this figure are plotted as a function of wavelength\* to eliminate the effect of differing sampling intervals (consequently the power distribution of the Squamish bed spectrum is reversed from Figure 5.4). The Squamish spectrum plotted in this figure is based on the Overall depth series, but was recalculated for a different set of wavenumbers.

It is interesting to compare the bed elevation spectra of the Squamish and North Alouette Rivers. As might be expected on the basis of the 'size' of the two systems, spectral density is skewed to longer wavelengths in the Squamish River. Both spectra have similar slopes in the wavelength range from 10 to 100 metres. However, there appears to be no simple measure of the 'size' of the systems (such as channel width) to collapse these two spectra onto a universal curve.

### The Local Site

As mentioned earlier, the velocity sampling program on the North Alouette was only instituted after problems occurred on the Squamish River. Velocity series were collected for one vertical during one day (April 21, 1978). As a result the overall data set allows comparison of velocity spectra from

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\*this is just the inverse of wavenumber -- similar to the frequency/period combination for spectral transforms of time domain data. Note the appearance of twice the sampling interval (the Nyquist wavelength) as the minimum wavelength.



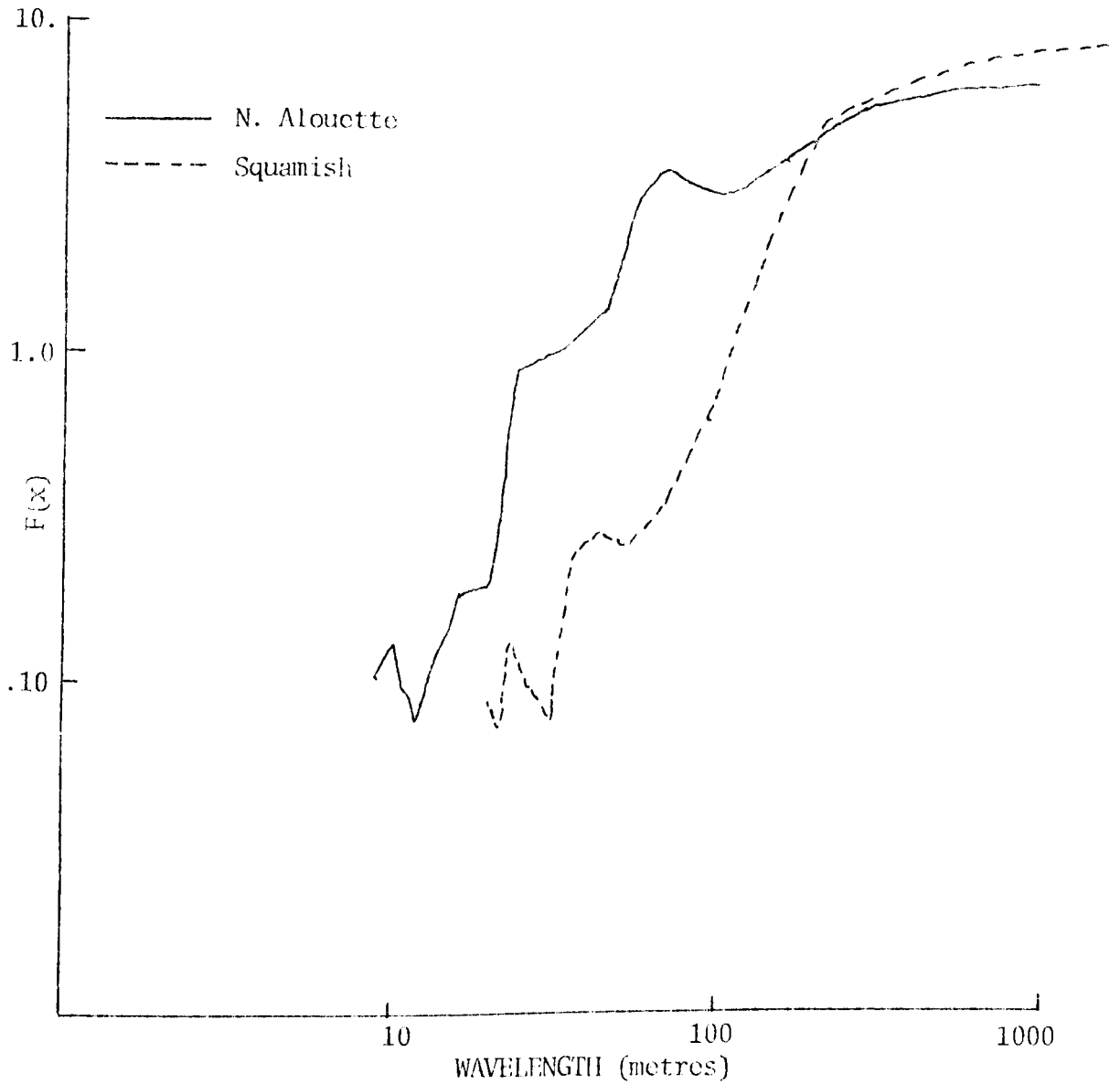


Figure 5.7: Comparison of N. Alouette and Squamish bed elevation Spectra

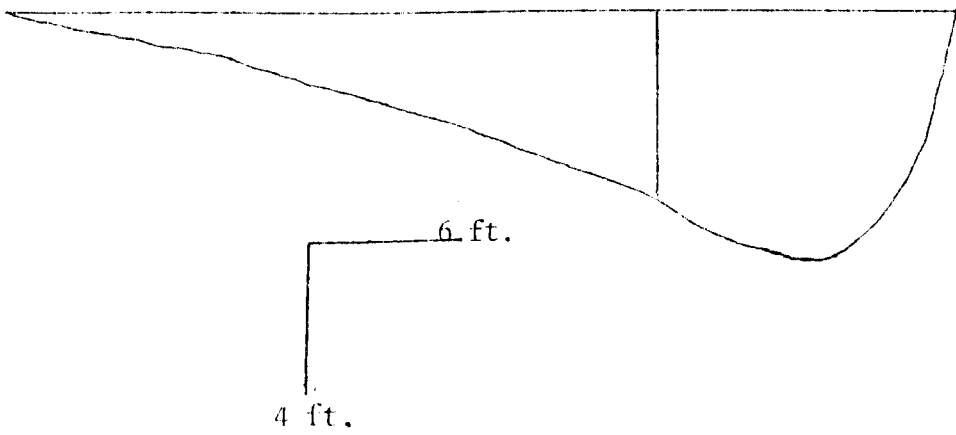


Figure 5.8: N. Alouette measurement cross-section

1. two rivers of different 'sizes'.
2. one channel at two widely different discharges
3. from one comprehensive vertical in one channel.

Figure 5.8 presents the measurement cross-section. Seven velocity records were collected at the marked vertical using a current meter suspended from a rope across the channel. These measurements were collected on a falling stage (See Footnote 1, Chapter VI, for more discussion).

Despite a comprehensive survey of the reach none of the classic macroturbulence features were observed. Conditions of the water surface during measurement were very calm and no 'kolk' activity was noted as on the Squamish River, despite being only thirty or so metres downstream from a shallow riffle area where the surface of the flow was very disturbed.

## VI. PROBABILISTIC STATISTICS

### Introduction

This chapter presents the variation of the general statistics of the velocity distribution; that is, its moments and some of their variation with sample size and relative depth. Where possible data from the Squamish and North Alouette Rivers will be compared to other results.

Data discussed in this section are presented in Table 6.I.

### Mean Velocity Profile

In general, the universal velocity profile for open channel flows is written as (Tennekes and Lumley, 1972):

$$U/u^* = 1/k' \ln y/k + \text{constant}$$

though a defect law is also appropriate.

More commonly, though, results from an individual channel are presented in the form of the local mean velocity against the logarithm of relative depth. Occasionally the local mean velocity is non-dimensionalized by the channel average flow velocity to compare profiles collected across a section (Raichlen, 1967) or scaled by the shear velocity ( $u^*$ ) or  $U(\text{infinity})$ , the velocity outside the boundary layer, which is

TABLE 6.1--STATISTICS OF THE VELOCITY DISTRIBUTIONS

	Y/D	MEAN VELOCITY (m/sec)	STANDARD DEVIATION (m/sec)	INTENSITY	SKEWNESS	KURTOSIS	
N. Alouette	0.78	.513	.059	.11	-.60	.52	
	0.71	.406	.086	.21	-.54	-.04	
	0.59	.377	.079	.21	-.34	-.01	
	0.48	.358	.094	.26	-.03	-.50	
	0.41	.235	.098	.21	-.49	-.01	
	0.35	.356	.100	.28	-.29	-.11	
	0.22	.182	.062	.34	-.14	-.24	
	Squarish	0.66	.862	.071	.08	-.26	.36
		0.33	.721	.087	.12	-.20	-.63
0.72		.961	.088	.09	-.38	.16	
0.80		1.475	.111	.08	-.37	.12	

often interpreted as the surface velocity in an open channel. Both of the last two parameters are appropriate outer layer velocity scales.

This logarithmic form generally provides a good fit to velocity data from the outer layer, though divergence from this form occurs near the bed and often, near the surface (e.g., Raichlen, 1967).

Mean velocity data for the North Alouette River are presented in Figure 6.1. As is obvious from a cursory examination of this figure the data are not well explained by a semi-logarithmic curve. The explanation of the poor fit lies in the sampling procedure. Velocity measurements were taken over a long period of time (approximately 8 hours) with the measurements falling into two groups. Series 1, 3 and 6 ( $y/D = .78, .58$  and  $.35$ , respectively) were collected early in the measurement period before instrument problems caused a gap of several hours in data collection. As records were collected on a falling stage<sup>1</sup> the second group of measurements (series 2, 4, 5 and 7) tend to be displaced to the left on Figure 6.1 relative to the first three. Treating these measurements as two separate groups (Figure 6.1B) slightly improves the fit to the curves and decreases the velocity gradient. However, several of the

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<sup>1</sup>stage change during the measurement period (April 21, 1978) was from 1.38 to 1.29 feet, and the corresponding discharge change was from 3.1 to 2.3 cumecs (data kindly supplied by Mr. C. Naqy, Inland Waters). Stage changed too slowly to induce any trend in the velocity series which averaged only 17 minutes in length.

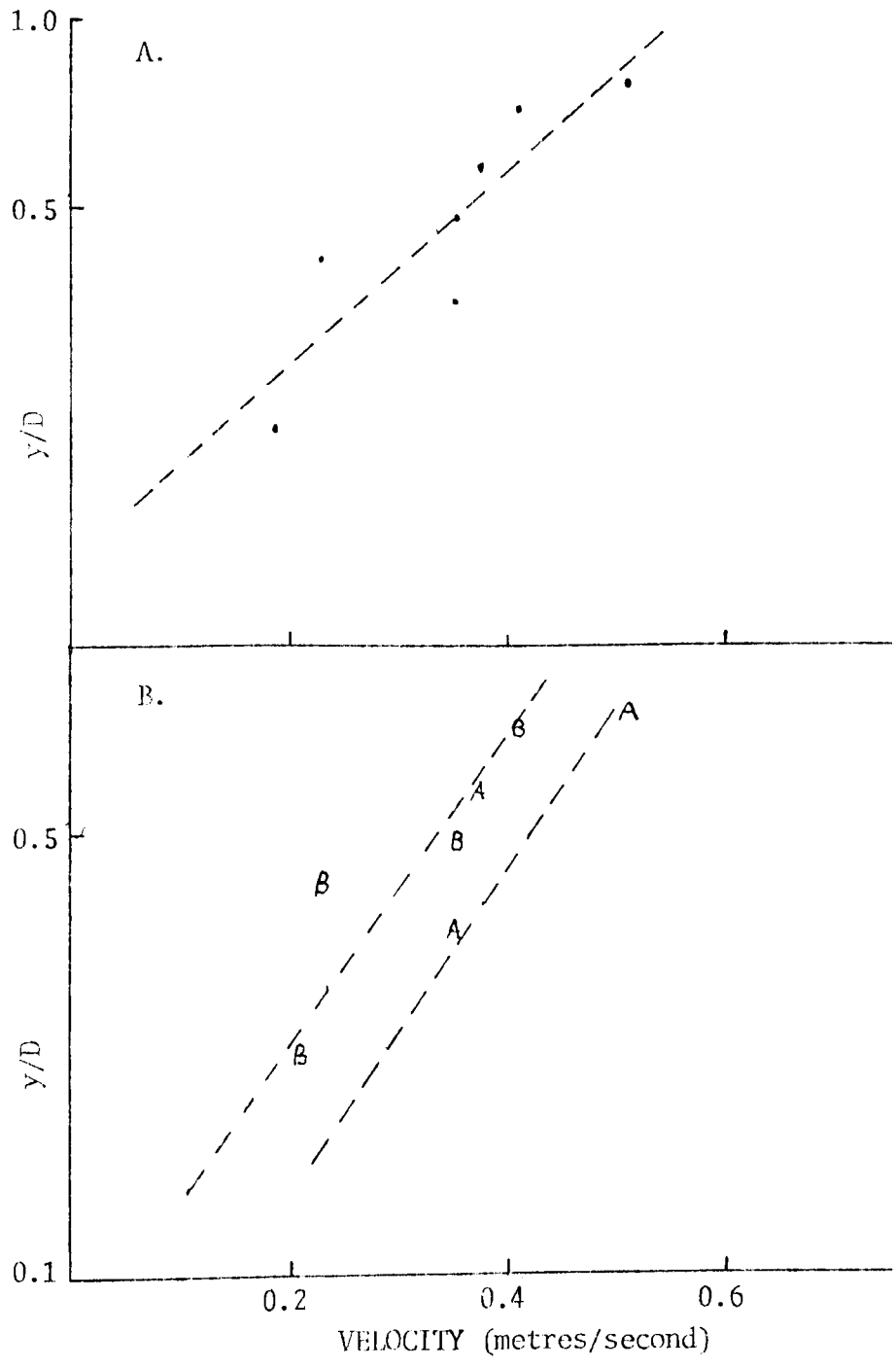


Figure 6.1: North Alouette River velocity profile

measurements still appear anomalous, notably those at  $y/D=0.58$  and  $0.41$ , both of which appear to be low. No explanation is known for the large divergence of these two measurements.

Certain problems are introduced by the variance of the mean velocities from semi log-linear form. The local mean velocity is used for scaling the R.M.S. velocity and also is used in producing non-dimensional frequencies for plotting universal spectra. The variations in the mean velocity will be reflected in parameters incorporating the mean velocity.

#### Accumulation of RMS and Mean Velocity

Analogous to common practise in turbulence research, an intensity statistic has been defined. This statistic is simply the RMS (root-mean-square) velocity scaled by the local mean velocity.

As was pointed out previously (Chapter IV) before intensity statistics from this paper can be directly compared with other results, several factors have to be taken into account. These are: differences in sampling interval and consequently sampling of energy at different scales, differences in record length (additional sampling of energy at large scales) and differences in intensity related to boundary type and Reynolds number. The first point was discussed in Chapter IV and will not be repeated here.



Figure 6.2 presents the accumulation of mean and RMS velocity with record length. The series used in this figure is Series 4 (Squamish River,  $y/D=0.8$ ) and was chosen principally because it is the longest one. Quantitative differences would be expected with different relative depth values. Demente'ev (1963) indicates that the mean error of shorter periods of measurement relative to the overall mean velocity decrease much more slowly near the bed than near the surface. Similar results would be expected here.

Figure 6.2A presents the effect of increasing sample size on the average RMS velocity; where the average RMS velocity is the average of a number of samples of some length drawn from the overall record. The number of samples taken varied with the length chosen; 25 independent (i.e., non-overlapping) samples were drawn for short record lengths while only five samples were drawn when record length had increased to 500 seconds. Figure 6.2C presents the effect of increasing record length on the standard deviation of the samples drawn from the population.

As can be seen in Figure 6.2A the RMS velocity increases rapidly with record length and after 50 seconds of record is within 4% of the value derived from the total record length (time to the first zero crossing of the autocorrelation function is 20 seconds, where this can be used to estimate the "largest" eddy). The variance stabilizes once sampling occurs across the major energy containing features. The RMS velocity

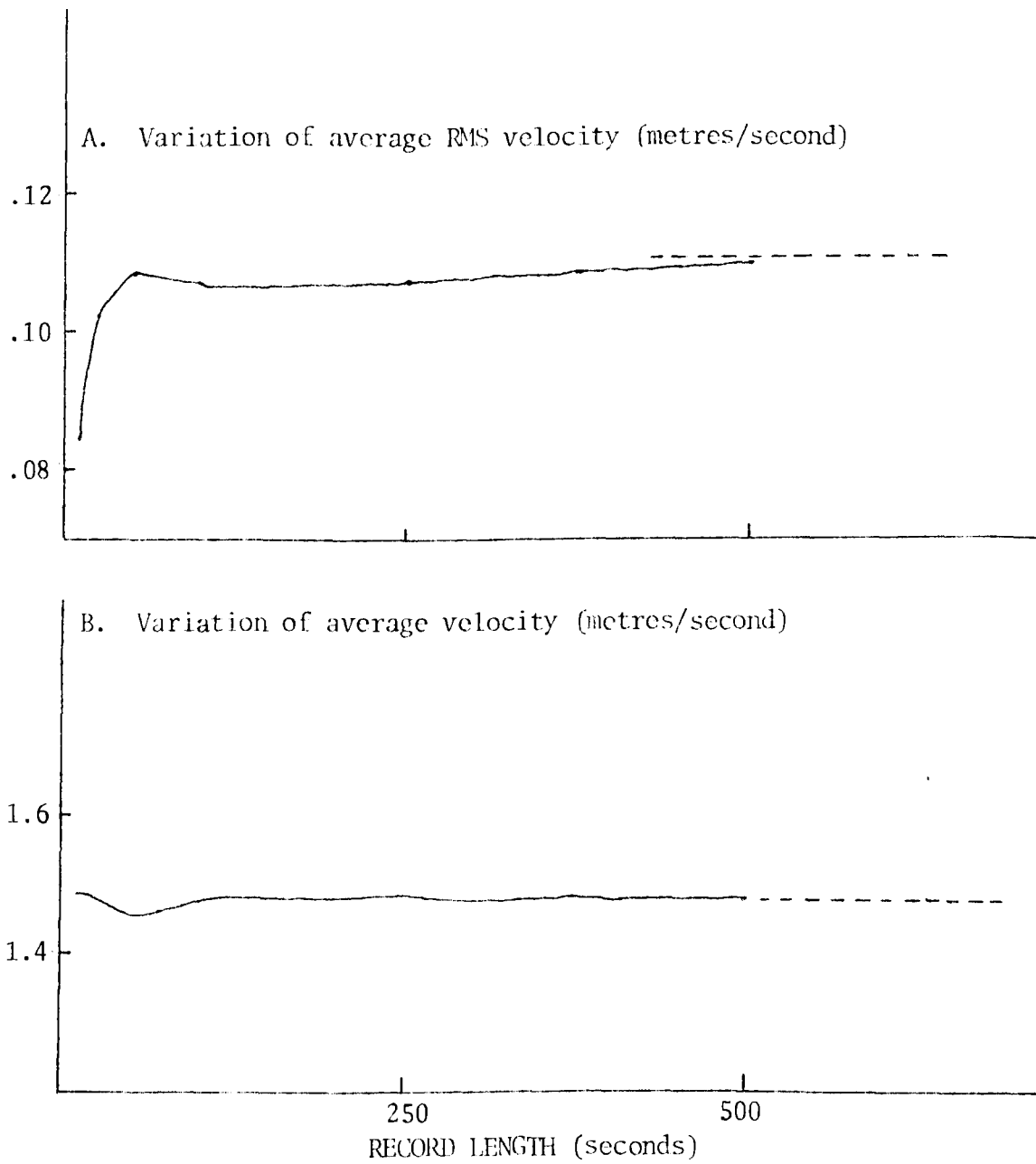


Figure 6.2: Effects of record length

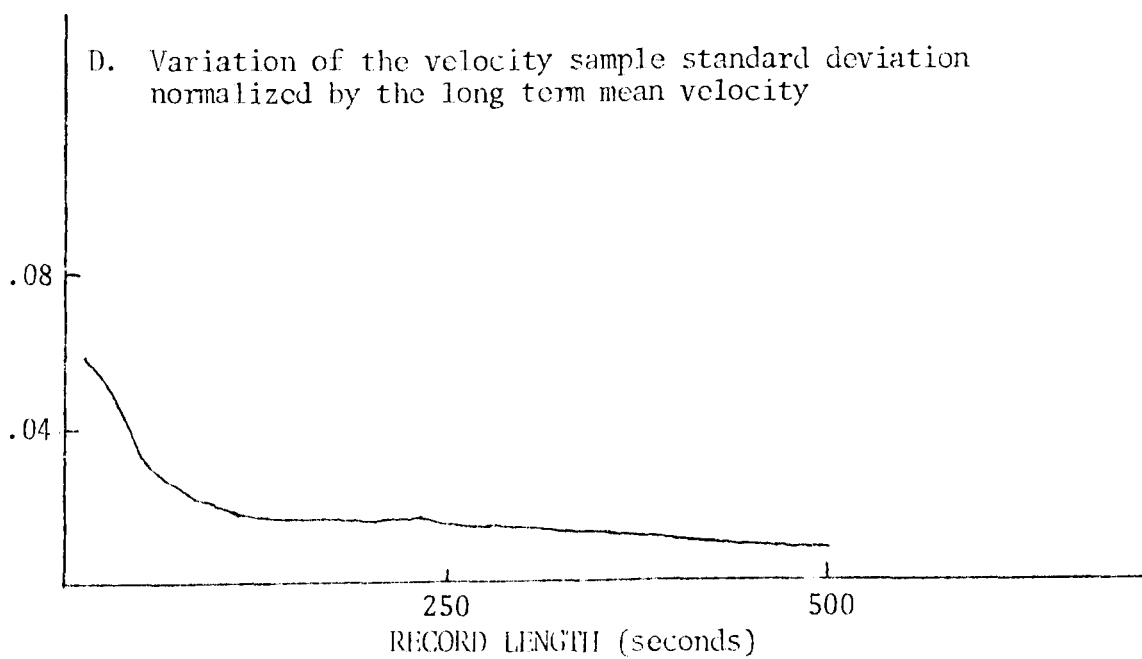
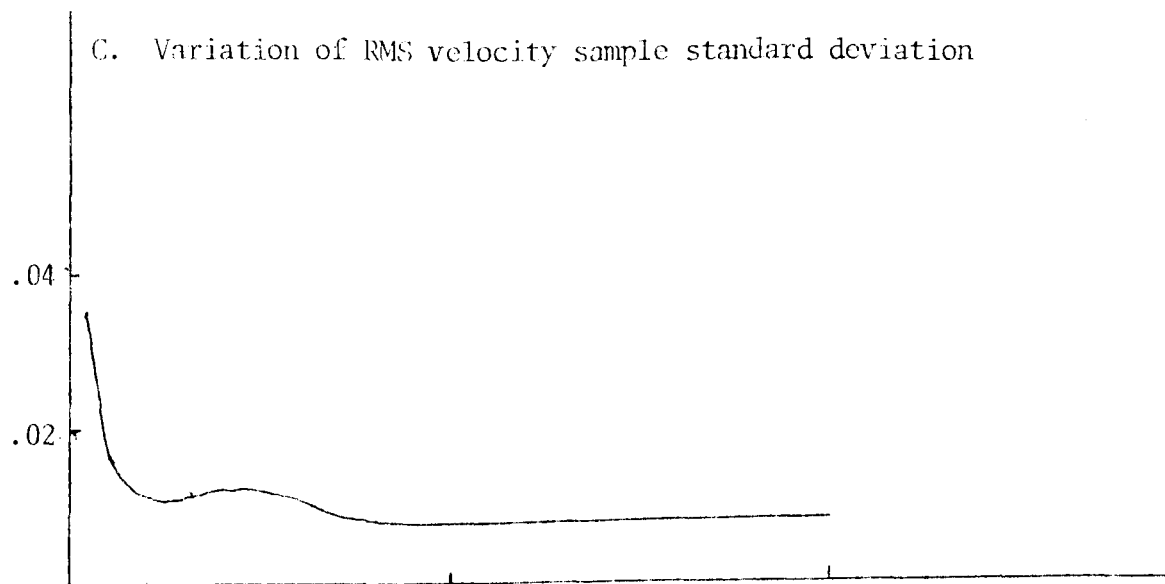


Figure 6.2 continued

continues to increase slowly to 500 seconds indicating the addition of small variance increments due to longer periods in the record. Figure 6.2A suggests that records exceeding 50 seconds are comparable in terms of RMS velocity, however, the sample standard deviation declines more slowly (Figure 6.2C; this may be considered as a likelihood of repeating the results of Figure 6.2A with one sample) to a constant value, suggesting record lengths should be extended to approximately 4 minutes to insure comparability (this value is equivalent to record lengths from McQuivey, 1973).

A few comments are worthwhile on a similar analysis of the accumulation of mean velocity and mean velocity sample standard deviation (Figures 6.1B and 6.1D). As Figure 6.1B demonstrates the average mean velocity is within one or two percent of the long term mean velocity for very small sample sizes and quickly stabilizes to a value equivalent to the long term mean. Figure 6.2D presents sample standard deviation results. The format and results are very similar to those of Demente'ev (1963), Savini and Bodhaine (1971) or Carter and Anderson (1963) where this type of analysis is used to indicate the length of record necessary to reduce the error in a velocity reading to less than some pre-specified percentage. The sample standard deviation (scaled by the long term mean) decreases rapidly and by 1.5 minutes is less than 2% of the mean. This result is equivalent to those of Demente'ev (1963) and Savini and Bodhaine (1971) for

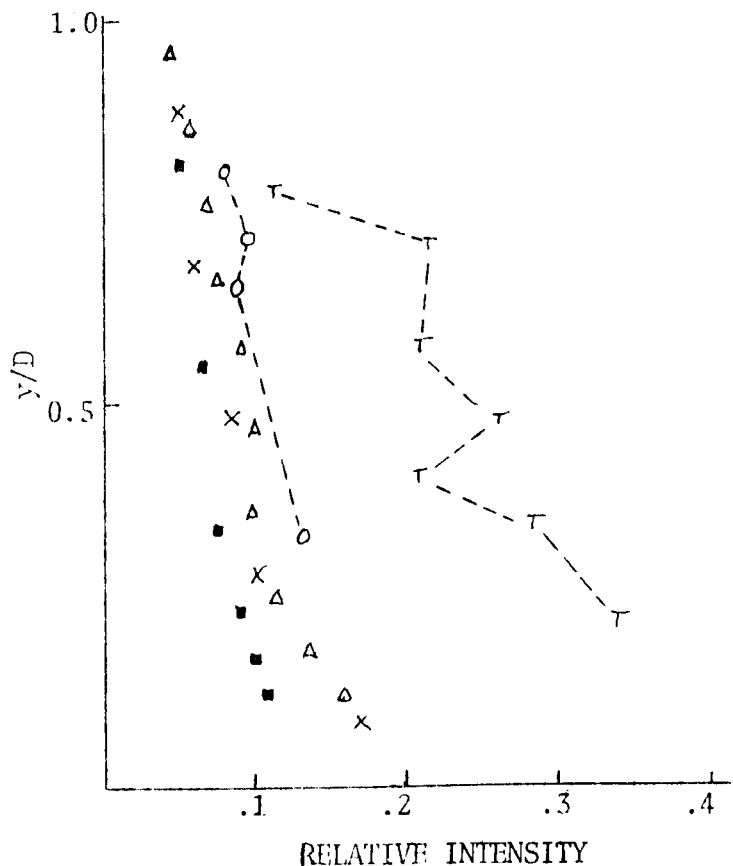
measurements near the surface.

The effect of variations in Reynolds no. and boundary roughness are more difficult to express. General results from Blinco and Partheneides (1971) indicate that for measurements at turbulent scales increases in relative intensity (when scaled by local mean velocity) occur for increasing Reynolds number though this effect decreases with increasing Reynolds number and is probably not important for the rivers under consideration. Second, relative intensity as a function of  $y/D$  increases strongly with bed roughness.

### Relative Intensity

Figure 6.3 presents relative intensity results from McQuivey (1973) and this thesis. Results from McQuivey (1973) and the Squamish River are very similar to other results from turbulent scale research (Laufer, 1953; Grinval'd, 1972) in flumes and rivers where the intensity shows a gradual increase from around 4% near the surface to around 15% near  $y/D = 0.1$ . Results from the four rivers plotted are similar in trend though values at any relative depth are scattered and results tend to spread more as the bed is approached. Data from the Squamish River plots to the right of the sandbed channel data possibly indicating the effects of increasing relative roughness.

Intensity results from the Alouette River are obviously wildly different from the other data in Figure 6.3 and by



- x Mississippi River: McQuivey, 1973
- Missouri River: "
- Δ Rio Grande Conveyance Channel: "
- T N. Alouette
- O Squamish

Figure 6.3: Relative Intensity against Relative Depth

$y/D=0.7$  are approximately twice as large. The explanation of these differences appears to lie in the local flow conditions. As was noted earlier (Chapter V; Figure 5.6) the measurement site was in a pool approximately 40 metres downstream of a riffle area (average channel bed slope to the measuring site was 1 in 65, though it is locally steeper near the measuring site). Similar intensity values were observed by Nowell (1975) under similar measurement conditions in a gravel bed river. His data varied from 10% near the surface to approximately 25% near the bed.

It is suggested that local intensity values are not in equilibrium with local conditions. It appears that the mean velocity stabilizes rapidly to local conditions and RMS velocities measured reflect large scale features imported from the upstream riffle area ( $5-10l$  is a life-distance estimate for large eddy features; where  $l$  is the length scale of this feature; hence some features may travel long distances downstream before losing their identity, Tennekes and Lumley, 1972; Roshko, 1976). Conclusions would be more appropriate if RMS values could be scaled by the shear velocity, which reflects local turbulence production (Blinco and Partheneides, 1971; Nowell, 1975).

### Standardized Velocity Distributions

A limited amount of comparative information is available for longitudinal velocity distributions in the outer layer. Data collected at similar measurement scales include only McLean and Smith (1979; over a sand wave field); while data collected in gravel bed rivers is limited to Nowell (1975; turbulent scale research).

Data are presented in two Figures. Figure 6.4 shows the vertical variation of skew and kurtosis.<sup>2</sup> Ensemble data from McLean and Smith are added to these figures since they represent data relatively close to the bed. While direct comparison at similar  $y/D$  values is not possible, sand and gravel bed rivers may be separated at these measurement scales by more indirect evidence.

The kurtosis values (Figure 6.4A) show a slight tendency to increase with relative depth. However, the values cluster around  $K=0$  (the normal distribution), though large deviations from this value occur that may be related to the length of the data series. This trend of increasing kurtosis with  $y/D$  is not supported by the data from McLean and Smith, though again it is unknown whether this different behavior is due to different bed or relative depth conditions. Relative normality of the fourth moment was found by Nowell (1975) for the longitudinal velocity

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<sup>2</sup>It is important to note that the number of samples per series is small for a stable determination of these values and the results should be treated with caution.



distribution (Cheekye River) and it is probably reasonable to assume that this is also appropriate for the data in this thesis.

A more obvious trend is observable for the skewness measures (Figure 6.4B), where values increase from negative skew near the surface. While the results from McLean and Smith are consistent with the trend of the data, it is not clear whether they are an adequate extension for skew values over gravel beds.

It is worthwhile to compare Figure 6.4B with skew results from Nowell (1975; Cheekye River), despite different measurement scales. While his data extends from  $n=0.01$  Hertz into the dissipation range, since it partly overlaps with data from this thesis it may be worthwhile to compare results. There is no other source of comparative data over gravel beds. His results indicated that skewness varied from -0.13 to -0.47 for  $y/D = 0.1$  through 0.9, respectively. These results are particularly interesting since the measurement at  $y/D$  was near or below the tops of the cobbles on the bed ( $D=0.7$  metres, average cobble size is 12 centimetres). Additionally, Nowell's data supports the declining trend in skewness measurements evident in Figure 6.4B.

While Nowell's results may be taken as indirect evidence supporting different skewness behavior at large measurement scales near the bed for sand and gravel rivers, other turbulent scale results differ. Grass' (1971) observations of skewness

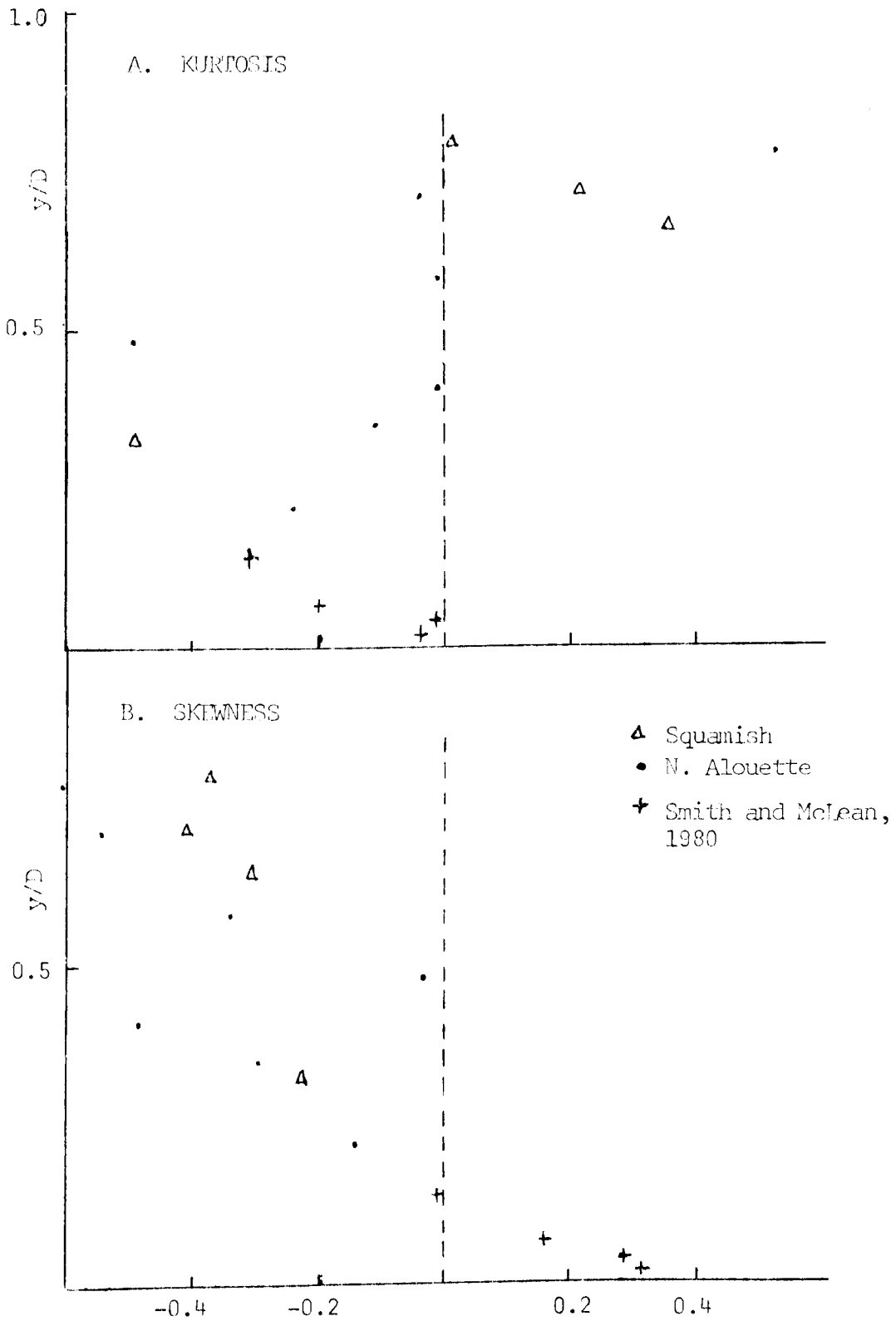


Figure 6.4: Kurtosis and Skewness as a function of Relative Depth.

over 9 mm gravel in a laboratory flume agree with those of Mclean and Smith (1979). These results may possibly be dismissed on the basis of low Reynold's Numbers and small sample sizes ( $N < 1000$ ).

It is worthwhile to discuss the nature of the skewness results from this thesis. First, negative skew is consistent with data collection problems induced by the meter suspension system used (as discussed in Chapter IV). The general effect of "swinging" and "near-stalling" should be to introduce occasional large negative excursions from the mean and, thus, negative skew. However, the trend of the skewness is inconsistent with this observation. The problem was observed mostly for low velocities (especially near the bed in the North Alouette River) and it would be expected that if meter effects dominated the distribution that the trend of skew would be reversed.<sup>3</sup> In the same vein, it is encouraging that similar results are recorded for the Squamish River where local mean velocities are at least a factor of two larger.

It is unclear, at present, what time scales are associated with the large negative velocities. In fact, the time scale and

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<sup>3</sup>It also could be argued that the range between the mean velocity and the minimum velocities is contracted by the presence of "zero" velocity and this effect keeps the distributions near the bed relatively normal. In other words, as mean velocities increase so does the range below the mean and thusly so does the negative skew. On the other hand, if the meter stability is related to the mean velocity, then the effect might be expected to decrease with increasing  $y/D$ .

distribution of these large negative velocities is probably more informative than the summary measures discussed so far. This is best approached through examination of the velocity traces.

A more detailed look at the velocity distributions is provided by Figure 6.5. The figure shows standardized velocity distributions both near the bed and the surface for the North Alouette and Squamish Rivers. All distributions are significantly non-Gaussian at  $p=.005$  (this result is based on a chi-square test outlined by Bendat and Piersol, 1968).

The principal feature of these distributions that is not apparent from the moments is their shape variation near the mean with relative depth. This variation is related to a slight excess of values near the mean at low  $y/D$  values. This slight excess is not reflected in the kurtosis values and is mostly concentrated into a narrow velocity range.

Velocity distributions from Nowell (1975) show opposite behavior with a slight lack of values near the mean very close to the bed. At larger measurement scales, in a sand bed river, McLean and Smith (1980) observed near Gaussian behavior very close to the bed. It is unclear how results from this thesis would vary relatively close to the boundary and thus, whether these other results are comparable.

An excess of values near the mean can be associated with intermittent flow phenomena though this is more commonly demonstrated by large kurtosis values of the  $u$ ,  $w$  or,

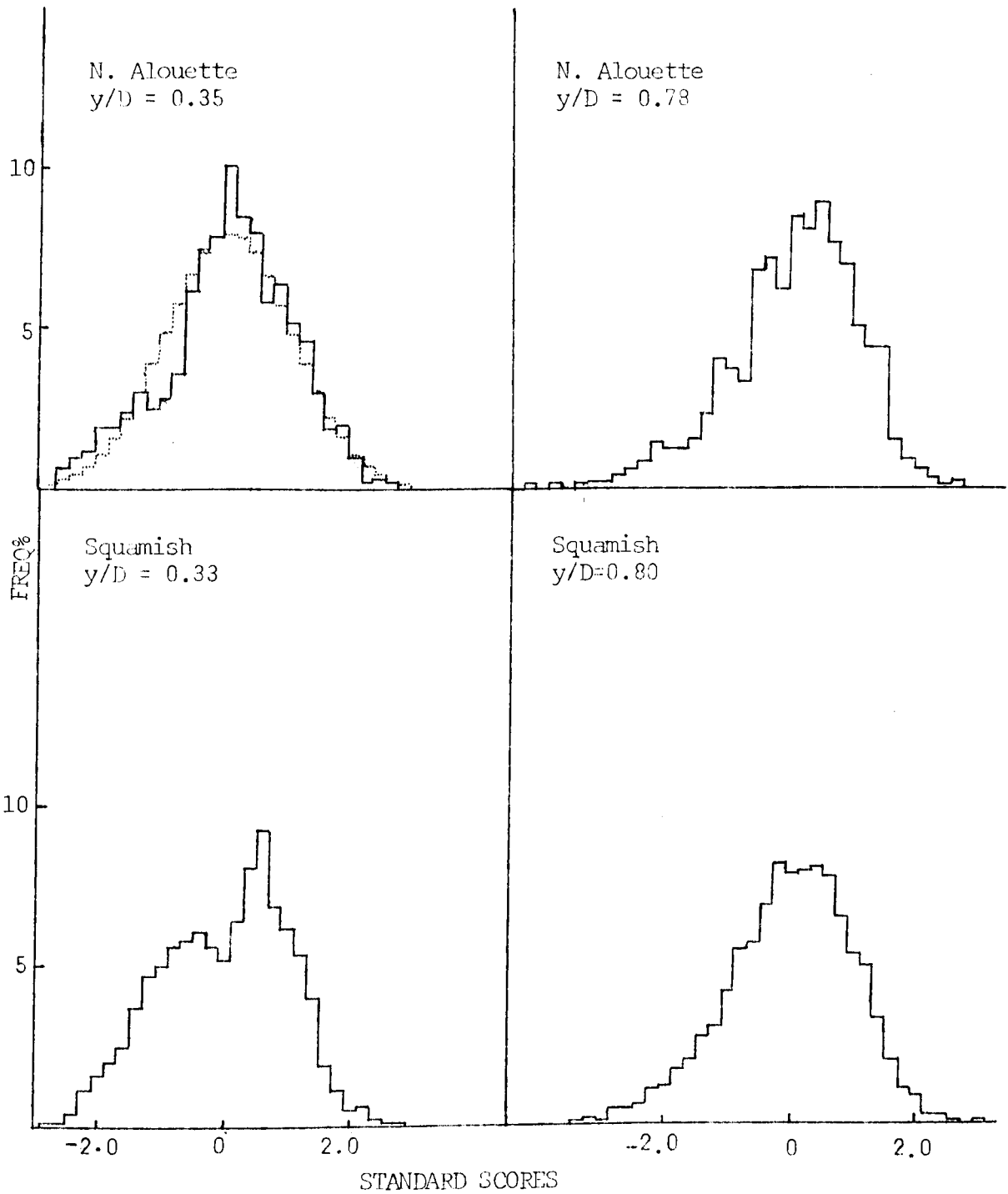


Figure 6.5: Standardized velocity distributions.

particularly the,  $uw$  distribution (e.g., McLean and Smith, 1979). Excess values in the longitudinal velocity distribution are typical of the type of intermittency that occurs with clear air turbulence (Dutton, 1969; this consists of patches of turbulence surrounded by patches of clear air). It is difficult to suggest that the observed excesses reflect intermittent behavior in the velocity records collected for this thesis since any intermittency would appear to be associated with only a small range of scales (see Chapter I) and must be seen against the background of a velocity field that is known to be continuous at other scales.

Consequently, it is more appealing to emphasize the concentration of these excess values (they also appear in some other velocity distributions) into a narrow spike. The presence of this concentration relatively near the bed suggests the diffusion of processes occurring at the bed, and dominated by a characteristic velocity, into the outer layer of the flow. This behavior might be expected from the presence of a wake layer associated with the gravel roughness elements.

It is worthwhile to directly compare results from the North Alouette and Squamish Rivers. As noted earlier, the local channel topography near the North Alouette measurement site appears to have influenced the intensity statistics. Despite the previously suggested non-equilibrium for the North Alouette, there is little difference between the higher moments or the

velocity distributions of the North Alouette and Squamish series (excluding the bimodality at  $y/D = 0.33$  in the Squamish). While it is not necessary that adjustment to channel changes apparent in the second moment be reflected in other statistics of the velocity distribution, the results suggest that the behavior of the intensity statistic more reflects changes in the local mean velocity than in the structure of the velocity fluctuations.

## VII. SERIAL APPROACHES

### Introduction

This chapter is the final approach to the velocity records collected for this thesis. In a detailed manner, this chapter examines:

1. comparison of velocity records in a vertical
2. comparison of records collected in one channel under different discharge conditions
3. cross-comparison of results from two rivers--the effect of "channel size"--and additionally comparison with other similar spectra

Two different methods will be used for these comparisons. Both spectral and rescaled range analysis will be applied to investigate the gross properties of the various data series.

It is worthwhile summarizing the approach of this thesis to comparison of the spectra. As has been pointed out, the phenomena that may exist at large scales show varying levels of "locality". There are events that are very local in origin and influence such as many of the features (bank eddies, bottom rollers) classified by Matthes (1947). Equally, shedding of separation eddies, as identified in this thesis, is dependent on



local conditions, and probably dependent on discharge conditions. The downchannel influence of these phenomena (i.e., the distance over which they retain their identity) is not known, though it is not expected to be large, implying adjustment may be rapid (see Chapter V). On the other hand, "pulsation" phenomena, if these existed in gravel rivers, would seem to be much less dependent on the location of the measurement site.

As a consequence of the above comments, spectral shape may be strongly dependent on conditions near the measurement site rather than any general channel properties. Comparisons of spectra between channels may then be dominated in some frequency bands by variation due to the relative measurement position (see Introduction to Chapter I), rather than the effects of "channel size". As such, the spectra from each river need to be investigated separately before comparison is made.

Most of the approaches to collapsing longitudinal low frequency spectra have come from atmospheric research. Their results tend to be discouraging. As Panofsky (1969, pp. 106-7; see his figure 4) states:

"It is clear that the shapes of spectra from various sites differ widely ... This suggests that mesoscale features determine the characteristics of the low frequency portion of the U spectra, not just the roughness length  $z$ , which is a measure of local roughness. Attempts have been made to fit U spectra to empirical formulas; however, such fits can be only of limited use for application over terrain for which they have been derived."

The quotation is interesting and the reasoning on "mesoscale roughness lengths" is best derived by analogy with turbulence research. On the microscale, roughness size and spacing control the statistical properties of the flow above the boundary (this control extends to the non-dimensional spectral peak; see Nowell, 1975). Reasoning by analogy and keeping in mind the hierarchy of skin through form roughness elements, it might be expected that different frequency bands of the spectrum may then scale with different elements of the roughness hierarchy (as was noted in Chapter I, some of these elements scale on channel dimensions). This is to say, if there is a hierarchy of roughness elements there equally well may be corresponding regions of the frequency continuum that scale in different ways.

An exploration of what a longitudinal mesoscale roughness element may be was begun in Chapter V. As was noted in that chapter, dominant wavelengths identified through spectral analysis are not sufficient to describe the effect of the roughness elements. However, on the microscale, spacing variation seems to be sufficient to produce peak separation in velocity spectra (Nowell, 1975) and, as such, dominant wavelengths may be useful for explanation in this chapter.

### North Alouette Speed Spectra

The common approach to presentation of low frequency spectra is to plot frequency times spectral density against either frequency or a non-dimensional frequency (i.e., Bullock et al, 1978; this is the usual form in atmospheric research; see Kaimal et al, 1972, for an example). This form of presentation will be followed in this chapter (see Chapter II for a discussion).

The most appropriate form of similarity scaling from atmospheric research is the Monin-Obukhov scaling. For neutral lapse rate conditions (this is the case for rivers) collapse should be obtained using the non-dimensional spectral density  $\hat{w}$  and the reduced frequency ( $f = ny/U$ ), where this is based on a frequency due to the height above the boundary.

This scaling applies only for  $y$  values in the "constant stress" layer where distance from the boundary is the only relevant length scale. This region excludes  $y$  values near the boundary and near the surface and also is limited to those scales above the frequency region of viscous domination (Monin and Yaqlom, 1975b). Consequently, it was of little surprise that Nowell (1975) found that this particular scaling was

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the square of the shear velocity is more appropriate, though less successful, for scaling than the variance, however, this was not measured. Additionally, this may be divided by a non-dimensional dissipation parameter to collapse the high frequency (inertial subrange) end of the velocity spectrum (see Kaimal et al, 1972).

inappropriate near the boundary and the surface. However, the variation of his results in a vertical, plus those of Perry and Abell (1975; pipe flow) and Bullock et al (1978; pipe flow), are useful for the purposes of comparison.

Speed spectra from the North Alouette River are presented in Figure 7.1. The most obvious feature of this figure is the general similarity of shape of all seven spectra. This is not entirely unexpected. As was noted in Chapter IV, the type of measurement apparatus limited the flow region that could be sampled to an area unaffected by nearness to the boundary (approximately 10 roughness lengths away in the Alouette) and also relatively distant from the surface. Results from Perry and Abell (1975) and Bullock et al (1978)<sup>2</sup> indicate that longitudinal spectral variation in a vertical is regionalized corresponding to inner/outer layer separation (Perry and Abell also include an area of "overlap"). While these results come from flows over smooth boundaries, somewhat similar results occur for flows over discrete roughness elements (Nowell, 1975).

While it is possible to distinguish these regions spectrally, there also appears to be consistent shape variation with  $y/D$  in the outer layer. First, there is a general tendency

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<sup>2</sup>Comparison must be of the most generalized form since similar quantities are not being measured or similar flow systems being sampled. Perry and Abell (1975) report results based on a circular frequency, while Bullock et al (1978) transform velocity data that has been rotated to correspond to the phase angle of wavefronts in the vertical. General agreement between these forms and Nowell's (1975) is encouraging.

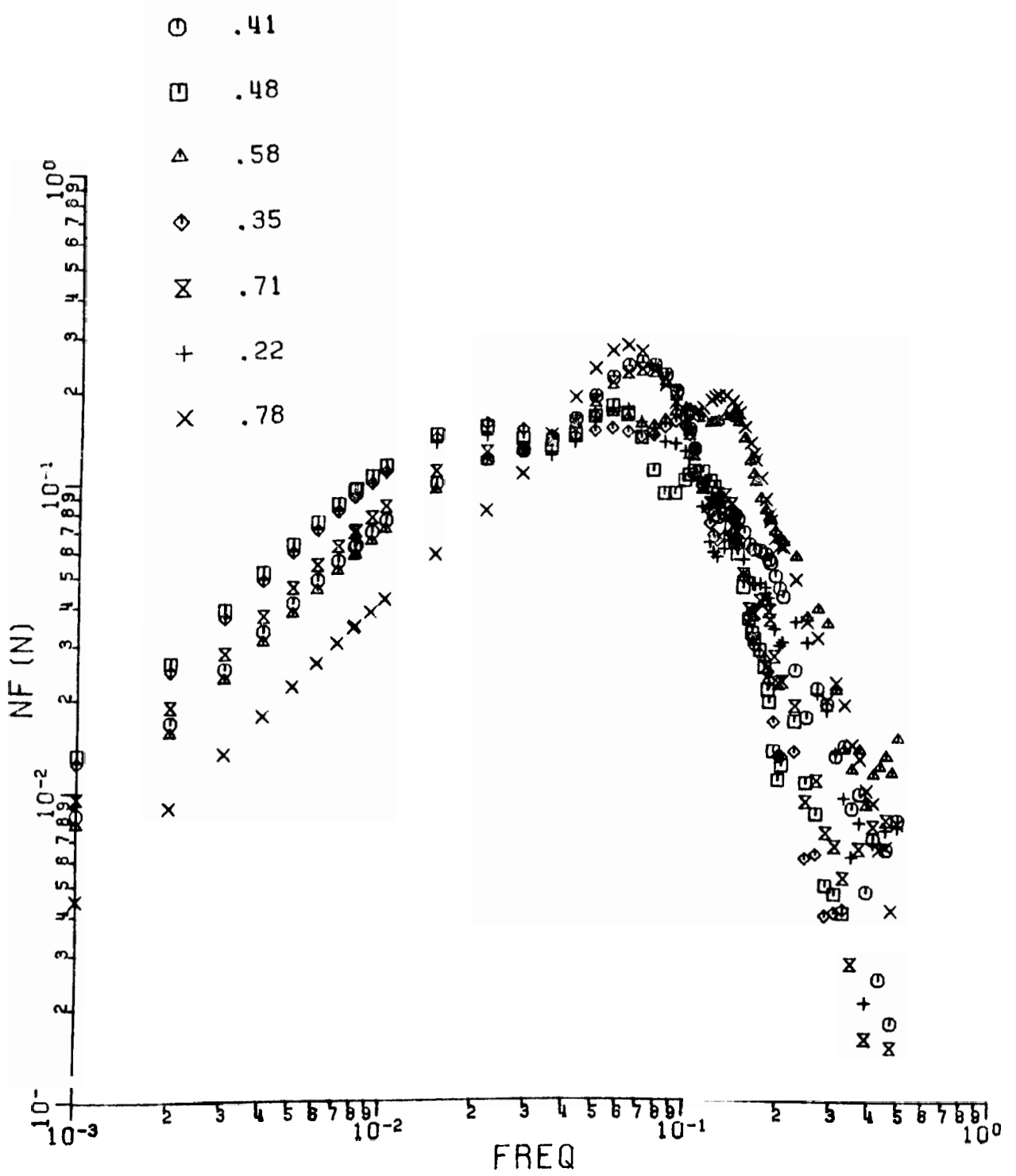


Figure 7.1: North Alouette speed spectra

for the spectra to become more peaked with increasing  $y/D$  (especially Perry and Abell, 1975 and Bullock et al, 1978; not really evident for Nowell, 1975). This tendency is associated with a low curvature or extensive flat region near the peak at relatively low  $y/D$  values.<sup>3</sup> Similar variation is observable in Figure 7.1 (see  $y/D = 0.22, 0.35$  and  $0.48$ ) however the flat region (or low frequency hump near  $n = 10^{**}-2$ ) extends to larger  $y/D$  values than might be expected on the basis of Perry and Abell's results. While there is a very general tendency to declining relative\* low frequency power with  $y/D$ , the only curve with different shape is that for  $y/D = 0.78$ . Differences between this spectrum and the rest are apparent in the cumulative spectra of the North Alouette from Figure 7.2A. This figure also gives a clearer idea of the relative importance of high and low frequencies in the spectra.

The change in the nature of the spectra at  $y/D = 0.78$  is consistent with Perry and Abell (1975) and Bullock et al (1978). The result is also consistent with the free surface spectra of Nowell (1975; his figure 26A) as evidenced by comparison of the

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<sup>3</sup>Under the spectral scaling used here this flat region corresponds to the traditional production (-1 slope) region. As was noted in Chapter II this region is most extensive near the boundary. Bullock et al indicate this flat region corresponds to the appearance of an increased low frequency component above the inner and buffer layers.

\*Since spectral density estimates are plotted each curve has the same area underneath it and ordinate values can be seen as percentages of the total variance.

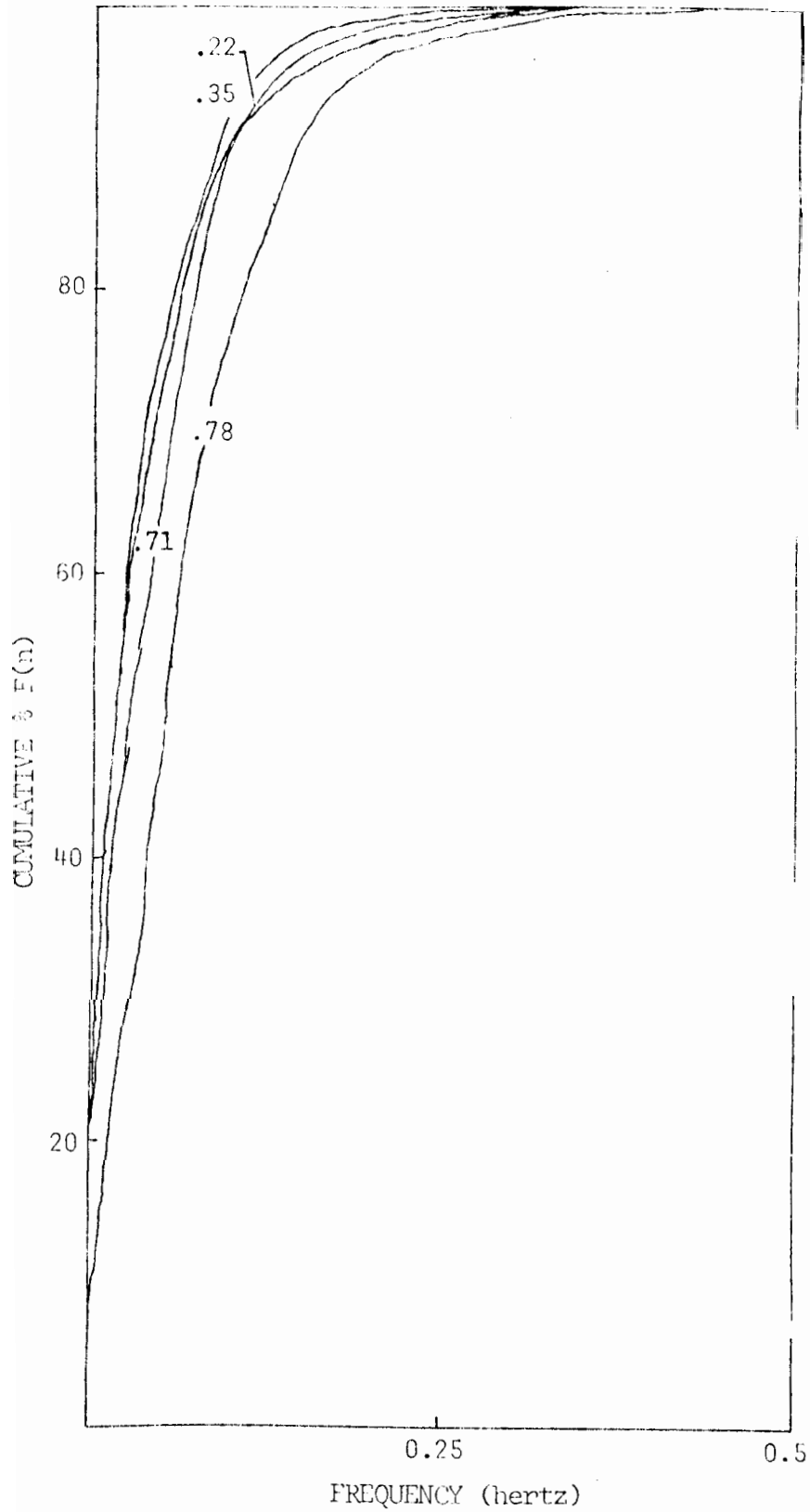


Figure 7.2A: Cumulative Spectra--N. Alouette

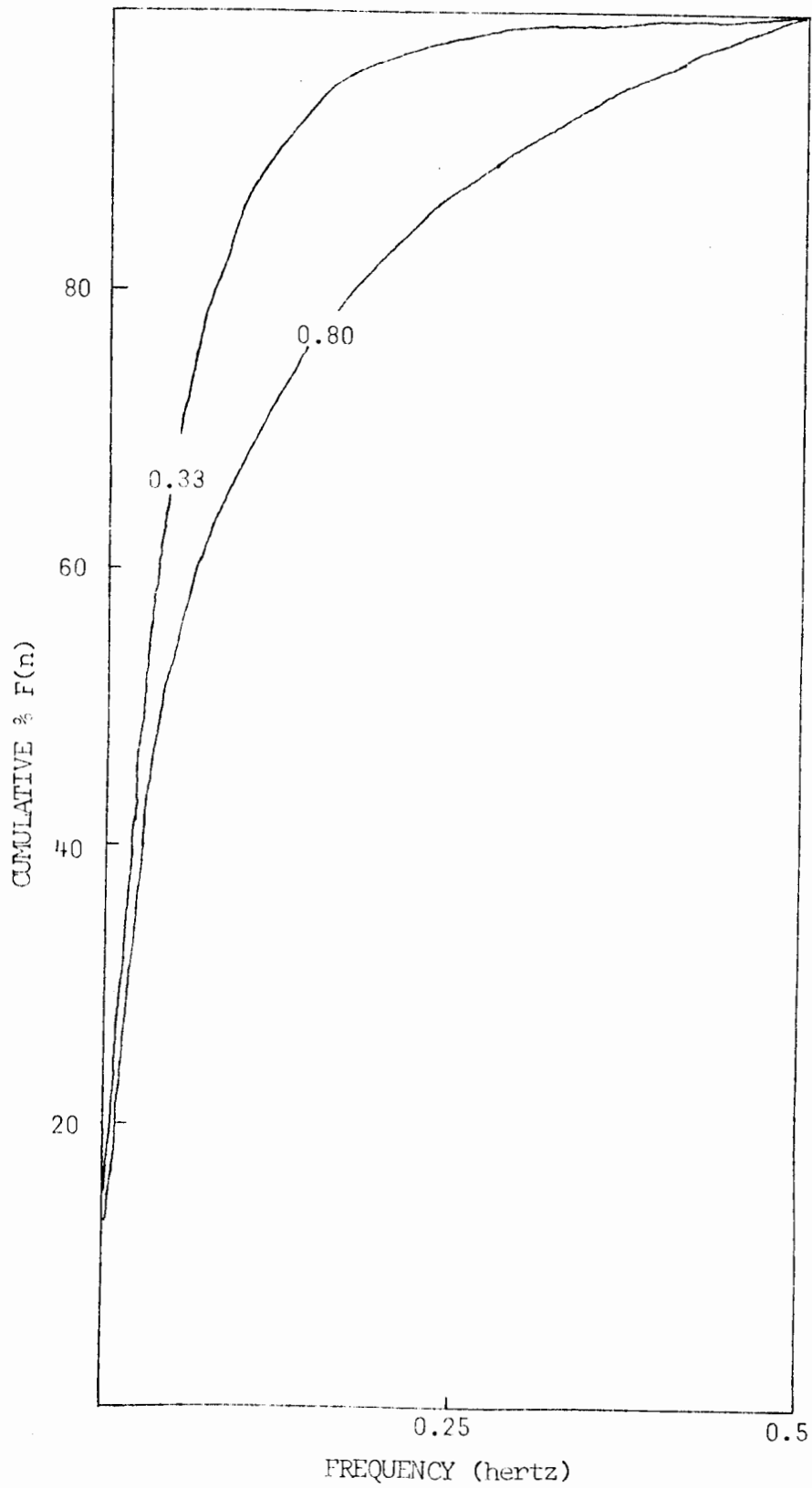


Figure 7.2B: Cumulative spectra--Squamish



cumulative curves, despite differences in "closeness" to the surface. The variation involves a relative shift of power to mid-frequencies. This change in the spectra with  $y/D$  will be discussed further in later sections of this chapter.

Similarly, there is variation in a vertical in the frequency band near  $n = 0.10$ . In this case, there appears to be a general tendency for energy to accumulate in this region with increasing  $y/D$ . Consequently, the largest peaks are observed for  $y/D = 0.78, 0.71$  and  $0.41$ .

It is unknown whether the two-component separation discussed in the last few paragraphs has physical significance or whether it is merely convenient for description. The results suggest that two frequency separated components which vary in their relative magnitudes may be necessary to explain the behavior of the low frequency spectrum in the North Alouette. At present, the system can be simply described as a lower frequency component dominant near the bottom of the outer layer with scales a small multiple of channel width, and a higher frequency component important closer to the surface with scales ranging between channel depth and width.

The fact that spectral shape varies with  $y/D$  and that peak location is relatively constant in frequency over changing  $y/D$  implies that reduced frequency scaling is unlikely to improve the collapse of the data. The principal effect of this scaling (see Figure 7.3A and B) is to provide a general increase in the

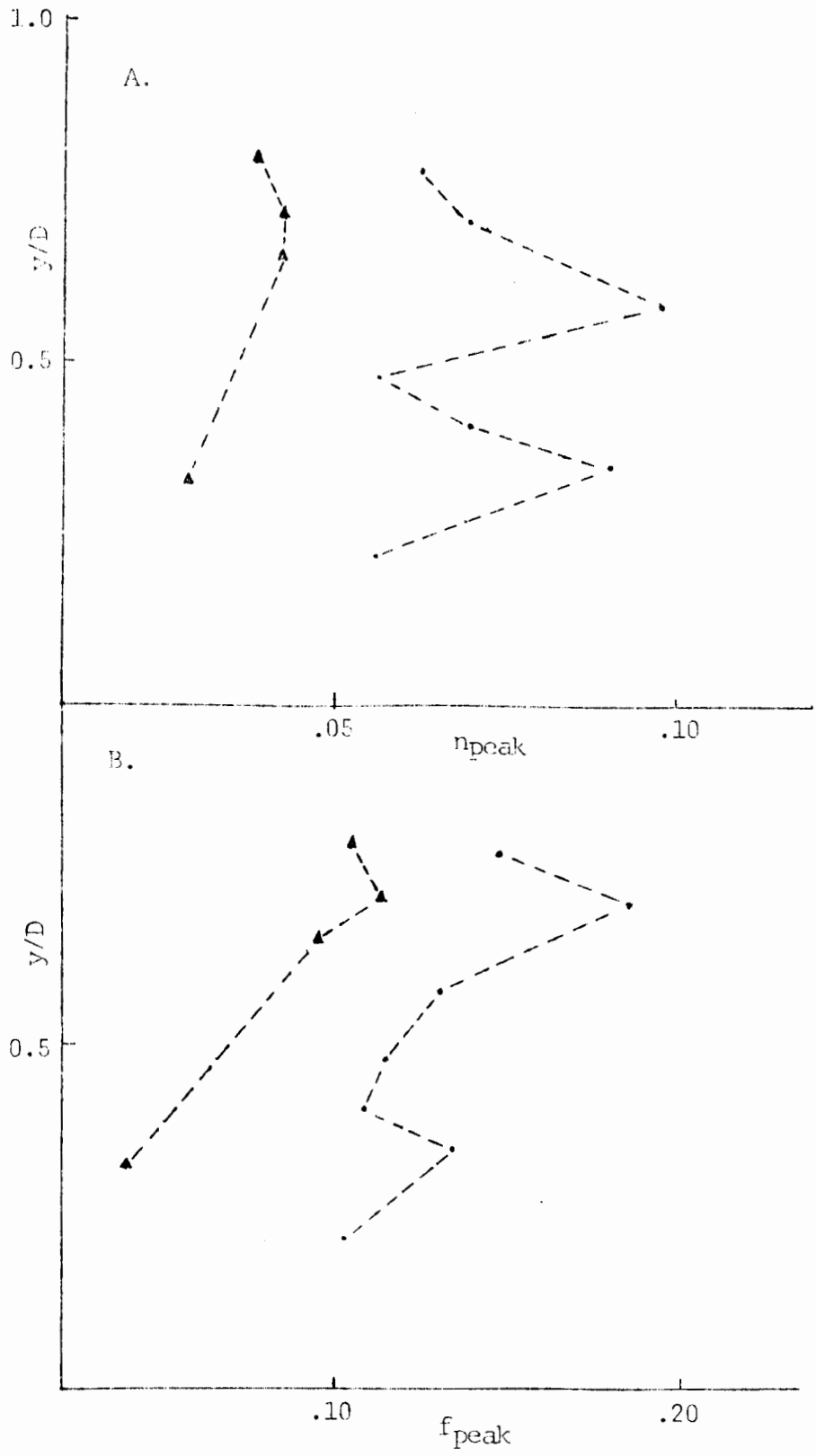


Figure 7.3: Peak frequency and peak reduced frequency against relative depth.

peak reduced frequency value with increasing  $y/D$ . In other words, the measured peak scale does not increase as fast as the scale associated with distance from the boundary. This is a fairly common effect of this scaling (see Nowell, 1975; Perry and Abell, 1975). The actual change in these peak values is not large compared to variations observed by others (often a decade or more); though much of this observed total variation occurs between inner and buffer layer spectra and outer layer ones. It is very difficult to choose peak values for some of the spectra since many are flat in the region of the peak. Consequently, simply the greatest value was chosen which leads to large scatter in Figure 7.3A. Hence, frequency peaks for  $y/D = 0.58$  and  $0.35$  extend to larger frequency values than the other spectra. It is also interesting to note that a minor peak of  $y/D=0.78^5$  extends out to the same higher frequency and that these three spectra extend their peak region out to slightly higher frequencies than the remainder of the spectra. The explanation for this behavior appears to be that these three spectra were collected at higher discharges. It is extremely interesting that the spectral results are sensitive to such minor changes in discharge. The result has implications for the interaction of bed materials and flow with changing discharge. Again, results from the Squamish River where a larger range of

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5 The gap between these two peaks is not significant in terms of confidence intervals.

discharges were observed will be more conclusive.

#### Comparison with Squamish River Speed Spectra

Figure 7.4 presents the speed spectra of the Squamish River. Several results are apparent from this figure. First, the shape is simpler in form than that of the North Alouette and there is no indication of the type of "two-component" spectral shape discussed for the Alouette.\* Second, the spectra near the surface are not different in shape from the one near the bed (this comment applies to the lower and mid-frequency parts of the spectra). However, the relative contribution by low frequencies declines as the surface is approached, similar to the North Alouette (see Figure 7.2B). Again, differences occur because of differences in the high frequency end which appears to be controlled by changes in discharge. Third, there is large variation in slope at the high frequency end as discharge changes. These points will be discussed in turn.

As is evident from Figure 7.3, the maximum peaks observed for the Squamish River occur at much lower frequencies than those of the North Alouette. While this might be attributed to the effect of channel "size", a direct comparison of some of the North Alouette and Squamish spectra (see Figure 7.5)

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\*Since the series from the Squamish are longer and were calculated for the same resolution bandwidth they should be smoother since the degrees of freedom (approximately  $2N/m$ ) are much larger.

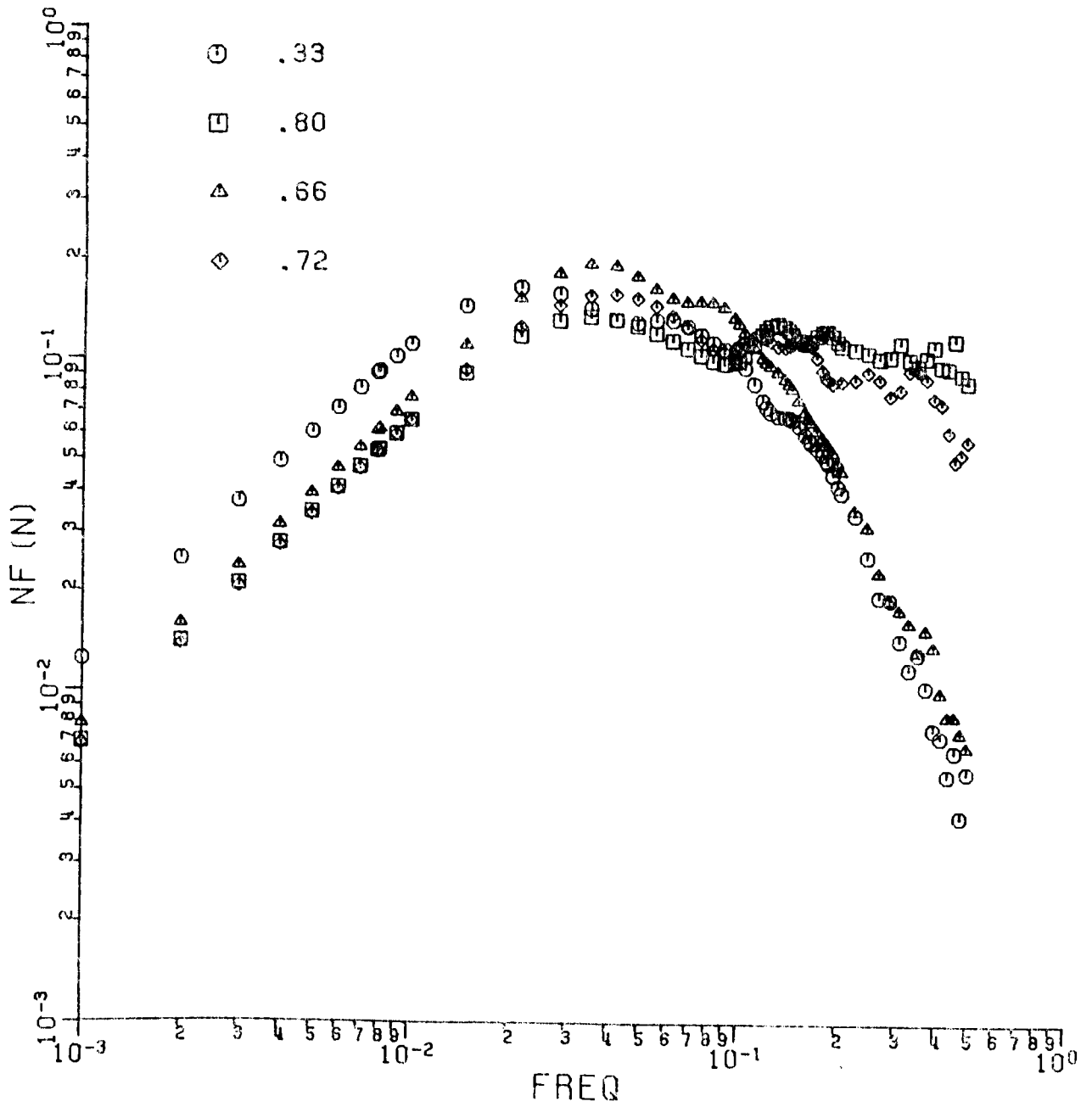


Figure 7.4: Squamish River speed spectra

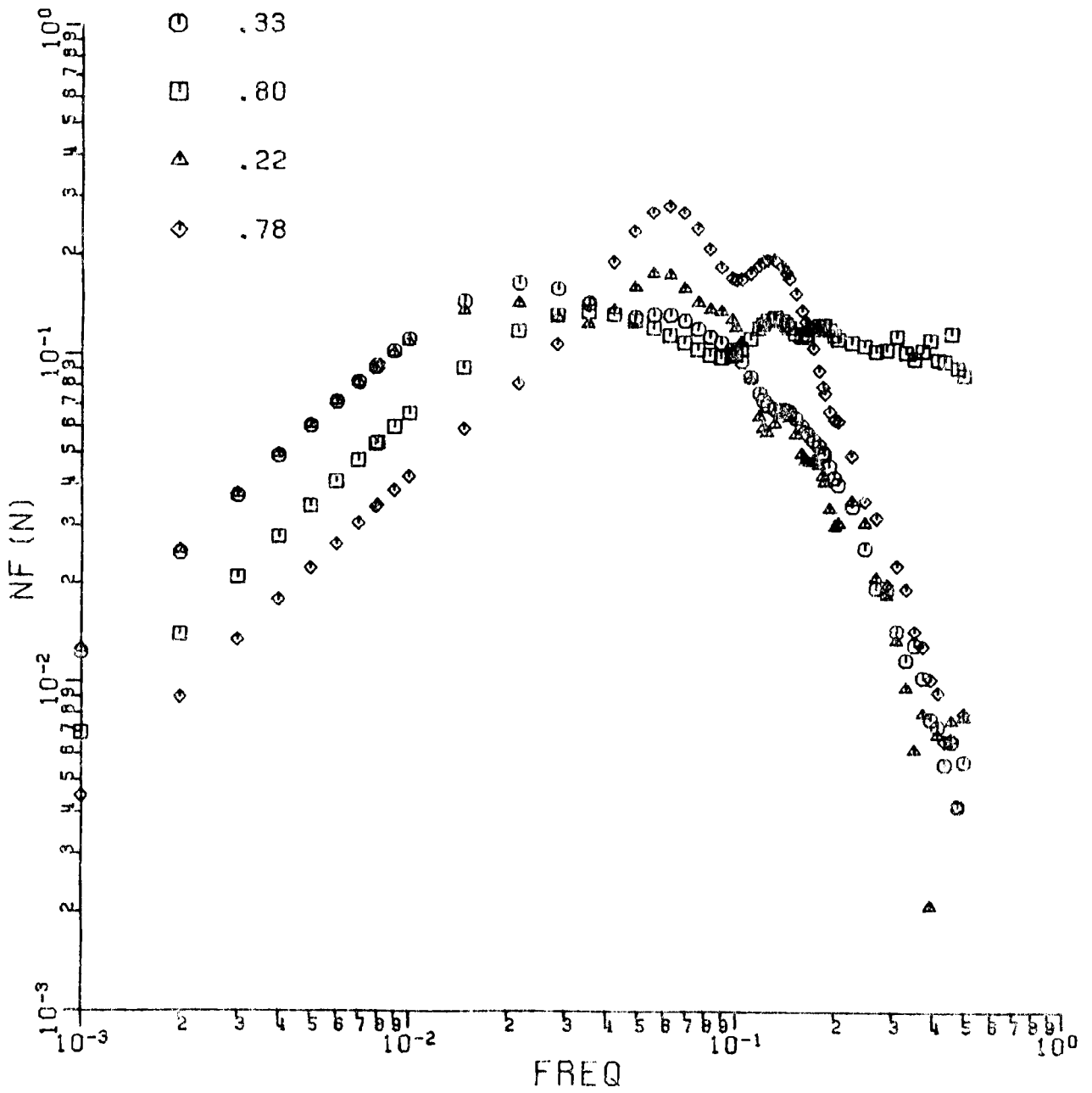


Figure 7.5: Comparison of some Squamish and North Alouette speed spectra

demonstrates that it is due to the different behavior of the spectra; especially evident when the two "near-bed" spectra ( $y/D = 0.22$  for the North Alouette and  $y/D = 0.33$  for the Squamish River) are compared. While the distributions are very similar at the low frequency end, including a "hump" near .01 Hertz, and at high frequencies, the peak occurring near frequencies of 0.1 Hertz is missing from the Squamish spectrum. The lack of this spectral component shifts the maximum peaks to lower frequencies for the Squamish River.

In fact, the similarities of the spectra of the two rivers, under this form of presentation, are unexpected. Judging from Figure 7.5 the variation in the relative distribution of variance over frequency is dominated more by relative depth within a river than between channel differences at similar relative depths.

To this point spectral variations have been discussed outside the framework of statistical similarity or difference. Consequently, before proceeding to discussion, the spectra were compared using a Test for Equivalence from Bendat and Piersol (1966; also mentioned in Jenkins, 1961). The test statistic,  $C$ , is based on the fact that the sampling properties of the logarithm of the spectral density are constant over the frequency interval of interest.  $C$  is of the form;

$$C = [ N(n) \cdot (2/dn(1) + 2/dn(2)) ] \log \sum_{i=1}^{N(n)} F_1(n) / F_2(n)$$

and the statistic has a standardized normal distribution.<sup>7</sup>

This statistical comparison was done for two reasons; first, to investigate whether the  $N(f)$  presentation has dominated differences between the spectra and second, if possible, to simplify the discussion of spectral differences by separating the spectra into groups.

Results of applying this test (at  $p = 0.01$ ; results would differ at higher confidence intervals) indicate that there are three groups that are internally homogenous but dissimilar to other groups. These are:

1. North Alouette,  $\gamma/D = 0.22, 0.35, 0.41, 0.48, 0.58$  and  $0.71$ ;  
Squamish,  $\gamma/D = 0.33, 0.66$ .
2. North Alouette,  $\gamma/D = 0.78$ .
3. Squamish,  $\gamma/D = 0.72, 0.80$ .

As a result, some of the spectral properties used in the last section are not discussable, especially the tendency to concentrate variance in the frequency band near 0.10 Hertz as  $\gamma/D$  increases. As noted earlier, though,  $\gamma/D = 0.22$  and  $0.71$  would be significantly different at a marginally increased

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<sup>7</sup> Additionally,  $N(f)$  is the number of independent spectral estimates and is equal to:

$$N(f) = B/B(e)$$

where  $B$  is the bandwidth of interest and  $B(e)$  is  $1/2m$  (Equivalent equi-variability bandwidth; from Jenkins, 1961), where this is defined for the particular window function used. Then  $df(1)$  and  $df(2)$  are the Equivalent degrees of freedom (Jenkins, 1961) and are equal to  $2N/m$ , where  $m$  is the truncation. Note that the test is applied to the spectral density estimates; this prevents rejection on the basis of total variance differences.



significance level. On the other hand, the 99.0% level seems to be appropriate for the type of gross comparisons and simplifications made in this section.

The most obvious and intriguing aspect of this classification is that it contradicts previous expectations. It has been argued (Chapter I) that a reasonable expectation was that energy would be skewed to lower frequencies (larger scales) as channel size increased, using an argument based on limiting dimensions (a slightly more sophisticated version of this was argued in the introduction to this chapter based on different frequency bands responding to different elements of a channel form hierarchy). Additionally, it was argued that it might be reasonable to expect local measurement site differences (see Chapter V) to dominate spectral comparisons. The similarity between the bulk of the North Alouette and two of the Squamish spectra (discharge variation is also operating here which complicates comparison) implies that the above arguments are unreasonable. This has very interesting implications for the understanding of low frequencies in gravel rivers, which will be discussed later.

It is important to discuss why differences occur between  $y/D = 0.72$ ,  $0.80$ ,  $y/D = 0.78$  and the rest of the observations. It appears that differences related to  $y/D = 0.80$  and  $0.72$  are a direct function of increases in discharge. The variation in these two spectra is consistent with the extension of the flat

peak region observed for the North Alouette spectra under discharge change. Also, it is apparent from the way that the C statistic accumulates that differences occur principally in the high frequency end of the spectra.

As mentioned earlier, the extension of this flat region is equivalent to an expanding  $-1$  slope region. Extensive  $-1$  slope regions can be detected in Mclean and Smith (1980) for similar frequency ranges. However, their results are from much lower relative depths. This  $-1$  slope region appears to extend only into higher frequencies as discharge increases; lower and mid-parts appear to experience little or no change. While, unfortunately, no comparative measurements are available close to the bed, the results are consistent with the idea that with increasing discharge the flow/rough boundary interaction tends to concentrate energy through vorticity amplification into a dominant frequency band (Nowell, 1975). This statement should be treated with extreme caution since it is unclear whether these processes would be diffused throughout the boundary layer.

It is somewhat more difficult to discuss the difference between the observation at  $y/D = 0.78$  and the bulk of the results. In this case, differences appear primarily as a relative lack of variance at low frequencies and a relative excess at mid frequencies. This result is difficult to treat since no similar variation occurs in the vertical in the Squamish River despite similar  $y/D$  values (i.e., the variation

is not related to "closeness" to the surface). Though, the two vertical structures are difficult to compare due to differing discharges and vertical locations. Second, the behavior of the spectrum at  $y/D = 0.78$  appears to be consistent with the vertical variation of the remainder of the North Alouette spectra.

If on the basis of comparison with the Squamish spectra it is argued that this variation is not a property of the equilibrium vertical structure of the spectrum (this is based on the fact that there is little channel variation above the Squamish River measuring site and that the Squamish spectra are similar to single peak low frequency atmospheric spectra); then the explanation is probably due to channel behavior near the measurement site. The vertical structure of the N. Alouette spectra is then seen to be a result of changes in flow structure induced by changing channel conditions. In other words, the two component spectral type discussed earlier represents a combination of scales imported from upstream and a diffusing upwards local equilibrium spectral shape (see also intensity statistics; Figure 6.3). In this scheme, it is assumed that the low frequency "hump" represents scales in response to very local depth increases and the variance concentration near 0.10 Hertz, which declines in importance with decreasing relative depth, scales imported from the upstream riffle area.

### Comparison with Other Spectra

Spectra from the North Alouette and Squamish Rivers were compared with an empirical curve from atmospheric research (Kaimal et al, 1973) and one spectrum from a flume over discrete roughness elements (Nowell, 1975; his figure 24b,  $y/D=0.36$ ). While other low frequency spectra are available, (see Chapter III) generally either it was impossible to extract reasonable spectral estimates from very small, poorly marked and highly variable graphs (esp. McLean and Smith, 1979) or vital information (e.g., variance estimates for converting to spectral densities; Yokosi, 1967) was missing from the paper.

Kaimal et al (1973) have produced empirical curves, covering various spectral parameters, for neutral lapse rates. For longitudinal spectra;

$$nf(n)/u^*2 = 105f/(1 + 33f)^{5/3}$$

This curve produces a  $-2/3$  slope in the inertial subrange (approximately  $f > 10$ ) and tends asymptotically to a  $+1$  slope at small  $f$  values.

Since the shear velocity was not determined for either the North Alouette or Squamish Rivers, the empirical curve from Kaimal et al was adjusted so that spectral estimates at the peak corresponded to that from  $y/D = 0.80$ . The general approach taken to comparing spectral shape is usually some form of peak matching. The approach taken in this thesis follows Nowell (1975). In this case  $nF(n)$  is plotted against  $ny(0)/U$ , where

$y(0)$  is a length scale chosen so that

$$n(\text{peak})y(0)/U = 0.1$$

The result of plotting these two variables is seen in Figures 7.6 and 7.7. It is apparent from these two figures that the Squamish spectra agree reasonably well with the two comparative spectra while the North Alouette results do not. Spectra, particularly the North Alouette, tend to underestimate the comparative curves at large non-dimensional frequency values.

However, it is felt that the method of peak matching used may not be an appropriate way to compare these spectra. Since, as discussed previously, the North Alouette spectra appear to have two peak components, comparison based on the absolute maximum peak implies that different parts of the North Alouette spectra are matched with the peaks of the comparative spectra relative to the Squamish river spectra. This matching of different parts helps to explain why the fit appears much better for the Squamish spectra.

The Alouette spectra were replotted (Figure 7.8) so that the low frequency component was matched with the other maximum peaks at  $ny(0)/U = 0.1$ . It is obvious that this change improves the fit to the comparative curves.\* Several features remain worth discussing, though. First, the spectra from this thesis

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\*Note that  $nF(n)$  will tend to be underestimated for the Squamish and North Alouette series relative to the other series, since the frequency range 0.001 to 0.5 Hertz only includes a proportion of the total variance (see Figure 7.2A and B), whereas this frequency band is scaled by the total variance.

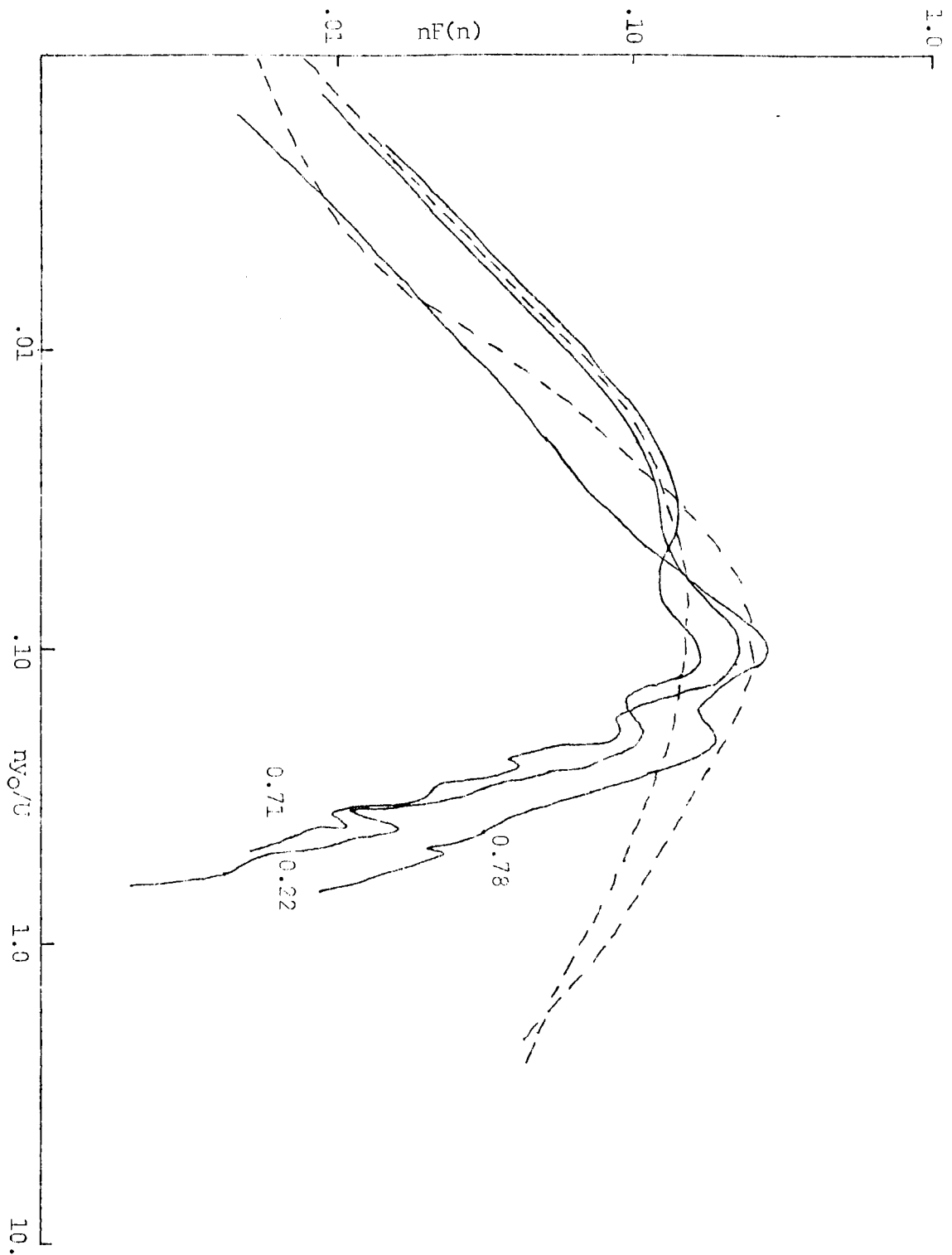


Figure 7.6: Shape comparison of N. Alouette spectra  
 (see Figure 7.8 for dashed lines)

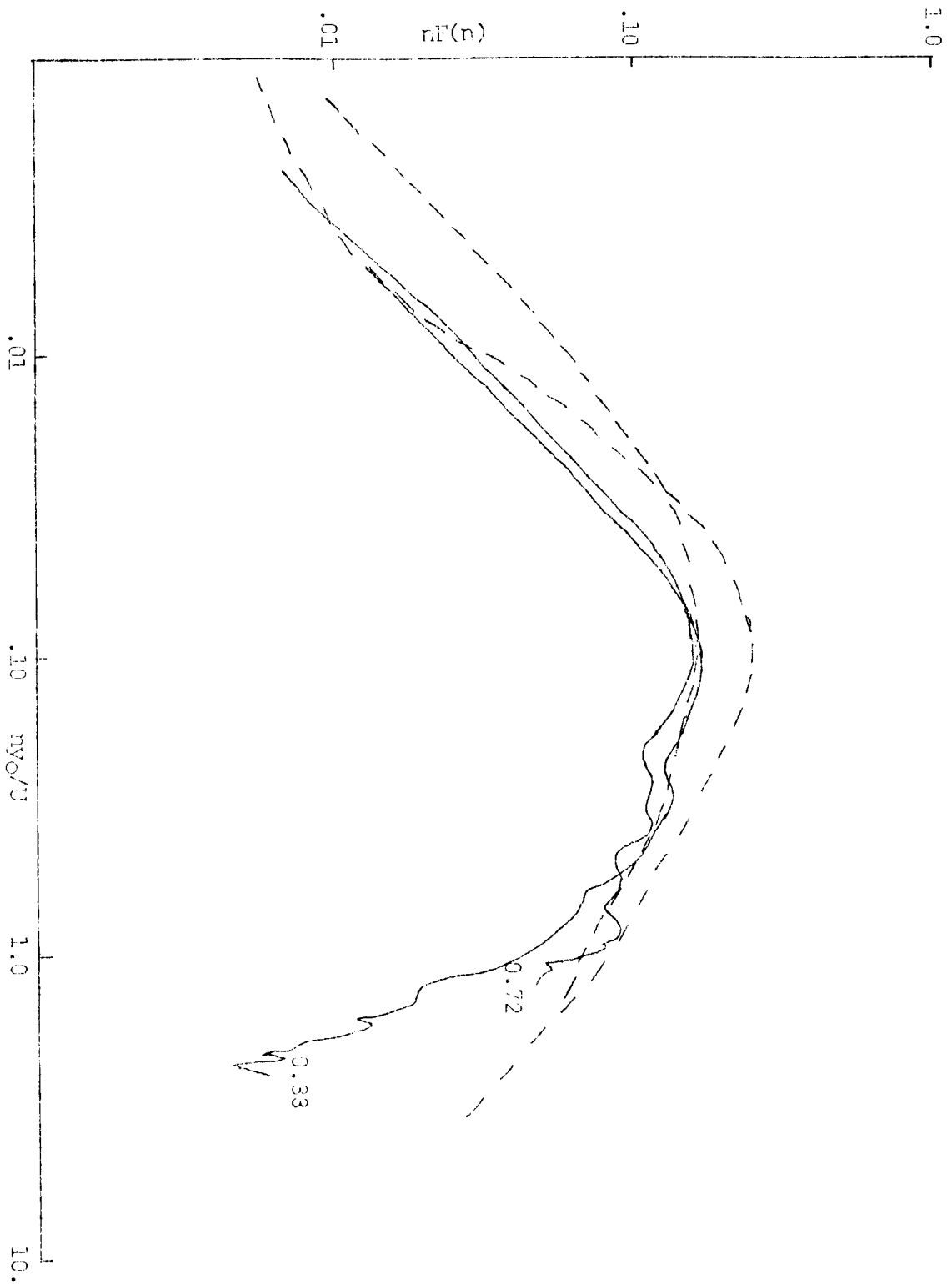


Figure 7.7: Shape Comparison of Squamish River spectra.  
 (see Figure 7.8 for dashed lines)

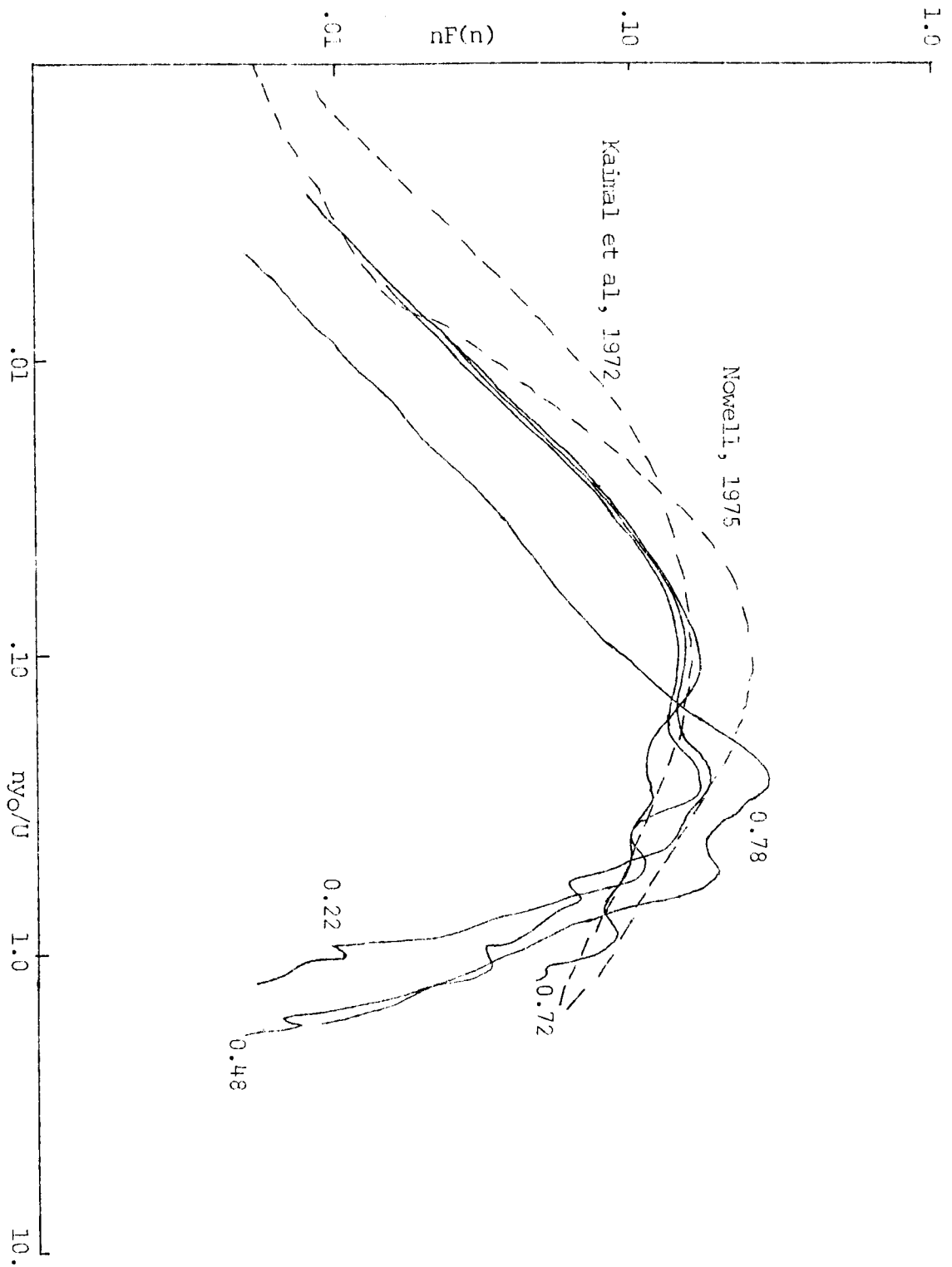


Figure 7.8: Comparison of spectral shape using lowest frequency peak.



(and Nowell, 1975) tend to underestimate the low frequency content relative to Kaimal et al's curve. Second, the comparative curves tend to be overestimated by the Alouette spectra in the mid to high frequency band. Third, the highest frequency part of the Alouette and some of the Squamish spectra appears to be severely underestimated. Fourth, the Alouette spectrum at  $y/D = 0.78$  appears to be a singularity.

It was felt, at first inspection of these figures, that the problem at the highest frequency end was related to meter instability problems. However, it appears more likely that the underestimation is due to subjective decisions taken while digitizing the original velocity traces. This opinion is partly based on arguments outlined in the last section of Chapter VI.

This is best illustrated by discussing the Squamish River data. Of the original four velocity traces, those collected on August 25, 1978 were unfiltered and those of October 3, 1978 were filtered. The charts were originally digitized at UBC; however, inspection of the data indicated that numerous data points were "skipped" in each series. Before this was discovered, spectral estimates of these four series showed that while slope variation in the high frequency end occurred that was qualitatively similar to that observed in Figure 7.4,

differences were not as large.<sup>9</sup> Subsequently, the series collected on October 3 were redigitized since the number of missing data points was large, while the series of August 25 were filled where missing data occurred with a value that was the average of velocity values prior to and after the missing point. When spectral estimates were recalculated for the new series it was found that the high frequency estimates had decreased (and slope had become more negative) for the series of August 25 and declined more dramatically for those of October 3. It is felt that the large observed change is due mostly to attempts during redigitizing to exclude frequencies less than the digitizing interval (Note--despite being filtered some of these frequencies are passed).

This lack of confidence in the highest frequency part of the data indicates that conclusions about the effect of discharge changes on spectral shape must be treated with extreme caution.

### Rescaled Range Analysis

The results of the rescaled range analysis are presented in Figures 7.9 and 7.10. These are not presented in the usual form of a box diagram (see Mandelbrot and Wallis, 1969) where all estimates at a particular sample size are plotted. It was felt

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<sup>9</sup> The obvious effect of missing data points is to shift variance to higher frequencies.

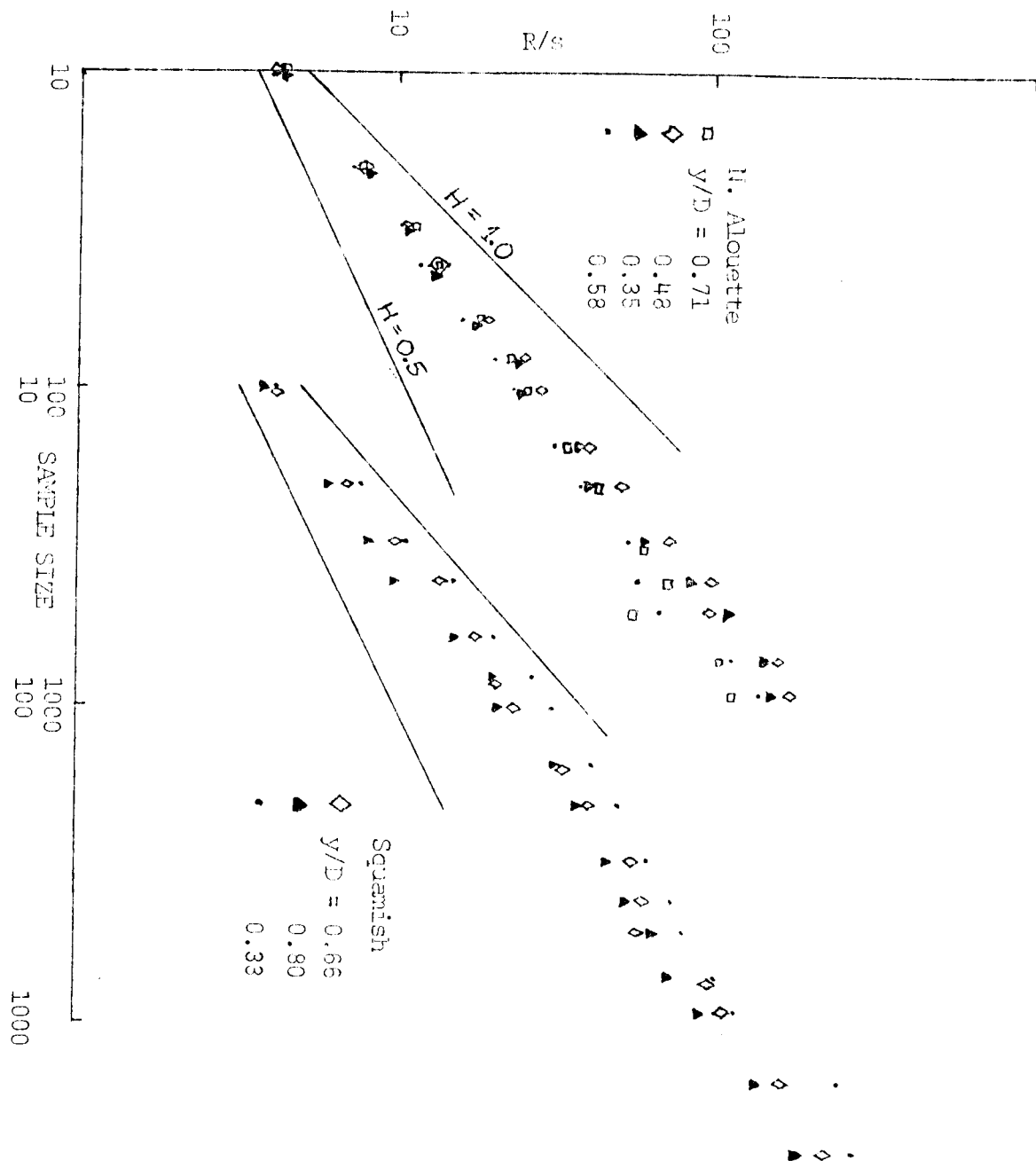


Figure 7.9: Rescaled Range Plots.

that this complicated the two figures to the point where the selected curves were indistinguishable. However, it is important to give some idea of the variability associated with the plotted mean values. Where 10 estimates were made for a particular sample size ( $10 < S < 100$  for the N. Alouette River and  $10 < S < 300$  for the Squamish River) standard deviations and maximum and minimum values were determined. Results indicated that standard deviations were generally of the order of 15-30% of the mean, while maximum and minimum values ranged to 50% of the mean value.<sup>10</sup>

As was discussed in Chapter I, it is particularly difficult to interpret rescaled range results, since the principal feature of the diagram (the slope) has no specific meaning, which is to say, there is no specific meaning attached to two series having different slopes.<sup>11</sup> Consequently, emphasis has been placed on detecting gaps or slope breaks in the diagrams. This will be

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<sup>10</sup>The technique is not designed to investigate internal variation in the record. Averaging at each sample size implies that the variation from section to section of the record is somehow random, in the sense of random variation around a "true" mean series behavior. Equally obviously, changes may reflect non-stationarity at certain scales in the velocity record. It is totally unclear how large the variation must be (or how it must behave) before some form of non-stationarity may be argued. However, this is a topic for another thesis.

<sup>11</sup>Nordin et al (1972) observed slope values increasing from 0.60 to 0.95 as the series under consideration changed from an 8 inch flume to the Mississippi River. They did not associate this variation with increasing channel "size", though if they are, it appear that slope variations are not sensitive to relatively small "size" variations (compare Mississippi and Missouri River results; their Table I).

discussed in this section though comparison and interpretation of results from the North Alouette and Squamish Rivers will be emphasized.

As mentioned earlier, various transformations of the speed data may produce slope breaks in the rescaled range plot and one of these, in particular, should be discussed. Nowell (1975) found that the derivative function produced slope breaks at sample size of approximately one second (300 sampling intervals; see his Figure 44b). This break was related to intermittent dissipation activity at that time scale (this is the usual transform for investigating dissipation).

Under a first difference transformation, similar results (not shown) can be observed for the series in this thesis; where an initial transient occurs with  $h > 0.5$ , then at sample sizes of approximately 80 seconds the slope declines to  $H < 0.5$ , and declines further at large  $S$ , without a particularly sharp slope break. This behavior becomes clearer when the nature of the data transformation is investigated. Both the derivative and the first difference operators (Note--the first difference function is the digital version of the derivative function; both have similar gain functions<sup>12</sup>; Koopmans, 1974) are linear high-pass filters. This suggests that the altered behavior of the rescaled

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<sup>12</sup>A filter operates on a series in such a way that if the gain at some frequency is low or zero, that frequency is not passed into the filtered data and consequently, its spectrum. The first difference filter also has large squared gain ( $>1$ ) at high frequencies and as a result amplifies some passed frequencies.

range plots under filtering is due to the removal of large excursions from the mean associated with long wavelengths by the spectral analysis and that whatever break is observed is a property of the gain function of the filter and the relative distribution of variance in the spectrum of the series. This view implies an intimate connection between the spectrum at given frequencies and the rescaled range plot at given sample sizes, which is to my knowledge, unsubstantiated by other research. In fact, this suggests that R/s values at large sample sizes will be related to variance estimates at low frequencies. This will be investigated later.

Since the behavior of the rescaled range plots of the North Alouette and Squamish Rivers (see Figure 7.9) appear different on first inspection they will be discussed separately. The selected North Alouette curves in Figure 7.9 show no tendency towards slope breaks (estimates are more variable at large sample sizes since the means are based on relatively few values). The apparent slope values vary between approximately 0.75 to 0.90, which are similar to values determined by other researchers (see Nordin et al, 1972; Nowell, 1975).

The selected series plotted in Figure 7.9 also show a tendency for slope to decline with increasing  $y/D$  (this was not observed by Nowell, 1975). Similarly, as discussed earlier, there is a tendency for the variance at low frequencies to decline with increasing  $y/D$ . Table 7.I shows the ranking of the

variance in the frequency band around  $n=0.001$  Hertz compared to those of R/s values at  $S=900$ . The similarity of the two rankings provides support for the idea that R/s values at larger sample sizes (and consequently, slope) are controlled by the relative distribution of low frequency energy.

However, there appears to be differences in the rescaled range behavior of the Squamish River series (see Figure 7.9). First, the variation with relative depth is not similar to the North Alouette case. Here the curves tend to be parallel with values for large  $y/D$  shifted to the right of the Figure (i.e., towards smaller R/s values). While the overall "average" slope and R/s values tend to be similar to the North Alouette (see Figure 7.10) implying that there is no need, on a gross level of comparison, to apply the type of characteristic time scaling suggested by Van Atta and Helland (1977), the internal distribution of R/s values appears to vary. While the results are not conclusive, a slope break near 150 seconds appears to occur for some of the plots. This break is clearly apparent for  $y/D = 0.33$  and  $0.80$ , very difficult to visualize for  $y/D = 0.66$  and not apparent for  $y/D = 0.72$ .

The inconclusiveness of these results increases the difficulty of explaining them. It is particularly difficult to explain why similar results are obtained for measurements at widely different  $y/D$  values and taken under different discharge conditions and why other measurements collected under similar

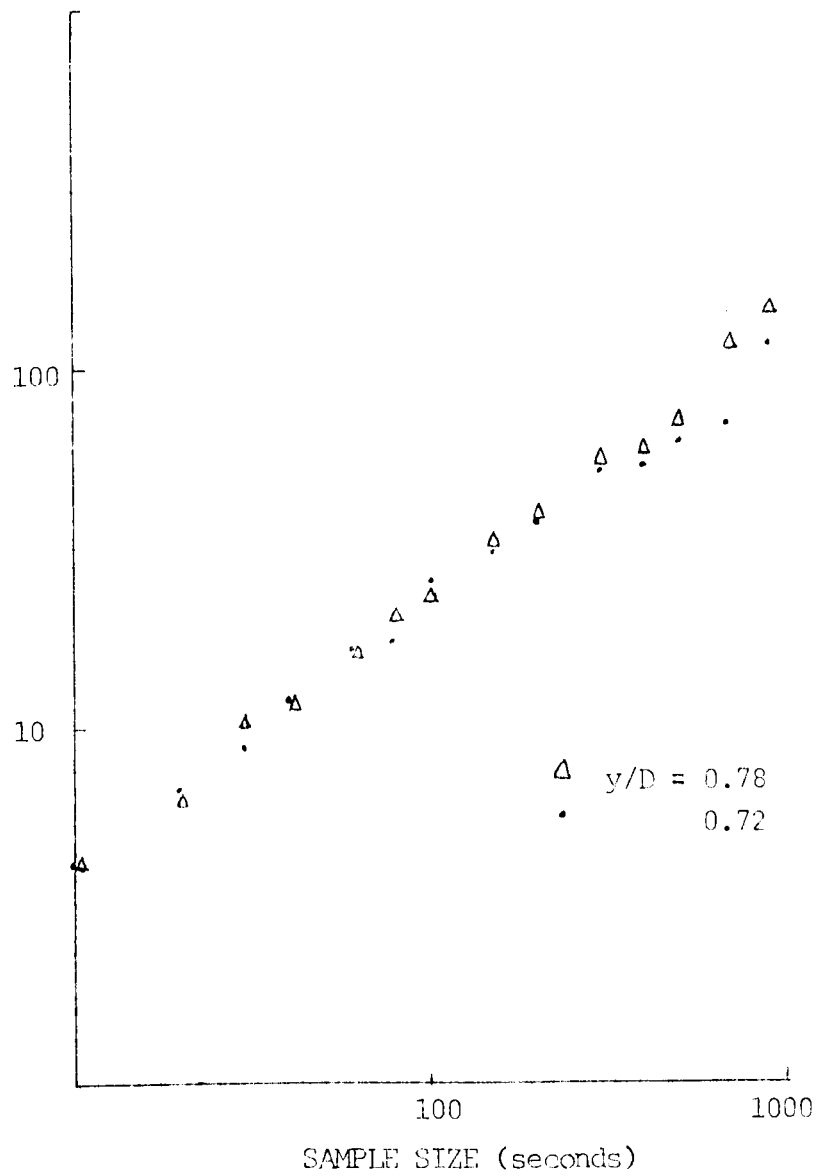


Figure 7.10: Rescaled Range plots compared for the Squamish and N. Alouette Rivers.



TABLE 7. I--RANK OF RESCALED RANGE AND SPECTRAL ESTIMATES

y/D	0.22	0.35	0.41	0.48	0.58	0.71	0.78
F/s value at S = 900	204.7	154.2	142.0	120.7	141.9	118.0	90.8
Rank	1	3	4	2	5	6	7
Rank of variance at n = .001	2	3	5	1	6	4	7

conditions do not demonstrate the same behavior.

## CONCLUSIONS

This thesis attempted a first order discussion of large scale flow features in gravel bed rivers in a simple and speculative manner. The limited literature concerning the subject was used to generate simple hypotheses and speculations that were testable with the gross analytic techniques used. Unreserved acceptance or rejection of hypotheses was usually not possible. Particularly limiting were the problems of resolving high frequencies, meter stability problems and a sampling design that was of limited application to the type of questions that were eventually asked.

Chapter I initiated and concluded a lengthy overview of what questions were suitable for investigation considering the type of data collected and the analytic framework proposed. This essentially involved framing ideas about large scale flow processes into the context of serial data and then investigating whether or not these ideas were testable. Spectral analysis and rescaled range analysis were seen as capable of providing gross summaries of the character of large scale flow features but incapable of investigating many of the commonly attributed features of these phenomena.

The experimental part of the thesis (Chapters V through VII) turned to a consideration of the time series collected in the Squamish and North Alouette Rivers. Since the time series were collected under different channel and flow conditions, the

thrust of the thesis was based on testing simple arguments that attempted to predict how the gross spectral and probability structure of the time series would vary under different conditions. Some of the expectations developed were contradicted by the results of data analysis.

Results suggest that no observable properties of the spectral or probability structure of large scale flow features were influenced by channel "size"--depth, width and along-channel variation. What this indicates is, contrary to expectation, that these features are independent of the gross features of the containing channel and that a simple collapse occurs for the spectra. Second, this absence of control by channel length scales on the spectral character implies that the various hierarchial subdivisions of the flow scale continuum concerning relations between flow and bedform scales, either do not exist or are not detectable with the robust analytic techniques used in this thesis.

Several qualifiers must be added to the above conclusions. Data series were collected only in the outer layer of the flow, especially in a region relatively unaffected by nearness to the channel boundary or flow surface. Within this region, similarity of the spectra is not unexpected; different conditions may prevail closer to the boundary. Second, it is felt that discharge variation may influence these results, despite the problems resolving the high frequency end of the spectrum.

Third, it is not clear if a larger range of "size" variation will influence the spectra.

The fourth qualifier concerns another argument generated from the literature. It was argued in the Introduction to Chapter I that the position of the flow sensor relative to large bed elements might influence the gross properties of the collected time series. This was based on the assumption that meso and macroscale roughness elements would be associated with different frequency bands and the relative decay of the features generated by apparent point sources would influence the spectral shape. While the reasoning appears incorrect (to accept this as correct would deny the previous conclusion) some results appear to be influenced by the bottom topography near the North Alouette measurement site; particularly, the intensity statistics and the mid-frequency energy content. It is suggested that the intensity statistics are dominated by scales imported from upstream and that large values are principally a result of declining local mean velocities. Second, it is argued that the variation in the mid-frequencies of the North Alouette spectra suggests that the observed shape is a combination of variance at certain scales imported from upstream and a new equilibrium structure diffusing vertically in response to local shear variation.

This argument has two implications. It suggests that the agreement noted between North Alouette and Squamish River

spectra may be slightly fortuitous--though, based on comparison to other spectra, there might be an equilibrium large scale spectral shape. Second, it implies that investigation of microscale parameters may be fruitful in elucidating the shape of the spectra of large scale flow features.

### Recommendations for Further Research

The recommendations fall conveniently into several categories; technical ones, recommendations concerned with extending this research and those that may extend investigation of large scale flow features to other areas. These will be discussed in turn.

Technical recommendations for duplicating the type of data collected in this thesis should be obvious from the data problems discussed. These include:

1. use a fixed current meter to measure longitudinal velocities--this is obviously difficult in large rivers. A more stable system for speed measurements should be designed for these cases.
2. use of electronic recording equipment for the velocity signal--this will eliminate subjective error involved in digitizing.
3. efficient low and high pass filtering of the velocity data.
4. use of smaller current meters to improve sampling near to the bed.

The recommendations concerned with extending the results of this thesis, using this comparative methodology, should be apparent from the "qualifiers" discussed in the Conclusions. Obvious procedures include comparison of spectra sampled along a channel reach (some distance that includes major bottom variation) and sampling at a point under a wide range of discharges. However, it is felt by the author that this would be close to a waste of time, since the results do little to elucidate flow behavior at large scales. This is to say, that when the inter-relationships of these large scales are better understood, then, it should be possible to predict the behavior of gross statistics under differing channel and flow conditions.

The remaining set of recommendations concerns investigation of variation within rather than only between collected time series. As was discussed in Chapter I, certain statements concerning possible non-stationarities or local (in the time series) behavior of large scale flow features were made but never tested. Several statistical techniques that may be suitable for investigating internal variation in the records come to mind. These include:

1. Pattern recognition studies--this is a partly subjective approach to internal classification of a time series.

Approaches have recently been adapted to analyze hydrologic data (Panu and Unny, 1980). Any visual study of the velocity records would be worthwhile.

2. Complex demodulation--this technique was outlined in Chapter I. There it was noted that a priori information on frequencies of interest was needed. These may be generated from the spectra in Chapter VII. For instance, the technique may be applied to the North Alouette series for frequencies near  $n=0.10$ .
3. Rescaled Range Analysis--investigation of the variation of R/s values at a given sample size. This may be helpful in a first order definition of the size and location of regions of interest in the series.

Additionally, based on the final comments of the Conclusions, it appears necessary to integrate the investigation of large scale flow features more completely with turbulent scale research. As a byproduct, it is probably worthwhile to consider the effect of microscale spacing and turbulent scale parameters on the behavior of large scale flow features.



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