ESTUARINE SEDIMENTATION AND EROSION WITHIN A FJORD-HEAD DELTA: SQUAMISH RIVER, BRITISH COLUMBIA

by

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THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF
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Geography

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Squamish River, British Columbia

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ABSTRACT

This study has determined the nature of tidal and riverine control on deposition along the lower 5 500 m estuarine reach of Squamish River; delta and floodplain sediments exposed along channel banks at 42 locations along Squamish estuary were examined. This study also has determined the rate and nature of estuarine channel bank erosion from 1957 to 1990 and has determined the nature and rate of Squamish delta accretion. Erosion rates are calculated from aerial photographs and erosion pins, while accretion data are calculated from deposition stakes.

Analysis of delta and floodplain sediments reveals that estuarine sequences comprise seven distinct facies which record varying degrees of tidal and riverine influence. While most of these facies yield evidence of their estuarine location, only the deposits of intertidal sands and tidal marsh are unequivocally tidal in origin. Sedimentation within this fjord-head environment primarily is driven by gradual channel abandonment and fill. Estuarine sequences produced by this process display a decreasing tidal influence both up estuary and up section, as evidenced by a change in sediment size, structure, and the form of facies contacts. Within this estuary the bayline is located around 3 100 m upstream from the delta front.

Erosion data indicate that the tidal deposits are very unlikely to be fully preserved in sequence because of their low preservation potential. Since 1957 the estuarine channel has shifted considerably, and predicted continued meander migration poses a threat to the future stability of the river training dyke, built in 1972 to isolate Squamish River to the west of the valley. Along certain reaches, rates of channel bank erosion have increased as a result of decreased channel and effective floodplain widths associated with dyke
construction. These factors have also led to an increased riverine influence near the river mouth which has in turn led to the formation of an anomalous coarse-grained sedimentary sequence along the lower 800 m of Squamish west delta. Accretion data reveal that Squamish west delta accretion has increased dramatically since dyke construction, and that this accretion is both spatially and temporally variable. This variability reflects the riverine source of sediment, and indicates that depositional processes differ within summer and winter months.

Comparative analysis of sedimentary sequence within fluvial and tidal reaches reveals that fine-grained deposits provide evidence of their environment of deposition. These deposits may be used as environmental indicators based on sediment-size characteristics, facies geometry, organic content and bedding form.
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First of all I wish to thank Ted Hickin for providing the initial impetus for this research, and for ensuring (through one means or another) that it be successfully completed. The content of this work has been much improved through interesting discussion with Dr. Hickin during these last months. I also wish to thank Drs. John Luternauer, Mike Roberts, and Lance Lesack for wading through this thesis with speed and good humour, and for providing thoughtful suggestions for its improvement. I am also indebted to Paul DeGrace, Ray Squirrel and Gary Hayward for technical assistance and unlimited access to computing facilities (Ray, you may have your room back now).

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DEDICATION

This thesis is dedicated to anyone who will reference it.
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CHAPTER ONE
INTRODUCTION

Background to the Study

Within the past three decades, fluvial sedimentologists have developed a number of generalised facies models of depositional environments to provide a basis for the identification and interpretation of similar systems in modern and ancient settings (Miall, 1978a). In recent years, however, it has become apparent that this type of facies modeling is of limited practical value in the field (Miall, 1984a). Environmental facies models have been found to be highly site specific, and often are of little value when taken out of the original context and applied to other systems. Another failing of these models is that they are too inflexible, and cannot account for the fact that too much overlap and gradation exists between depositional environments to warrant their simple subdivision and grouping. These comments refer specifically to fluvial facies models, though similar comments also can be made with regard to deltaic and estuarine models. Much of our understanding of deltaic systems comes from numerous studies of both modern and ancient Gulf Coast deltas, particularly the Mississippi. While this work has been extremely valuable, Miall (1984b) notes that there has been a lack of recognition of the site specificity of these facies models. This oversight has tended to bias interpretation of other depositional systems which may not adequately be explained by facies models based on Gulf Coast deltas. (Miall, 1984b)

In 1987, Miall observed that the previous nine years had seen a great number of newly proposed facies models, and modifications to older models. It had become obvious that there were many departures from the norms, and almost all facies models were based on
over-simplifications (Jackson, 1978; Bridge, 1985). Model classifications have limited real-world value, as specific processes and deposits are common to many different environments, and gradations between all end-member conditions exist (Miall, 1985; Bridge, 1985). Facies models are highly site specific, and as such, virtually unique. Despite this reservation, facies models can be of value as an interpretive tool, but only if their limitations are fully understood.

Nevertheless, a number of inaccurate interpretations have been made because of an over-reliance on existing facies models (Smith, 1987). More recent research is showing that there are many more variations in facies style than have previously been realised. The obvious conclusion to be drawn from this realisation is that there needs to be a reappraisal of many of the commonly accepted facies models such as Allen's (1970) model of meandering point bar deposits. This reappraisal can only come through more research to find which features are common to all environments, and which are unique to one particular depositional environment, or to a specific research site.

This research represents a case study of depositional and erosional processes operating within a contemporary, high energy fjord-head delta. Previous studies which have examined the nature of estuarine deposition have concentrated on environments within which tides exert considerable control. It is not yet fully known whether the conclusions drawn from these studies can be applied to high-energy environments which are controlled by fluvial rather than tidal action. Fjords are excellent locations in which to study this problem, as fjord-valley rivers commonly are high energy systems which control both the supply of sediment to the delta and its subsequent erosion. Syvitski and Skei (1983) state that fjords provide a detailed sedimentary record of both terrestrial and marine processes and dynamics. Presumably, delta stratigraphic sequences will also
reflect the interaction of these two processes, and their relative dominance, though these questions have not been fully investigated.

The remainder of this chapter is organised into a literature review, a statement of aims, and a description of the study area. The literature review first attempts a summary of sedimentary criteria which are considered diagnostic of tidal deposits. Discussion will also centre on the changing focus of research over time, and the current state of knowledge. The chapter concludes with two sections on sedimentologic research specific to estuarine environments, and to fjord deltas.

**Literature Review**

**Recognition of Tidal Deposits**

Tidal deposits are genetically defined as sediments deposited in water influenced by tidal flux. It is important to note that these deposits need not display any characteristic sedimentary evidence of their tidal origin, although discussion in this section clearly will have to be restricted to those which do. Tidal deposits may form in a variety of environments, both inter- and subtidally, and are detectable at a variety of scales. These scales range from millimeter thick mud drapes and small-scale reactivation surfaces, to the scale of environmental systems represented by coastal shelf or estuary. The following discussion summarises those features which have been recognised as diagnostic of tidal influence, and synthesises the most important research to date. It should be stressed at this point that there is a clear distinction to be made between sedimentary criteria which are diagnostic of tidal influence, and sedimentary deposits which are unequivocally tidal in origin. This point will be discussed in more detail in later sections.
There have been a number of different approaches to the study of tidal deposits. The first studies which examined the extent of tidal sedimentary control were small-scale descriptive studies of specific features observed in the intertidal zone along the coast of northwestern Europe. As few of these studies were published in English, however, their importance was not fully realised until Häntzschel published a synthesis of European research in 1939 (Klein, 1976). Post-war research into the origin and character of intertidal deposits was led by van Straaten, who published numerous papers describing intertidal facies assemblage (van Straaten, 1950a, b, 1953, 1954, 1959), later followed by Evans (1965) and Reineck (1967, 1972). The aim of these early studies was to identify zones or subenvironments within the intertidal zone. These subenvironments were differentiated on the basis of morphology, and internal structure. The range of deposits covered by these studies included tidal channels, tidal flats, and salt marsh deposits. While local variability was commonly observed to be great, typical lithofacies assemblages were often produced, showing the upward transition from tidal channel to salt marsh (eg. Yeo and Risk, 1981).

Major advances in the understanding of tidal sedimentation were made in the 1960s from a series of papers which discussed inshore subtidal deposits from the North Sea Basin (Belperio and Nio, 1988). These studies were found to be of greater importance to geologists than previous work, as most of the tidal deposits preserved in rock are of subtidal origin, a consequence of their higher preservation potential than intertidal deposits (Nio and Yang, 1991). Vertical sections, such as those presented by Yeo and Risk (1981) and Reineck and Singh (1980) not only show the nature of subtidal deposits, but also the transition from subtidal to intertidal environment (preservation permitting).
The following discussion will focus on those deposits which are considered diagnostic of tidal deposition. The results of early descriptive studies of tidal flat sedimentation will be presented first, followed by results of research conducted on inshore subtidal estuarine deposits.

**Bedding in Tidal Flat Sediments**

The discussion of tidal flat sedimentology is organised into three sections: intertidal flats, tidal channels, and salt marshes, following van Straaten's (1959) tripartate subdivision of modern tidal flat environments.

**Intertidal flats**

Many bedding types may be preserved in the sediments of intertidal sand flats. Cross-bedded sands produced by the migration of current ripples are the most common deposits. Climbing ripple lamination is very rare, and restricted to areas of high current activity and an abundance of sand, such as areas close to gully mouths. Climbing ripple lamination may develop in three basic forms (as shown in Figure 1.1), though transitional forms between each of these do exist. The transition from climbing ripple lamination in-phase to type 1 then type 2 in-drift forms, is produced by increasing flow velocity and an increasing mobility of bed sediments (Jopling and Walker, 1968).

If subordinate currents are strong enough, and if there is a source of sand to transport, then herringbone cross-bedding may develop. This is described by Reineck and Singh (1980) as cross-bedded units which display alternating foreset beds adjacent to one another in vertical sequence. In tidal flat environments this form of bedding only
Figure 1.1. Schematic representation of the three main forms of climbing ripple lamination, showing a gradual increase in bed load transport with increasing height in section. Modified after Jopling and Walker (1968), and Reineck and Singh (1980).

devlops under exceptional circumstances. Singh (1969) states that a special kind of herringbone cross-bedding is seen to have a thin mud layer separating cross-beds. This bedding is said to be typical of (though not restricted to) tidal environments. This description by Singh appears similar to the description of specific tidal bedding which would later be termed tidal bundles (discussed later in this chapter). The difference between these two deposits is that the bi-directional cross-bedded units in the example given by Singh, appear to show no dominant current direction.

Mixed tidal flats are characterised by the presence of wavy, flaser, and lenticular bedding (Nio and Yang, 1991). Reineck and Wunderlich first described the genesis of
these deposits in 1968. They state that these will form in environments in which sand and mud are available, under conditions of fluctuating current activity. As both these conditions are met within tidal environments, wavy, flaser, and lenticular bedding are commonly observed. The sand contained within these structures may indicate unidirectional or bi-directional flow. Despite this, however, these deposits again are not unique to the tidal environment, as they have been observed in modern fluvial point bar deposits (Woodyer et al., 1979; Calverley, 1984).

In some modern tidal flats, current ripple bedding is not distinguishable in any form. In these cases, extremely finely laminated or thinly interlayered bedding may be observed. These are general descriptive terms for bedding comprised of alternating thin layers of material which differs in texture, composition, or colour. These rhythmic variations of sediment are produced by pulsations in current activity or in sediment production. They may be formed over a variety of time scales from short-term fluctuations in current activity, through tidal cyclicity, to seasonal changes of flow characteristics (Reineck and Singh, 1980). Rhythmic bedding produced by tidal fluctuations is termed tidal bedding (Johnston, 1922). In these deposits, the mud and sand layers, which may be less than one millimetre thick, are deposited during slackwater periods, and during flood and ebb tides respectively. Unfortunately, without further evidence of tidal influence, tidal bedding can be extremely difficult to distinguish from other forms of thinly interlayered rhythmic bedding, which may form in non-tidal environments (Reineck and Singh, 1980).

**Tidal Channels**

In stratigraphic sequence, tidal channels are seen eroding into salt marsh and intertidal flat deposits. Megaripple bedding commonly is found in tidal channels (van Straaten,
Megaripple bedding is similar in form to ripple bedding, differing only in size, and is produced in any environment which has high enough energy conditions to form megaripples. Thus, while megaripple bedding is common in subtidal environments, it is not restricted to them, and so can only be used as a possible indication of tidal influence. As water flows over the ripple surface, a backflow current is produced by the vortex shed from this obstacle. If the obstacle is large enough, as in the case of a megaripple (greater than 60 cm in length (Simons et al., 1965)), these backflow currents may be great enough to transport material against mean flow. This produces backflow ripples (Boersma et al., 1968), which may develop into climbing ripples. As the tidal channel bed is subaqueous through all stages of the tidal cycle, wave ripple structures generally are absent, with the exception of weakly developed structures in shallow water (van Straaten, 1959). Because tidal channels remain inundated throughout the tidal cycle, there is a lack of emergence features such as animal tracks and dessication cracks. In addition, bioturbation structures are of far less importance than those found within intertidal flat sediments.

The inaccessibility of modern tidal channel deposits prevented more detailed description of sedimentary structure in these pioneering studies. Far more detailed descriptions from studies in the Netherlands, which examined recent subtidal deposits exposed in a construction trench, are reported later in this chapter.

**Salt Marsh**

Salt marsh deposits are described in terms of their structural and textural properties, organic composition, and their stratigraphic position relative to other tidal flat deposits. Evans (1965) stated that salt marsh deposits comprise well-laminated silty-clays with
some sand, plant and shell debris, although the presence of shells may only be a rare occurrence (Howard and Frey, 1973; Yeo and Risk, 1981). These layers within the salt marsh facies are described as "wavy, nodular" by van Straaten (1954), who attributes this to the deposition of sediment on an uneven plant-covered surface. Campbell and Oaks (1973) similarly described marsh sediments containing very thin wavy parallel beds which are internally structureless. Despite this apparent lack of internal structure, however, Bouma (1963) showed very fine laminations revealed by radiograph which were not visible in laquer peels. These beds may show considerable lateral continuity (van Straaten, 1959). Bioturbation and mud cracks rarely occur, and are restricted to unvegetated patches (Yeo and Risk, 1981; van Straaten, 1959). Because of this lack of burrowing fauna, bedding is well preserved, being disturbed only by roots. In sandy marsh deposits, ripple bedding may be observed, but only rarely (Bouma, 1963).

Evans (1965) argues that the salt marsh deposits may be differentiated from other tidal flat sediments in terms of sedimentary characteristics. This point was suggested earlier by Bouma (1963) and Yeo and Risk (1981). Plots of grain size for a number of tidal flat subenvironments (Evans, 1965, p. 214; Yeo and Risk, 1981, p. 249) show that marsh sediments are characteristically finer than in other intertidal subenvironments. Sediment commonly fines upsection, while organic content increases, and bedding becomes thinner and more difficult to detect (Bouma, 1963; Yeo and Risk, 1981; Martini, 1991).

Basan and Frey (1977) described salt marshes as "well-vegetated intertidal flats", a term which is perhaps something of an oversimplification, yet describes the importance of the organic matter as a distinguishing characteristic of these deposits. Barwis (1978) concluded that the organics contained within overbank fine units of tidal environments are far less diverse than those in fluvial overbank fines. In addition, because of their
estuarine setting, they will also be halophytes. Salt marsh deposits, which Barwis terms "Spartina-rooted muds" commonly contain little else other than Spartina. Occasionally, small organic detritus termed "coffee ground" organics may be present in the sediments, particularly where peat is being locally eroded (Buller and Green, 1976). Evans (1965) argued that the salt marsh environment may be differentiated from others on the basis of the amount of plant debris, as well as its diversity. The overbank units within tidal environments are said to be thicker than those in fluvial reaches of rivers (Thomas et al., 1987), with the exception of high suspended-sediment rivers (Woodyer et al., 1979; Jackson, 1981). This distinction reflects differences in flood frequency, and very different velocity-stage relationships between the two environments (Barwis, 1978).

In rock sequences, identification of salt marsh deposits is often made from an examination of the stratigraphic position of the facies in relation to bounding facies. For example, Rahmani (1988) based his interpretation of salt marsh primarily on its position between intertidal flat and swamp deposits.

**Research in Modern Estuarine Environments**

Estuarine deposits are characteristic in that they display the transition from fluvial to marine environments in a transgressive sequence. This transition may be seen both laterally and vertically. The bulk of research in estuarine environments has focused on vertical facies associations in subtidal deposits. This line of research began with work published by Oomkens and Terwindt in 1960, which reported an idealised section from the lower estuarine reach of the Rhine and Maas rivers. This research created renewed interest in subtidal deposits which had largely been neglected by previous workers. The
The discussion below is an attempt to describe an idealised section of subtidal to intertidal deposits, based on the work of Oomkens and Terwindt (1960) and others.

At the base of the idealised section, a coarse lag of shell debris and well-rounded clay pebbles overlie an erosional base (Bosence, 1973; Barclay and Davies, 1989). The shell and pebble debris may be imbricated in the direction of dominant flow (van Straaten, 1959). These deposits lie beneath large-scale cross-bedded deposits of coarse sand. Internally, these sandy foreset beds may contain evidence of bi-directional flow, separated by thin clay drapes, shell beds and peat lumps (Oomkens and Terwindt, 1960; Land and Hoyt, 1966; Van Beek and Koster, 1972; Goldring et al., 1978). The preservation of these clay layers is related to the scour-lag effect, whereby the ebb tides are unable to erode material brought in by the flood tides (van Straaten and Kuenen, 1957). The fact that these clay layers can develop to such thicknesses during one tidal cycle may point to the importance of flocculation and aggregation of suspended particles in estuarine waters (Haven and Morales-Alamo, 1972; de Mowbray and Visser, 1984).

These cross-bedded sands may also be characterised by a high frequency of reactivation surfaces (Terwindt, 1981), which may show evidence of bi-directional flow. These deposits are interpreted as being laid down by migrating megaripples in a flood-dominated tidal channel. Above this facies, both wavy and tidal bedding are reported (Oomkens and Terwindt, 1960) with occasional flaser (Van Den Berg, 1981) and lenticular beds (Terwindt, 1971), which are interpreted as being subtidal in origin. This bedding has been described in detail by Kasse (1986), who examined a one metre laquer peel taken from the bedding unit described above. Van Beek and Koster (1972) describe a very similar sequence which they interpret to represent a gradual change in environment from tidally-influenced channel to sandy tidal flats. In the upper sequence, flaser, lenticular, and wavy bedding are dominant, grading upward into tidal bedding at
the top of the sequence. Preservation permitting, this sequence is capped with salt marsh deposits, which have been discussed earlier. With the exception of this upper-most facies, this sequence is characterised by a general increase in bioturbation upsection. Evidence of marine biota and foraminifera is present throughout the section. Many of the above features are shown in the idealised vertical section presented by Yeo and Risk (1981), which is shown in Figure 1.2.

Bosence (1973) reported similar deposits from Lower Tertiary beds in southeast England. The base of the estuarine channel was scoured and filled with large mud flakes overlain by planar cross-bedding. This facies graded upward into cross-bedded silts and sands with a 10 to 20 degree dip, containing flaser, lenticular, and wavy bedding. This facies was previously described by de Raaf and Boersma (1971), and was interpreted as estuarine point bar deposits. A more complete discussion of estuarine point bar process and product is given by Thomas et al. (1987). Point bar deposits which develop in environments with strong tidal influence clearly differ from their fluvial counterparts. Intertidal point bar deposits have been modeled by Bridges and Leeder (1976), and their morphology and structure discussed in relation to tidal flow characteristics (de Mowbray, 1983). A comparison of the point bar deposits of these two end-member environments is made by Barwis (1978).

It must be stated at this point that the discussion above is of an idealised section, which may rarely be preserved in the rock record. Indeed, many authors feel that this sequence is unlikely to be preserved in its entirety. Terwindt (1971) and Jouanneau and Latouche (1981) recorded great lateral variability in lithofacies, which are often difficult to trace more than ten metres. Similarly, Goldring et al. (1978) reported rapid facies changes in Tertiary estuarine sediments, both laterally and vertically. Many researchers have
Figure 1.2. Idealised vertical section of intertidal deposits. From Yeo and Risk (1981).
commented on the internal variability of various tidal subenvironments. Dalrymple *et al.* (1991) stated that mudflat structures displayed considerable local variability, so only general trends could be noted. Similar conclusions were made concerning structures in the high intertidal flat zone (Syvitski and Farrow, 1983). This suggests that tidal environments are characterised by great within-site variability (Howard and Frey, 1973), making facies modeling a difficult task. In their attempt to distinguish between fluvial and tidal deposits in a modern estuarine setting, for example, Land and Hoyt (1966) were forced to conclude that there were more similarities in depositional product than there were differences. Terwindt (1981) reports that tidal deposits display an absence of a general vertical macro-sequence, which agrees with the earlier observations of de Raaf and Boersma (1971). On a similar note, Campbell and Oaks (1973) note that the facies cyclicity commonly noted in the fluvial environment generally are absent in estuarine sediments. Fining-upward sequences cannot be said to be the norm within estuaries.

To conclude, tidal deposits seem to be characterised by a general lack of stratigraphic order. The lack of order within estuarine sediments is not fully understood, but Goldring *et al.* (1978) believe that it is because these environments are subject to many spatially and temporally variable processes. The highly variable flow regime is caused by the continuous alteration of flow by tidal currents, wave-induced currents, and by storm events which may be either fluvial or marine in origin.

This flow variability can lead to the formation of reactivation surfaces. Klein (1970) suggests that reactivation surfaces within sedimentary facies might be used as an indicator of tidal influence. These features form in a number of environments (tidal and non-tidal) which exhibit flow unsteadiness. Klein states, however, that these surfaces in tidal environments are likely to show some degree of regularity. This regularity or
periodicity is produced by the rhythmic action of tides. This point was taken up by de Mowbray and Visser (1984), who further opined that reactivation surface morphology will differ depending upon the velocity of current action. In bi-directional flows where one current is dominant, reactivation surface morphology may reflect tidal diurnal inequality. If the dominant and subordinate currents are able to erode channel-bottom sediments, reactivation surfaces may themselves be bi-directional (Nio and Yang, 1991). De Mowbray and Visser (1984) also argue that neap-spring cyclicity may be observable in the cyclic variation in reactivation surface extent, which is determined by flow velocity and water depth. Despite these statements, however, the authors conclude that the interpretation of depositional environment must not be based on these criteria alone, but should be considered within the general framework of the stratigraphic sequence of deposits.

The above statement is typical of the cautionary notes given in papers which discuss tidal deposition. For instance, Reineck and Singh (1980) state that none of the sedimentary structures within tidal flats are restricted to that environment. Similarly, de Raaf and Boersma (1971) discuss several sedimentary characteristics which are diagnostic of intertidal and subtidal deposition, but none of these can be said to be unique to tidal environments. Ginsburg (1975) grouped sedimentary structures which are diagnostic of siliciclastic tidal flats into four categories. These categories comprised sedimentary evidence of 1) flow reversals, 2) fluctuations in current velocity, 3) intermittent subaerial exposure, and 4) alternating erosion and deposition. He stated that if evidence from only two or three of these categories could be found, interpretation should not be based on sedimentologic evidence alone. Zaitlin and Schultz (1984) similarly argue that interpretations of tidal deposits are based on a combination of
sedimentologic features and textures within sequences that comprise part of the regional stratigraphic setting.

It is clear from the above discussion that there are many characteristics and attributes of tidal deposits. Unfortunately, while a combination of many of these deposits points almost conclusively to tidal deposition, no single deposit can be said to be unequivocally tidal in origin. This statement accurately describes the state of research in tidal environments prior to 1980. After this date, Visser's research on tidal bundles provided the first unequivocal evidence of tidal deposition.

**Tidal Bundles**

The work of Visser (1980) is of critical importance to the field of tidal sedimentology. It represents a reappraisal of the genesis of large-scale bedform deposits which Allen and Narayan (1964) previously thought were laid down by dunes migrating during storm events. These bedforms comprise superimposed, bi-directional cross-bedded sets, each of which is separated by mud drapes. These mud drapes form couplets, which encase cross-bedded sands deposited by the subordinate current. Visser states that the cross-bedded sand units are deposited during successive flood and ebb tidal flows. The dominant currents deposit more material on the slipface of the bedform (which Boersma (1969) termed a bundle) than subordinate currents. These are separated (ideally) by thin mud drapes deposited during slackwater periods (Terwindt, 1971), and may be partially or completely eroded by the following tide. The cross-bedded sands display rhythmic variability in thickness, reflecting the fact that ebb current velocities exceed those of the flood currents.
In addition to this ebb-flood rhythmicity, the tidal bundles show a cyclic increase and decrease in thickness (Roep, 1991) reflecting the semi-lunar neap-spring tidal fluctuations (Visser, 1980). Since publication of these papers, tidal bundles have been reported from a number of recent (Visser and de Boer, 1982; Nio and Yang, 1983; de Mowbray and Visser, 1984) and ancient environments (Allen, 1982; Allen and Homewood, 1984; Smith, 1988c; Deynoux et al., 1993). Nio et al. (1983) investigated the internal variability and periodicity of beds within tidal bundles in an attempt to reconstruct palaeoflow conditions. The authors reported that their calculations of flow velocities and tidal ranges from these Holocene deposits agreed well with data recorded in a modern environment of similar setting.

The concept of tidal bundles has been further developed into the concept of tidal sigmoidoids. These sigmoid-shaped tidal bundles contain regularly spaced reactivation surfaces and mud drapes, and may contain evidence of changes in flow velocity within one tidal cycle (Kreisa and Moiola, 1984, 1986).

Tidal bundles provide convincing evidence of various tidal cyclicities. It may be argued, however, that these deposits are of limited value to the geologist, as the conditions for their deposition and preservation rarely are met. Middleton (1991) states that many hydraulic conditions must be operating in conjunction with each other for tidal bundles to form. Firstly, the environment must be the inner, subtidal bend of a laterally migrating tidal channel. This channel must experience bi-directional flow, where one flow direction is dominant. Tidal flows must be weak enough to allow the preservation of this bedding. In addition to this, the dominant tidal flows must be competent to transport sand, while the subordinate flows transport little to no sand. This flow asymmetry enables the sedimentary structures formed by the dominant tides to be preserved.
Finally, the environment must contain high concentrations of suspended mud which is able to settle through processes of flocculation, agglomeration, or pelletisation.

Sedimentary structures which reflect neap-spring cycles have also been found in different tidal environments. Tessier and Gigot (1989) found vertically accreted tidal bundles in an abandoned channel. Huang and Wang (1987) reported flood-ebb cyclicities in modern estuarine point bar deposits. In addition, neap-spring cyclicity may be preserved in tidal laminites (Roep, 1991), which may form either inter- or subtidally. These deposits, termed tidal rhythmites by Reineck and Singh (1980), are described in an example from the Bay of Fundy (Dalrymple and Makino, 1989). They comprise alternating sand and mud couplets, perhaps only millimetres thick, which represent deposition during periods of tidal current activity and slackwater periods respectively. Three orders of cyclicity are detected in these sediments, which are discussed fully by Nio and Yang (1991). Similar deposits were reported from the ancient record by Williams (1989), although they were only tentatively interpreted as being subtidal in origin. Middleton (1991) concluded that these other forms of tidal rhythmicity are of less value in geologic studies than tidal bundles, as well-developed cyclicities rarely are preserved.

**Transitional Environments: Fluvial-Estuarine Reaches of Rivers**

Estuarine reaches of rivers are hydrodynamically complex systems affected by a variety of processes. The interaction of fluvial and marine processes combine to produce depositional sequences which cannot be classified as simply fluvial or marine in origin. During periods of high discharges, estuarine reaches of rivers are dominated by fluvial processes which effectively dampen-out tidal influence. During low flow conditions,
however, river flow may be markedly regulated by tidal cyclicities operating over various time scales. Estuarine reaches of rivers then, are highly transitional environments, and this may be expected to be reflected in stratigraphic sequences.

Within the past decade there has been a general recognition that these dynamic systems cannot be described accurately by a generalised spatially static model. Previous attempts to produce definitional lithofacies models for fluvial and deltaic environments are of limited value in an estuarine environment. No single model can adequately describe an environment which is characterised by considerable change throughout. Despite this realisation, very few papers have attempted to identify sedimentologic evidence of fluvial-marine transitions within estuaries. In 1980, King noted that studies of estuarine sedimentation are limited in number and in scope, stressing the possible importance of different estuarine environments. Because of this, sedimentary models of these transitional environments are poorly developed (Nichols and Biggs, 1985). As these papers are few in number, and are highly relevant to this research, each is worth outlining in detail here.

The earliest English language papers to specifically examine the changing sedimentary character of deposits across the fluvial-marine gradient were presented in a special issue of \textit{Senckenbergiana Maritima} in 1975. Dörges and Howard (1975) investigated sedimentological and biological indicators of the fluvial-marine transition in a Georgia estuary. Examination of primary physical and biogenic sedimentary structures and textures enabled the authors to delineate six depositional environments within the estuary. These environments recorded the transition from fluvial-dominated, through estuarine, to marine conditions. Each environment contained characteristic sedimentary structures (shown more fully in Howard and Frey, 1975) and textural composition.
Boundaries between environments are described as gradational, so the facies associations described for each environment are idealised. Primary sedimentary structures revealed in box cores along the estuarine reach are shown in Dörjes and Howard (1975, p. 160). The most important evidence of marine conditions are interbedded sands and muds, increased bioturbation, and the presence of shells.

Howard et al. (1975) studied a series of cores taken from 5 point bars along the same estuary reach as the study above. These authors examined sedimentological and biological evidence of fluvial-marine transition from a fluvial-dominated environment to marine-dominated. The estuary is subdivided into lower, middle, and upper regions on the basis of chemical and hydrographic characteristics. Point bar sedimentary structures and textures show a definite change along the length of the river, although no attempt is made to define idealised sections from the established estuarine regions. Sediment texture fines with increasing fluvial dominance, and the dominant structure changes downstream to upstream from large-scale cross-bedded sands to ripple lamination, wavy, flaser, and lenticular bedding, and to interlaminated mud and sand.

In 1978, Jackson noted discrepancies between reports of deposits found in modern meandering point bar environments, and deposits reported from ancient point bar environments. The most important discrepancy appeared to be that the epsilon cross-stratified (ECS) deposits of modern point bars were dominated by sand, whereas most ECS reported from ancient sequences contained regularly interbedded shales within sandstone. These discrepancies prompted Smith (1987) to investigate point bar deposits forming in several different depositional environments. Smith examined point bars within low and high energy rivers, and within estuaries with varying degrees of tidal influence. From this research, Smith notes that several lithostratigraphic trends exist
within meandering river point bar deposits. Smith (1987) combines his research in meso-tidal point bar environments with the fluvial sandy point bar model (Allen, 1970; Walker and Cant, 1984) and research in low-energy meandering streams (Jackson, 1981; Calverley, 1984). From this, he proposes a threefold lithofacies classification of meandering river point bar deposits (Figure 1.3). Smith notes that each of these styles can occur within one fluvial-estuarine system, and that transitions from one facies style to another do occur.

![Diagram](image)

**Figure 1.3. Smith's (1987) threefold classification of meandering river point bar deposits**

This early research by Smith was later refined by Thomas *et al.* (1987), and Smith (1988a, b, 1989). This work is particularly relevant to the present study as it models sedimentologic change along the fluvial-estuarine reach of a river, and is the first to
develop a classification of fluvial influence based on sedimentary sequence. The most important diagnostic criteria used to identify increasing tidal influence within floodplain deposits are discussed below.

In purely fluvial reaches of rivers, point bar deposits are dominated by sand and are capped with a thin unit of overbank fines. Sequences generally record the facies trends discussed by Allen (1970). With increasing distance downstream (increased tidal influence), mud becomes increasingly important in the ECS of point bar deposits (Smith, 1987). In upstream transitional reaches, these mud layers seem to be randomly placed, but with increasing tidal dominance they appear much more rhythmic throughout, and eventually dominate the sequence (Thomas et al., 1987). With this decrease in the amount of sand in section, it is also noted that the mean grain size of the sand fines both upsection and downstream. Further evidence of tidal influence is the presence and extent of bioturbation structures, which increase in number with decreasing marine influence (Howard et al., 1975; Smith, 1987). Smith (1987, 1988a) also notes that with increasing tidal influence the overbank fines unit thickens, perhaps up to 7 m. Within lower estuarine reaches of rivers, Smith terms these overbank units "marsh mud deposits" as opposed to "overbank mud" in upstream fluvial reaches of the river. While this terminology clearly implies a genetic distinction between these two facies, no attempt was made to differentiate the two facies by stratigraphic, textural, or biological analysis.

Thomas et al. (1987) noted that, with the exception of the work discussed above, no research had examined the sedimentology and genesis of meso-tidally influenced point bar deposits. Furthermore, they are critical of previous terminology (such as Allen's ECS (1963)) used to describe these type of deposits because it implies a genetic association, or is not descriptive enough. This led Thomas et al. (1987) to propose that the terms
Inclined Heterolithic Strata (IHS) and Inclined Strata (IS) be used to describe "large-scale, waterlain, lithologically heterogeneous and homogeneous (respectively) siliciclastic sedimentary sequences, whose constituent strata are inclined at an original angle to the horizontal or palaeohorizontal" (p. 125).

Since 1987, few authors have attempted to continue this line of sedimentologic investigation in estuarine reaches of rivers. The most recent sedimentological studies of modern fluvial-marine transitions have been undertaken by Allen (1991), Nichols et al (1991), and Cooper (1993). Allen (1991) reported results of new research in the Gironde Estuary, and a summary of research previously published in French. Previous work has suggested the division of the Gironde Estuary into three zones: upper estuary channel, estuarine channel, and inlet. This tripartate zonation was initially based on morphological elements, although each zone is said to contain distinct sedimentary facies which reflect the transition from fluvial to tidal to tidal-wave energy regimes. These facies associations are shown (Figure 1.4), and a longitudinal (down estuary) facies section has been determined from cores (Figure 1.5). Fluvial sections are dominated by gravel and sand point bar deposits overlain by silty-clays. With increasing influence of tides, point bar deposits contain numerous mud interbeds and flasers. There is also an increase in thickness of the overbank fines with increasing tidal influence. In addition to these indications of tidal influence, which agree with those found by Smith (1987), Allen (1991) found more sedimentologic evidence of tides in the sand bars of the estuary funnel. These are listed as clay clasts, tidal bundles, reactivation surfaces, clay drapes, and bi-directional cross-beds, all of which been discussed earlier in this chapter. Very similar results are given by Allen and Castaing (1993) in their investigation of the fluvial-marine interface in the Garonne valley fill. The characteristic tidally-influenced sand-mud IHS point bar deposits are said to be controlled by salt
sand with flasers.
thickening, medium grained sand with
up to crossbedded

Figure 1.4. Facies associations of estuarine sediments produced by Allen (1991).

Figure 1.5. Longitudinal facies transitions determined from core by Allen (1991).
wedge intrusion and the consequent trapping of suspended-sediments within the turbidity maximum. The significance of these results are discussed in relation to valley fill lowstand, transgressive, and highstand systems tracts by Allen and Posamentier (1993).

Nichols et al. (1991) examined sedimentary facies within the microtidal James Estuary from the bay mouth to the meandering tidal river. From this examination the authors state that the estuary may be morphologically divided into three compartments. These morphologic compartments are said to agree approximately with the tripartate zonation of the estuary (sand-mud-sand) based on surficial sediment texture. Examination of sedimentary facies led the authors to conclude that no structures are either limited to or characteristic of any of the three zones identified. Nichols et al. (1991) conclude that facies appear more variable and complex within the meander zone (furthest upstream) than at locations further seaward. They attribute this to increased energy conditions and sediment supply within this zone of greatest fluvial influence.

The work of Smith (1987), and Thomas et al. (1987) has prompted a search for analogous fluvial-marine estuarine transitions in the rock record. Rahmani (1988) reported evidence of tidal influence in a Late Cretaceous channel-fill sequence in Drumheller, Alberta. He stated that documented evidence of estuarine channel-fill was scarce because they are difficult to recognise, and may easily be misinterpreted as fluvial in origin. Rahmani's interpretation of tidal influence resulted from a detailed investigation of sedimentary structure and larger-scale stratigraphic analysis. Several indicators of tidal influence are noted, the most important of which are evidence of current bimodality (de Raaf and Boersma, 1971), mud couplets and tidal bundles (Visser, 1980), and rhythmic mud interbeds within the point bar IHS (Smith 1987;
Thomas et al., 1987). Rahmani (1988) mapped the lateral and vertical facies relationships exposed along this study reach, which records a gradual change from fluvial, through coastal plain, to marine depositional environments. These facies relationships have previously been recorded in modern (Jouanneau and Latouche, 1981) and recent (Oomkens and Terwindt, 1960) estuarine environments.

In a similar study, Herbert (1993) interpreted a fluvial to marine transition from floodplain to barrier islands, tidal deltas, and tidal inlets. The main sedimentological evidence of increasing tidal influence is a longitudinal (parallel to palaeoflow) increase in the number of shale interbeds within sandstone units, to the point where shale dominates sequences. Further evidence of marine environment came from trace fossil identification.

The most recent sedimentologic research within an estuarine environment is that of Cooper (1993), who stresses the paucity of data regarding sedimentation within river-dominated estuaries. Cooper worked within the Mgeni estuary, South Africa, which he describes as a spatially static feature produced by a long-term balance between sedimentation rates and relative sea-level rise. As this estuary is not a prograding form, Cooper (1993) classifies it as a valley-fill sequence as opposed to a delta, as the valley fill does not display the typical deltaic tripartate sedimentologic sequence typified by a Gilbert-type delta. This static estuarine form is not accounted for by the classification of coastal depositional environments developed by Boyd et al. (1992).

Cooper (1993) states that river-dominated estuarine morphology is controlled spatially and temporally by the river. In the tripartate division of estuaries by Dalrymple et al. (1992), which is based on the relative importance of riverine and marine (tide and wave)
processes and associated energy levels, the central zone is that region where fluvial and marine energy (primarily tidal current) balance in the long term. Cooper (1993) argues, however, that within river-dominated estuaries this central zone is completely fluvially dominated, resulting in an estuarine sequence which displays only a bipartate longitudinal division of facies.

**Limitations of Research in Tidal Environments**

From the research undertaken by a number of workers in modern estuarine environments to date, several common conclusions have been drawn. The general conclusion appears to be that estuarine sedimentary structures do indicate their tidal origin, as sediment texture and structure gradually change down estuary. There is also another common theme to these studies. Some of the most important research has been conducted in the Gironde Estuary (Allen, Jouanneau and Latouche), the Bay of Fundy (Dalrymple, Makino, Middleton, Zaitlin), the Oosterschelde (Oomkens, Terwindt, Visser, de Mowbray, Kasse) Ogeechee River (Howard and Frey, Dörjes, Howard, Greer), and Daule River (Smith). The common theme throughout these areas is that they all display considerable tidal influence. These estuaries are characterised by tidal dominance over fluvial processes throughout - at least toward their seaward edge. This is an extremely important point because this relative dominance of tides drives the circulatory patterns within these estuaries. As such, all of the systems mentioned above are characterised by considerable flow reversals during tidal flood and ebb. Further to this, it can be said that most of the sedimentary deposits which are noted as evidence of tidal influence are produced by these bi-directional flows.
Very similar conclusions may be drawn concerning much of the sedimentologic research in tidal environments. It has already been shown that the only unequivocal evidence of tidal deposition are tidal bundles. Without evidence of these, de Raaf and Boersma (1971) state that the most convincing evidence of tidal conditions are structures such as herringbone cross-bedding. These deposits cannot form in environments where flow reversals of some form do not occur. We can see from this that the sedimentary deposits which are considered most indicative of tidal influence are produced under bi-directional flow conditions. This fact has been used in the past as an aid in environmental interpretation of ancient deposits. For example, in his description of an ancient sandstone fluvial-dominated deltaic sequence, Pulham (1989) stated that "there is no evidence of reversing flows that might indicate tidal processes, and therefore the channels are all considered fluvial in origin" (p. 188). This interpretation clearly would be invalid if these deposits formed in an environment subject to tidal flux but which did not experience flow reversals.

The importance of the above discussion is that current research of estuarine sedimentation and depositional product is inadequate. There is a need to expand the range of environments under study, as we are not yet fully aware of the specificity of estuarine models. For example, it is not known whether high energy fluvial-dominated estuaries fit into the same model or category as that developed for Daule River by Smith (1987). More research is required in environments where flow reversals do not occur, such as those found in some fjord estuaries, for example. Results of a study by Smith (1989) give an early indication that fluvial-dominated estuarine sediments bear little resemblance to those within tidally-dominated systems.
Recent research by Dalrymple *et al.* (1992) highlights the need for greater understanding of the full range of estuary types. These authors propose a conceptual framework for estuarine classification based on a geologically-oriented definition of an estuary. The reasoning employed in this paper, however, is somewhat flawed, resulting in an incomplete definition and classification of estuaries. For example, Dalrymple *et al.* (1992) state that one of the primary distinguishing features of an estuary (as opposed to a delta) is the landward movement of sediment derived from a marine source. The authors go on to define an estuary as: "...the seaward portion of a drowned valley system which receives sediment from both fluvial and marine sources and which contains facies influenced by tide, wave, and fluvial processes." (p. 1132). From this, an estuarine classification system is proposed, which contains wave and tide-dominated categories, but which ignores the riverine-dominated category. The authors justify this by stating that riverine influence primarily determines sedimentation rate, but does not control the system morphology, and presumably (though this is not explicitly stated) neither does it control sedimentary sequence.

It is clear that there exist a number of estuarine environments which fail to be classified as estuaries under the proposed classification system of Dalrymple *et al.* (1992). One obvious example is that of fjord-head delta environments which develop estuaries, yet which most commonly are progradational forms with a dominant seaward movement of fluvially-derived sediment. Within these and other systems where riverine processes dominate, it seems highly unlikely that riverine influence fails to control system morphology or sedimentary sequence. Despite this statement, research in such fluvially-dominated estuarine systems is limited.
Fjord estuaries provide similar contrast to many of the environments discussed in the literature review on tidal deposits. This contrast is again related to the fluvial-dominance of fjord estuarine systems, and concerns the transport of sediment into the delta environment. Within most fjord systems, sediment input primarily is fluvial in origin. In contrast, an important conclusion of early research in tidal flat environments is that most of the sediment originates from a marine source, and is transported by tidal flows. This is particularly true of studies undertaken within the North Sea Basin. It remains to be seen how this difference in sediment source and transport affects the character of tidal deposits.

**Fjord Delta Sedimentation**

Fjords are transitional regions between land and sea, regions of strong physical and chemical gradients. These strong gradients also exist in terms of sediment distributions, as coarse fluvial sands become finer seaward (Syvitski *et al.*, 1987). Fjord systems are dynamically unique environments. Because fjord-head deltas are laterally confined by narrow valleys, littoral transport of sediment is negligible, and deltas are almost always progradational. Climate exerts strong control within fjords, as it directly and indirectly determines the nature of estuarine circulation, and sedimentation rates and patterns. Wind is an important force in many fjords, as wind induced waves and turbulence can create distinctly different circulation and sediment distribution patterns (Syvitski *et al.*, 1987).

Canadian mainland fjords are characterised by high input rates of sediment, large river discharges which display high seasonal variability, and by strong winds. These processes operate within narrow (less than 3 km) valleys. Fjord-estuarine circulation is
driven by the fluvial input to the system, so is strong in summer months, but weak in the
winter when river flow is low. British Columbia mainland fjord oceanography is
discussed more fully in Pickard and Stanton (1980), and Thomson (1981).

The main input into fjord systems is the input of freshwater and terrestrial sediment from
rivers. In most cases this input is primarily from one source, which enters the system at
the head of the fjord (Syvitski et al., 1987). As sediment enters this deep semi-enclosed
body of water, it is deposited rapidly, forming fjord-head (bayhead) Gilbert-type deltas.
The sedimentological character of the topset beds of these deltas is determined primarily
by 1) strength and periodicity of river flow 2) river thalweg slope 3) climate 4) relative
sea-level history 5) sediment supply 6) wave energy and direction and 7) tidal energy
(Syvitski et al., 1987). Given the fact that fjords are defined, in part, by their
characteristic hydrologic and climatic regimes, deltas which form in these environments
may contain stratigraphic evidence of their fjord setting. In other words, the fjord
environment may impart a degree of control on sediment depositional form which is
detectable, and characteristic of that environment.

At present, it is very difficult to state whether fjord deltas can be identified and classified
as fjord in origin based on sedimentologic evidence alone, as very few studies have
examined the sedimentology of fjord deltas. Fjord deltaic sedimentologic research has
focused on subaqueous prodelta and delta front depositional process and product
(Gilbert, 1982, 1983; Bogen, 1983; Kostaschuk and McCann, 1983; Bornhold and Prior,
1990). These papers discuss the genesis and morphology of the bottomset and foreset
beds, but do not mention topset beds. The most recent publication dedicated to fjords
(Syvitski et al., 1987) included 1,100 references, but only five of these discuss the
sedimentology of fjord delta topset beds. Two of these publications (Ovenshine et al.,
1976, and Ashley, 1979) discuss rare and highly site specific conditions, and as such are of little relevance to studies of modern, more typical fjord deltas. This lack of research is despite the fact that fjord deltas may be ideal environments in which to study delta topset beds (Syvitski and Farrow, 1983). This is because fjord estuaries are characterised by low salinities and high rates of sediment accretion. This inhibits macrofaunal species diversity and number, which increases the potential for sedimentary structures to remain intact.

Of the few papers which have examined fjord delta topset beds from a sedimentologic perspective, the paper by Syvitski and Farrow (1983) is the most interesting. In this paper, the authors examine two bayhead deltas in terms of their subaerial sedimentary structure, sediment character, and resultant lithofacies. They are able to distinguish a number of “discrete depositional deltaic subenvironments” on the basis of sedimentary structure. These subenvironments show varying degrees of fluvial or tidal dominance. The authors found that the two deltas were dissimilar with regard to physical structures and facies development, despite the fact that they were similar in terms of the processes operating within, and building the deltas. Stratigraphic dissimilarity is believed to be related to differences in the wave angle at the delta fronts, and differences in the two river channel gradients (Syvitski and Farrow, 1983).

Bell (1975) divided the intertidal region of a fjord-head delta into three zones: upper intertidal, intermediate intertidal, and lower intertidal zones. General comments were made regarding vegetation type, sedimentary structures, and bioturbation within each of these zones, but no detailed sedimentologic analyses were undertaken. Similarly, Kostaschuk and McCann (1983) divided the intertidal zone of the Bella Coola delta into a number of subenvironments, but this was based on morphologic rather than
stratigraphic evidence. Bartsch-Winkler et al. (1983) recorded the progradation of a fjord-head delta in core. The core was interpreted to represent a change in depositional environment from subtidal, through intertidal, to supratidal fluvial deposition over a period of 14 000 years.

An examination of abandoned Holocene fjord-head deltas can reveal well-preserved stratigraphy of topset and foreset beds. Corner et al. (1990) describe the morphology and sedimentology of an emergent fjord-head delta in Norway. At this site, topset and foreset beds are revealed in raised delta terraces. While this is a fjord-head delta, however, it differs markedly from most such deltas, as it is prograding into a wide fjord. Because the delta is laterally unconfined, it has the classic fan delta morphology. Because of its unconfined nature, the delta plain may be divided into two zones: fluviually-dominated, and wave and tide-dominated. Within high-energy confined fjord-head delta systems, such zonation will be far less developed, and perhaps undetectable. Postma and Cruickshank (1988) describe the sedimentology of an abandoned Holocene fjord delta in Norway. This delta, however, is also a fan delta which displays considerable wave re-working of topset beds. These have been greatly eroded by wave action during emergence, and replaced with a beach facies.

In conclusion, there are few reported investigations of the sedimentology of fjord-head deltas, in contrast to the vast body of existing work within other delta environments. There clearly is need for greater understanding of these depositional systems. Even one of the most recent classifications of coastal depositional environments (Boyd et al., 1992) states that the classification system adopted does not specifically apply to fjord systems.

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Aims of Research

The general objective of this research is to conduct a field study of the sediments in Squamish River estuary and delta, in order to characterise the dynamics of sedimentation within a high energy fjord-head delta environment. This study is undertaken with three specific aims in mind:

1) to document the sedimentology of this fjord-head delta;

2) to determine the character of tidal and non-tidal deposits, and the interaction between tidal and riverine depositional processes along the tidal gradient;

3) to determine the rate and character of Squamish delta erosion and accretion, and the consequent preservation potential of estuarine and delta deposits.

These aims are addressed more specifically within the results chapters, and specific data requirements are discussed within the proceeding methodology chapter. Sedimentologic changes along the tidal gradient are determined from a number of detailed analyses of sediments exposed in river banks at various sites along Squamish estuary. The rate and processes of delta erosion and accretion are determined from aerial photographs, erosion pins and deposition stakes, and field observation.

Study Area

The study area chosen for this research is Squamish River, which enters Howe Sound fjord by the city of Squamish, 40 km north of Vancouver. The Squamish valley is a 150
km long glacially-cut, fault oriented valley running northeast-southwest within the Coast Mountain Range of British Columbia. Squamish River and its main tributaries, Cheakamus and Mamquam Rivers, drain over 3 600 km² of the Coast Mountains. Where Squamish River discharges into the fjord, all river flow is confined to one channel on the western side of the delta; the valley is approximately 2 400 m wide.

The study reach of Squamish River and delta is shown in Figure 1.6. Toward the top of this figure, Mamquam River confluence marks the approximate upstream limit of tidal flux, and as such, the limit of the riverine estuary. Squamish delta is divided into a number of sectors or distinct interdistributary bay environments labeled west, central, and east deltas in Figure 1.6. Squamish central and east deltas are separated by the central channel which once carried all Squamish and Mamquam River flow, but which is now isolated from flow by the river training dyke (indicated).

The length of the riverine estuary is seasonally variable, and dependent on tidal height and fluvial discharge. During winter months, when flows are low, the estuary extends about 5 500 m upstream to the confluence of Squamish and Mamquam Rivers, and the salt wedge intrusion may extend 1 500 m upstream (Levings, 1980). During summer months, when river flow is higher, the estuary is shortened, and tidal influence is far less evident. At discharges over 500 m³s⁻¹, the salt wedge is pushed out of the river altogether (Levings, 1980).

**Squamish Estuary Tidal Regime**

The tidal regime of Howe Sound and Squamish estuary is mixed semi-diurnal, typical of the Pacific coast. The average tidal range is 3.2 m, with a maximum range of 5.1 m.
Figure 1.6. Location of study area, showing the estuarine extent of Squamish River. Also shown are the major physiographic regions referred to in the text. Distance from Howe Sound fjord to Mamquam River is 5 500 m.
which classifies this estuary as macrotidal (Hayes, 1976). This tidal flux modifies river
flow in a number of ways; estuaries display large variability in flow velocity, water-
surface slope, and depth over a number of time scales. A number of these estuarine flow
characteristics are discussed below.

The purpose of this section is not to present an extensive review of the processes and
characteristics of tidal modification, but to determine the degree to which this particular
environment under study is influenced by tides. All results presented below are taken
from a study of Squamish estuary flow hydraulics undertaken by Babakaiff (1993).

With increasing distance down estuary, variations in discharge become of lesser
importance in explaining variations in flow velocities. This decreasing fluvial influence
is interpreted as evidence of a transition zone between fluvial and tidal environments
(Babakaiff, 1993). This effect is felt more on the tidal ebb than the tidal flood, where the
regions of fluvial and tidal dominance are less distinct. Maximum variability of surface
flow velocities occur at different times on the tidal flood and ebb, and are highly
influenced by tidal drop. As such, both flow velocity and variability vary with discharge
fluctuations, and with neap-spring variations in tidal drop. In addition, the timing of
velocity maxima during tidal flood and ebb (time-velocity asymmetry) display
variability with location within the estuary.

Water-surface slope data display similar temporal variability driven primarily by tidal
flux. During high and low tide periods, water-surface slopes are quite stable, while
considerable variability occurs during tidal rise and fall. This variability is poorly
explained by variations in discharge or surface flow velocities. Water-surface slope
variability appears greatest on the ebb tide, and slope increases are less rapid than those
during the flood tide. In addition, the rates of increase at one location in the estuary are not the same as those rates recorded at other locations in the estuary. This highlights the spatial variability which is induced by relative changes in the degree of fluvial and tidal influence along this transitional reach. These slope data also reveal information regarding tidal-wave asymmetry; flood tidal wave celerity is faster than ebb tidal-wave celerity (Babakaiff, 1993).

Variations in depth of the estuarine water body are principally explained by tidal drop and timing within the tidal cycle, though flood and ebb periods do not respond in exactly the same manner. In addition, predictive models of depth for locations nearest the river mouth differ from those developed for reaches further upstream. While flood and ebb tide models differ, both show a decreasing importance of discharge toward the river mouth, indicating that the estuarine reach may be broken down into a number of zones (Babakaiff, 1993) which essentially indicate relative tidal influence.

In conclusion, the above results indicate the importance of tidal influence on flow velocities, water-surface slope, and water depths within this estuary. These results also highlight the fact that the estuarine reach is highly transitional, as tidal influence increases with increasing proximity to Howe Sound. In addition to these results, one can also expect the degree of tidal influence to vary seasonally. During summer months, maximum tidal drop is greatest, yet tidal influence likely is less than in winter months. Tides are less able to modify riverine flow during summer months because the high discharges associated with freshet flows effectively dampen tidal influence. Although tidal amplitudes are smaller in winter, lower discharges increase the relative influence of tidal modification. More detailed information regarding the seasonality of riverine flow is given below.
Squamish River Hydrologic Regime

Mean annual discharge for Squamish River is about 300 m³ s⁻¹, although discharges are highly seasonal. Mean summer and winter flows typically vary from 600 m³ s⁻¹ to 100 m³s⁻¹ respectively (Hickin, 1989). Squamish River mean monthly discharge records a high degree of seasonality. In summer months, discharges peak in association with the snowmelt freshet. During winter months, discharges are, on average, lower than summer months, as precipitation falls as snow over much of the catchment headwaters. There is, however, a second peak in Squamish mean monthly flow which commonly occurs between September and February. These floods, which are associated with intense rainstorms falling on ice fields and frozen, relatively impermeable ground surfaces, generally are larger (but of shorter duration) than summer freshet floods.

Suspended-Sediment Characteristics

Only limited conclusions can be reached concerning Squamish River suspended-sediment concentrations, as little data have been collected. All results presented here are taken from suspended-sediment load measurements made by the Water Survey of Canada (WSC) in 1974 and 1975, and a limited number of surface grab samples collected by Rood and Hickin (1989) in 1987-88. The total suspended-sediment loads for these two years are estimated at 2.334 and 2.653 x 10⁹ kg respectively (Hickin, 1989). Analysis of the WSC data has been performed by Syvitski and Murray (1981), Syvitski et al. (1987), and Hickin (1989). Hickin (1989) calculated that variations in mean discharge explained 60% of the variance in suspended-sediment concentrations, although the relationship does not hold for discharges below 200 m³s⁻¹. Below this level, suspended-sediment concentrations are independent of discharge. This explains
much of the 40% unexplained variance. Temporal analysis of the concentrationdischarge data for the years 1974-75 revealed considerable seasonal hysteresis (Syvitski et al., 1987), which led Hickin (1989) to determine two suspended-sediment rating curves. The rating curve for the months April to July is associated with a period of rising discharges with the onset of the freshet. Over this period, variations in discharge explain 90% of the variance in suspended-sediment concentrations. The rating curve for the months August to March, however, shows an unexplained variance of 36%. Hickin argued that this poor fit to the rating curve was related to heavily sediment-laden winter storm events, which are superimposed on the general falling limb of the annual hydrograph. These storm events, which commonly are larger than the maximum freshet discharges, have been known to deliver over 20% of the total annual suspended-sediment over a period of five days (Hickin, 1989).

An analysis of 13 suspended-sediment samples taken by the WSC (unpublished data) showed an average composition of 58.2% sand, 37.4% silt, and 4.4% clay. More detailed size analysis was performed by Syvitski and Murray (1981) on samples collected in summer and autumn of 1976 and 1977. The authors presented full grain-size distributions of suspended particulates, and concluded that as the total suspended load increases, its mean grain size also increases. They also noted that very large daily fluctuations in suspended-sediment concentrations can occur.

More recently, Rood and Hickin (1989) have shown that suspended-sediment concentrations are greatly influenced by boil production in the estuary, and are variable over the tidal phase. This boil activity transports sandy sediments from the bed to the water surface, and distributes them throughout the water column. Sediment sampling
both within and between these boils or surface disturbances, shows that boils inject higher concentration, coarser material (than background levels) into the upper flow.

Rood and Hickin (1989) determined that the concentrations of sediment within and between boils was dependent on flow velocity, and that the concentration of boils varied with tidal phase. They concluded that boils are responsible for the suspension and transport of coarse bed material in Squamish River flow, and that they play a key (if not dominant) role in the determination of suspended-sediment concentrations and size-distributions in Squamish estuary.

Effects of Wind

Wind plays an important role in Howe Sound fjord and estuary. During summer months, heating of the land mass sets up strong land-sea breezes. The inflowing sea-breeze velocities commonly reach 40 km per hour, with gusts up to 55 km per hour (Bell, 1975). These sea-breezes occur on at least half the days over the summer period (Hoos and Vold, 1975), and can produce waves of considerable height. When combined with an incoming tide, waves approaching 2 m in height have been observed in the estuary. These waves are surface instabilities, induced primarily by incoming tidal waters, which then become enhanced by strong up estuary winds. These large tide- and wind-generated waves are restricted to the estuarine channels; they do not form within the interdistributary bay environment during periods of flood, nor do they spill over and onto the tidal marsh surface.

Nevertheless, these waves are of extreme importance to the circulation and sediment distribution patterns at the river mouth. Buckley (1977) showed that strong up estuary
winds may produce reversing surface currents, which Syvitski and Murray (1981) suggest will increase the chance settling and deposition of material held within the surface layer. They also state that even small winds can create enough turbulence to prevent the settling of clay-sized material. Buckley and Pond (1976) investigated the relative effects of wind, tide, and river flow on surface-water circulatory patterns within Howe Sound. Their study shows that wind is the major factor influencing water-surface velocities within the fjord. River flow was also noted as an important factor, but daily variations in wind speed affected velocities an order of magnitude more than daily variations in river flow and tides combined. It is important to note, however, that these studies were conducted within Howe Sound fjord and so do not fully describe the relative influence of wind, tide and discharge within the estuarine reach of Squamish River.

**Delta Sedimentation**

Little data exist regarding sedimentation rates, process, or product in Squamish delta intertidal and subtidal environments. Hickin (1989) analysed bathymetric charts of the delta front and prodelta environment taken in 1930, 1973, and 1984. Analysis of these data showed that the delta is prograding at an average rate of 3.86 m/yr. Yearly advances up to 20 m, and retreats caused by mass movements may also occur in a given year (Hickin, 1989). Bell (1975) states that the delta is prograding at a greater rate in the western sector than in the east, reflecting the engineered source of sediment input. The location of the river mouth to the western side of the delta also influences sediment-size distributions across the delta front. Delta front grab samples show that sediment is coarsest on the western side of the delta, and fines eastward, reflecting the sediment
plume path (Bell, 1975) which is shown in Figure 1.7. Both Bell (1975) and Hickin (1989) give average vertical accretion rates of the subaqueous delta since 1930.

Bell (1975) gives a brief summary of the intertidal depositional environments of Squamish delta. Discussion is limited to a description of the location and morphology of river mouth bars, tidal flats, and tidal marshes. Grain size characteristics were determined from five cores and 23 surface grab samples. In addition, observations were made regarding ripple bedding exposed on river mouth bars and lower tidal flats.

Figure 1.7. Squamish sediment plume path spreads from west to east since the river has been artificially restricted to the west of the delta system.

It is clear from the above discussion that Squamish delta is prograding into the Howe Sound at a fairly rapid rate, at least in historic times. It is also clear that certain sections of the delta are eroding at a similarly rapid rate. This is very evident in the field, and
from an examination of aerial photographs. The nature of this erosion is quite complex, as it appears to result from a combination of natural and anthropogenic processes. The nature and rate of this erosion will be discussed in chapter five. The recent history of delta building and erosion is discussed below, and has been divided into two sections, reflecting the different processes operating within the delta environment. The aim of this brief discussion is to highlight the processes operating within the delta environment with a few cited examples. It is not intended as a complete inventory of stages of delta building and erosion.

**Delta Erosion**

**Channel Stability**

The earliest aerial photographs of the Squamish delta date from around 1930. Comparison of this and more recent aerial photographs documents changes to channel planform and delta morphology. It is immediately clear from these photographs that Squamish River and delta distributary channels are dynamic systems. They display considerable channel shifting, as would be expected within such a high energy, steep, laterally confined valley.

The first recorded planform changes in this region occurred during a large flood in 1921, when Mamquam River avulsed westward. The river, which previously flowed to the east of the town of Squamish, now enters Squamish River 5 500 m upstream from the river mouth. Flow down the original Mamquam channel was greatly reduced, and had completely stopped by 1946. This resulted in a depleted sediment supply to the east of the delta, and enhanced the deposition of river mouth bars to the central and eastern regions.
At this time, the main Squamish River flow passed through the central channel of the delta, which is indicated in Figure 1.6. However, over the period from 1945 to 1972, a combination of natural channel adjustment and human-induced change led to a total shift of Squamish River flow to the western distributary channel. The human-induced changes were significant, as discussed in the next section, but it seems highly likely that they merely speeded-up the natural process of channel-shifting which would have occurred anyway. This point can be seen clearly from an examination of the 1930, 1946, and 1963 aerial photographs (Figures 1.8, 1.9, and 1.10). Figure 1.8 shows that a large point bar is developing as this meander system migrates southwestward across the valley. By 1958 (Figure 1.9) it is clear that this point bar has grown to the point where flow cannot enter the channel during low-flow conditions, and has become stabilised by vegetation. Despite this growth, flow at this time clearly was able to enter the central channel during higher discharge periods, or even periods of moderate discharge and combined high tide. Some evidence of the velocity of this flow is indicated by the presence of channel bedforms toward the upper limit of the partially-abandoned channel reach. Tidal flows in this region are not considered strong enough to have produced these bedforms. No data for this period exist regarding flow levels through the largely abandoned channel during high tidal stages or flood periods. Any flow that did exist, however, ceased by 1972, when construction of a river training dyke completely blocked flow to the meander loop, as shown in Figure 1.10.

Anthropogenic Changes to the Natural System

Early development of the delta was restricted to the eastern sector, with the growth of the town of Squamish and the Pacific Great Eastern Railway loading docks. Disturbance of the western sector of the delta began in 1945, when the westernmost channel (now the
Figure 1.8. Oblique view of Squamish town and delta taken in 1948. This shows the large unvegetated point bar and the artificial channel section. This allowed freshwater flow through the western channel, which today carries all flow.
Figure 1.9. Closer view of the channel meander and point bar shown in Figure 1.8. This photograph (taken in 1958) shows that the point bar has grown and prevented low-water flow through the central channel, creating a partially-abandoned cutoff. The point bar has become considerably stabilised by vegetation.
Figure 1.10. By this stage (1980) the central channel has been cutoff from riverine flow by the presence of a river training dyke, built 8 years previously. Although Squamish flood flows are unable to enter the channel section, tidal waters inundate the majority of this abandoned channel.
main channel) was artificially connected to Squamish River. This stretch of artificial channel can be seen in Figure 1.8. Bell (1975) states that this flow alteration would have occurred naturally with time. He estimates that the channel meander would have continued migrating southward under natural conditions, and opened up flow to the western channel by the 1960s. This artificial alteration of the flow regime was an attempt to reduce the effect of possible flooding to the developed eastern sector of the delta.

The main human disturbance to Squamish River and delta was caused by construction of a river training dyke in 1971-72. This dyke was the first phase of plans to develop a harbour and coal loading facilities in the central delta region. The 6 000 m dike was completed in June 1972, although construction of the harbour never went ahead. The purpose of the training dyke was to control and restrict the extent of flooding and prevent siltation of Squamish Harbour. The dyke restricted river flow to the western channel, and completely sealed off flow to the abandoned meander 3 500 m upstream from the river mouth.

In addition to construction of the river training dyke, the lower reaches of the west channel were dredged and artificially straightened in 1972. The Squamish Estuary Management Plan (Government of Canada, 1981) reported that over 250 000 cubic metres of material was removed from a strip that extended from the river mouth to a point 1 500 m upstream. Bell (1975) stated that this deepening of the western channel through dredging enhanced salt-wedge intrusion. This was only a short-term response, however, as Zrymiak and Durette (1979) reported that the dredged area had been completely filled by 1978.
Bell (1975) noted that construction of the dyke resulted in changes in the modes of sediment transport and deposition in the west and central deltas, which was reflected in the surficial sediment-size distributions. Bell qualified this statement, however, by noting that delta sediment-size distributions prior to construction were not known. The training dyke has also altered salinity distributions across the delta. Because of the isolation of flow to the west of the delta, the central and east deltas now experience more saline conditions. Despite this change in salinity regime, however, Levings (1980) reported that no changes to the distribution or composition of the west and central delta communities had occurred. Bell (1975) also noted that the training dyke may act as an effective breakwater protecting parts of the central delta from intense wave action during sea-breeze events.

Since construction and dredging in 1972, a large river-mouth bar has been building off the western section of the delta (Government of Canada, 1981), although part of this build-up may be attributed to the influx of sediment brought in by the realigned Mamquam River.

Other alterations to the delta system were made during construction of the F.M.C. Chemicals plant on the eastern delta, and the Squamish Forest Products ship terminal, which occupies the central delta. In addition to this construction within the central delta intertidal zone, the central channel was extensively dredged to permit docking of large cargo ships. The material dredged from this site and from the west channel, was dumped on the central delta where it may still be seen in Figure 1.6.
Squamish River estuary has responded in several ways to the forced change of regime. The main impact of the dyke has been a reduction of channel width. The major impact of this flow restriction appears to be increased erosion of channel banks along the lower 3,000 m of the river. Bell (1975) reported results of a survey in 1974 which recorded erosion of tidal marsh along both east and west banks. Considerable bank slumping along the edge of the tidal marsh of the west delta was noticed by Bell (1975). Levings (1980) reported that the five year period after dyke construction saw a westward shift of the channel, resulting in the loss of approximately 10 m (width) of the west delta marshes, which extend around 1,700 m in length. Zrymiak and Durette (1979) investigated morphological processes operating in the estuary, and reported that "considerable slumping" was taking place along the northwest bank of the estuary. This erosion was attributed to higher flow velocity in the constricted channel, and forced researchers to relocate survey stations which slumped into the river. Wiebe (1976 - reported in Zrymiak and Durette, 1979) speculated that migration of the river meander through erosion would force flows across the channel, causing erosion of the bank along the dyke. In 1979, Zrymiak and Durette reported that this erosion was indeed occurring. Erosion of this eastern bank is still occurring today, though no studies have examined its rate or extent. Zrymiak and Durette (1979) also reported that increased flow velocities due to channel constriction caused considerable bed scour, though this observation is based on a very limited data set.

It is clear from the above discussion that human activity in the delta has been extensive, and that this has led to a disturbance of the natural hydrologic regime of the delta system. The channel bank erosion and bed aggradation are the early response of the river system to adjust to the new regime conditions which exist in the estuary. What is not yet clear is how long the river will take to reach some new form of equilibrium, or quasi-
equilibrium. It is also very difficult to differentiate between natural and human-induced
morphologic adjustments of the channel. It is clear from an examination of aerial
photographs prior to 1972, for example, that channel bank erosion and channel shifting
are ongoing natural processes within this system.
CHAPTER TWO
METHODOLOGY

This chapter describes the data collection and analysis techniques employed in the field and laboratory. Where necessary, a justification of the choice of technique will be given with reference to previous studies. The chapter is organised into two main sections which deal with data acquisition and analysis separately. Within these two sections, discussion focuses on the different aspects of this research: stratigraphy, contemporary sedimentation, and erosion.

Data Collection

The majority of field data were collected over the period from early May to late August 1992. During this time, all stratigraphic work was completed, all sediment samples were collected, and erosion and deposition stakes were monitored. The collection of sediment erosion and accretion data continued on a number of occasions throughout 1992 and 1993.

Collection of Stratigraphic Data

Stratigraphic data were collected from detailed examination of sediments exposed in river bank sections at a number of locations along the tidal gradient. At each location, a number of features such as sedimentary structure, bedding form, grain-size, and colour were examined in detail. The aim of stratigraphic analysis is to determine the nature and extent of tidal control on sedimentation within Squamish River floodplain and delta
environments. As it is unclear at what scale this tidal sedimentary control will be detectable, stratigraphic analysis has to be flexible enough to detect subtle micro-scale tidal characteristics and extensive elemental-scale characteristics.

The following sections discuss the determination of site location in the field and the methods of stratigraphic analysis. These choices were governed by the aims of research, the level of detail required from analyses, and opportunities and limitations provided by field conditions. The identification of changing depositional environments throughout Squamish estuary required that a number of sections be examined along the reach of the river from estuary mouth, to purely fluvial reaches. The criteria used in the selection of sites, and site locations, are discussed below.

**Location of Bank Sections**

Squamish River is a highly dynamic river system which is constantly shifting in both fluvial and estuarine reaches, as discussed in chapter one. This lateral shifting of the river produces near-vertical eroding banks which provide good, clean, extensive sections. These bank sections provide ideal locations to log stratigraphic sections of floodplain and delta sequences. Unfortunately, these ideal conditions do not exist along the entire fluvial-estuarine reach of the river, so sampling could not be undertaken systematically throughout the study reach. Despite this, existing bank sections are considered to provide enough coverage to detect stratigraphic and sedimentologic changes down-river. Because of the extent of channel shifting within this environment, bank sections reveal floodplain and delta sediments which are considered highly representative of this system as a whole. No attempt was made to dig and log additional
sections, as this was considered too time consumptive and destructive, and produced extremely poor sections.

Specific site location was determined by a number of factors in the field: the availability of cut banks, accessibility, and condition of the face. Where possible, several sites were logged along each bank exposure at regular intervals. In total, 42 sections were logged along the lower 3,500 m of Squamish River. The locations of these sites are shown in chapter three.

One common problem with this methodology (where site location is primarily driven by the availability of exposures), is that there is a tacit assumption that all forms of river deposit are represented within the available sections. Within this study, however, it became clear that not all forms of deposit were fully represented in the study reach. Within the upstream fluvially dominated sections, there is an absence of any major overbank fine-grained deposits. This is because the Mamquam and Cheakamus are extremely high energy rivers which inject very coarse sand and gravel into Squamish River. They exert a large degree of control on the sedimentary environments of Squamish River. For this reason, major overbank fines are not visible in bank exposures between a point 5,500 m upstream from the Cheekye fan, and 1,500 m downstream of the Mamquam confluence.

It was felt, however, that these units would be of great importance to this study, as they would enable comparison of estuarine and fluvial fine-grained deposits. Early field investigations indicated that a comparison of these facies would perhaps yield the most conclusive results concerning the nature of tidal sedimentary control in lower estuarine reaches. As such, a second study reach was examined, located around 13,000 m
upstream from the river mouth. Within this reach, several exposures (located in chapter four) containing thick, fine-grained material were logged and sampled in detail. These upstream sites will essentially act as a 'standard' for comparison with estuarine facies, particularly for the bed-scale analysis.

**Logging Sections**

The methodology employed to log sections must be determined from the aims of the research, the scale of information required, and field conditions. These factors will determine the level of detail required from section logs, and as such, will almost certainly be specific to the particular study.

Despite the fact that many sedimentologic studies of this nature employ the methodology outlined by Miall (1978b), where facies coding schemes are employed to describe deposits, this approach has not been used here. I agree with many of the criticisms of lithofacies coding schemes and architectural analysis forwarded by Bridge (1993), some of which are discussed below. Formal definition and codification of facies implies that all locations display these defined characteristics. Prespecification of a limited number of facies codes may discourage detailed examination at a variety of locations in the field. This in turn may lead to inadequate description and subsequent inaccurate interpretation. Also, this rigid coding scheme inadequately describes internal variability of facies and transitional facies (Bridge, 1993). These concerns are of critical importance within this study, which attempts to determine changes in depositional product along a transitional reach of river.
A similar decision was made not to examine deposits in accordance with the architectural element analysis approach espoused by Miall (1985). It was felt that the available exposures did not provide great enough degree of three-dimensional control, which is essential for the correct identification of elements. Despite this, some of the methodologies employed in architectural analysis are used in this study. These have been incorporated because of the importance of examining facies geometry, scale, and bounding surfaces.

For the reasons outlined above, sections were logged in detail at each site without the framework of a formal coding scheme. While this approach is undoubtedly more time consuming, it is considered to provide more precise data. Once all data had been collected, field notes, photographs and stratigraphic sections were examined in detail, and deposits broken down into a number of distinct facies. This division was considered necessary to reduce the vast amount of data into some simpler form.

The definition of the term facies has been a topic of much debate, as it has traditionally been used as a simple descriptor, as a form of genetic association, and in a broader sense to denote depositional environments (Reading, 1986). The term facies is used here to denote a channel bedform-scale depositional unit which is defined by distinctive, observable characteristics, following the suggestions of Miall (1977; 1978b) and Bridge (1993). Facies identification and interpretation is based on detailed examination of composition, sedimentary structure, bedding characteristics, sediment grain-size, and colour (Miall, 1984a; Bridge, 1993). Interpretation must also be based on an examination of vertical facies associations (Walker, 1984) and lateral facies variability (Reading, 1986). This study attempts to examine sedimentary facies which are characteristic of the depositional environments under study, and which are produced by
both normal and catastrophic processes. In addition to this, this study will examine depositional products which are considered to be of unique character. While these unique deposits are not characteristic or fully representative of any environment under study, they may yield the most conclusive evidence regarding the origin (e.g. tidal) of sedimentary deposits.

**Sediment Sampling from Logged Sections**

The general aim of sampling is to test whether sediment-size characteristics reflect the increasing importance of tidal processes down estuary. Sediment samples were collected and analysed at a number of sites to test whether estuarine deposits display sediment-size characteristics which reflect their tidal origin. The choice of sediment sampling technique and scale were primarily determined by the level of information that was required from sediment analysis.

Sedimentologists must make the decision whether to collect a small number of highly detailed samples, or to collect a greater number of samples with a lesser degree of accuracy and detail (Brierley, 1989). This decision can only be taken in the field, and will be based on the type of research that is being conducted and the environment under study. In this study, it was felt that samples should be collected at a number of different scales. Firstly, data had to be gathered over a relatively large area to be able to determine the extent and character of a number of different facies or architectural elements. This was required to assess the relative importance of the tidal parameter throughout the study reach. Secondly, a number of highly detailed samples were required within specific tidal facies. This was felt to be important as initial observations suggested that
differences between fluvial and estuarine overbank units would be more easily detected from micro-scale depositional characteristics than at a broader-scale level of analysis.

Sampling was performed on a facies-scale and a bed-scale. Facies-scale sampling enabled the average grain-size characteristics of each facies to be determined. It also enabled the identification of facies trends (such as fining-upwards sequences), and facilitated more accurate presentation of section data. Bed-scale sampling enabled highly detailed sediment-size analysis of specific deposits of interest. Samples collected at this scale were analysed in detail for small-scale sediment-size characteristics within particular facies. As different levels of detail were required from the facies and bed-scale samples, different grain-size analysis techniques were chosen for each. These techniques will be discussed in detail later in this chapter.

Facies-Scale Sampling

The aim of facies-scale sediment sampling is to determine general sediment characteristics and trends within facies, and to enable more accurate presentation of section data. It was felt that highly detailed sediment-size analysis was neither necessary nor practical for this purpose, so a simple, rapid analysis was performed in the field. This was achieved with a particle size analysis card and hand lens, as described by Miall (1984a). This method is commonly used by researchers (Smith, 1987; Vanderberg, 1987; Brierley, 1989) to determine mean grain size and sorting; Miall (1984a) states that this method provides adequate and accurate results.
Bed-Scale Sampling

Before discussing techniques employed to sample sediment at this scale, it is necessary to discuss the nature of the beds that were sampled. This is merely intended as an aid to discussion here, as complete descriptions are given in chapter three. A bed is defined here as a single sedimentation unit which formed under essentially constant physical conditions (Otto, 1938), and has no limiting thickness (Reineck and Singh, 1980).

Bed-scale sediment sampling was performed on two individual facies in riverine and estuarine settings. These facies were relatively thick units of fine overbank material comprising rhythmite beds bounded above and below by thin layers of organic material. Many problems were encountered during this stage of the research. It proved extremely difficult to extract samples from a relatively undisturbed site, particularly within the estuary. At estuarine locations, water was constantly draining through the sediments at the face, making samples exceedingly wet and difficult to separate. This water was also causing contamination of beds as sediment ran down the face. Because of these difficulties, it became necessary to remove small blocks of bank material, and then sample from these blocks. As this technique is clearly destructive, all sites were logged in detail prior to sampling. Each bed was numbered and labeled, and its thickness and depth in the section was noted before the site was disturbed.

The best method of sampling proved to be peeling beds along the bounding organic layers with the aid of a trowel. As detailed sediment-size analysis was to be performed on these sediments, hands and tools were washed after every sample to avoid cross-contamination. Also, the outer layer of sediment (at the bank face) was scraped away before beds were sampled because it was felt that the outer layer of the bank material
was not representative of the beds as a whole. This sediment may have become partially oxidised due to aerial exposure, and the textural properties of the sediment may have been modified. This modification may occur by the leaching-out of fine material, through exposure to atmospheric processes (Amos et al., 1988), and due to the increased access to marine flora and fauna. In all locations where detailed bed-sampling was undertaken, parts of the facies under investigation could not be sampled using the methods outlined above. In certain locations within the facies, beds either became too small to sample, or bedding planes became too difficult to detect. At these times sediment sampling was continued upsection in one or two centimetre intervals, to ensure that the entire facies was sampled. The removal of these samples from the face was performed using a series of metal plates driven into the face at known intervals (usually one centimetre), in accordance with the technique outlined by Allen and French (1989).

**Delta Erosion**

Both the rates and locations of bank erosion along the estuarine reach of Squamish River were assessed by indirect (aerial photography) and direct (field study) means, producing an assessment of the short-term (annual) and longer-term (up to 33 years) rates of erosion. Historic rates of bank erosion were assessed as an aerial loss of vegetated delta and floodplain lands recorded from aerial photographs dating from 1957 to 1990. The difficulties involved in accurately calculating these rates include the need to standardise photograph scale, and the need to find some relatively stable feature from which erosion may be recorded. The aerial loss of material was determined from an overplot of 2 photographs of different age (Alexander and Nunally, 1972; Hooke, 1984) from which rates and locations of erosion may be determined (eg. Sundborg, 1956; Brice, 1974). To enable two images to be superimposed, traces of channel bank locations were first made,
then these traces were scanned into a software package where they could be overlaid and resized to match one another. To permit accurate resizing of scans, a number of stable reference points were marked on each trace (Hooke and Kain, 1982). Up to 6 reference points were used, including road intersections, buildings, and railroad tracks. In those cases where the correlation of these reference points was poor, scans were rejected and not used in analyses. The remaining series of superimposed scans were then printed, and areas of bank retreat were calculated with a compensating polar planimeter.

From these data, bank erosion rates (expressed as a yearly average areal loss of material) were calculated over a number of different time scales between 1957 and 1990. The volume of material being eroded could not be calculated, as channel depth at the bank was not accurately known. It is clear from sonar scans that much of the slumped bank material is not immediately eroded and removed. The near-bank area is a shallow-water zone characterised by a highly uneven, blocky bed. The fact that this zone is littered with trees (which were locally eroded), highlights the fact that removal of this material is not immediate. It is not clear what time scales are involved in removal, though several large trees remained stationary in the flow over the freshet peak, and throughout the summer of 1992.

Short-term rates of bank erosion were measured in the field along two sections of the lower estuary. These areas (indicated in chapter five) were chosen for detailed surveying, as they were areas of maximum channel erosion which often was evident on a daily basis. These sites include the regions of maximum post-dyke erosion reported by Bell (1975), Zrymiak and Durette (1979), and Levings (1980). Erosion rates of the west delta and the western section of the central delta were measured from 114 survey pins. These reference 'pins' generally were trees along forested sections of the floodplain, or
rebar stakes along unforested delta sections. Surveys were conducted throughout the summer of 1992 and on a number of occasions afterward. During each survey, distance to channel bank at a known angle was surveyed. A total number of 135 recordings were made during each survey, and these data were used to plot bank location at the time of each survey.

No attempt was made to determine rates of deposition or delta progradation from aerial photographs as there were too many difficulties in collecting accurate data. Without the aid of bathymetric surveys, the visible extent of intertidal sands is determined by tidal stage and river discharge at the time the photograph was taken. These factors, however, do not affect determination of the extent of vegetation (as long as that vegetation remains exposed). Despite this, indirect determination of sediment deposition and delta progradation from vegetation encroachment is not considered useful here, as vegetation colonisation lags considerably behind the deposition of subtidal and intertidal sands.

**Delta Accretion**

There exists a difference between short-term sediment accretion rates, and longer-term vertical building of the delta. This is primarily because marsh sediment becomes compacted during burial (Bloom, 1964), a process which is aided by the break-down of organic material over time. Unfortunately, a number of contradictory terms have been used to describe short-term and long-term rates of delta accretion. This study will employ the terminology of Ranwell (1964), who defines accretion as the depth of sediment deposited per unit time. True accretion is defined as the depth of sediment deposited per unit time, minus the decrease in that thickness caused by all processes of settlement.
Rates of delta accretion may be determined by a number of techniques, each of which has associated benefits and drawbacks. The technique most commonly employed is the use of marker beds (Stearns and MacReary, 1957; Harrison and Bloom, 1974, 1977; Letzsch and Frey, 1980; Stumpf, 1983; and Stevenson et al., 1985). These markers, for example red brick dust or aluminium powder, are sprinkled onto the delta surface then wetted to prevent their erosion during the rising tide.

Despite the fact that this is a commonly used technique, there were thought to be too many limitations and errors involved with this technique to justify its use in this study. Some of these limitations and sources of error are as follows: (1) marker beds cannot record any erosion which occurs at the site if that erosion occurs immediately after placement of the marker, or if erosion exceeds the depth of sediment deposited above the marker bed. (2) The marker bed may be mixed and vertically displaced by the process of bioturbation, or by the growth of marsh vegetation in summer months (Richard, 1978). (3) If the material used as a marker has a higher density than the water-saturated mud, the marker bed may sink through sediment layers. Richard (1978) recorded this process in laboratory studies, and so had to apply a correction factor to his data to account for the sinking aluminium powder. (4) Site disturbance takes place every time data is collected, as the site has to be dug out or cored to find the depth of the marker. Several problems arise from this site disturbance. Firstly, inaccurate recordings may arise due to core compaction during retrieval. Secondly, the extraction of a core at a specific site may lead to disturbance (settling) of the remaining marker bed, thus making future readings inaccurate. (5) Measurement cannot be continued at exactly the same location (Ranwell, 1964) (6) The surface microflora is disturbed when the marker bed is placed down (Ranwell, 1964), and the physical characteristics of the surface are altered. This last point is potentially an important one. The placement of an artificial marker on the
surface is very likely to alter surface roughness, which may accelerate or reduce local sedimentation rates. The marker also buries the surface layer of sediment, which is being bound by mucus secretions of benthic diatoms and bacteria. This binding has been shown to increase the stabilisation of estuarine sediments (Holland et al., 1974; Anderson, 1983; Grant et al., 1986; Montague, 1986; Vos et al., 1988), and may also induce further deposition because of its cohesive nature (Carey and Oliver, 1918; Coles, 1979).

As the sum of these possible sources of error is considered quite large, accretion rates were measured with vertical stakes. The decision to use vertical stakes was also based on the ease and rapidity of data collection, and the ability to collect data a number of times without disturbing the site. This method is often employed in modern delta studies (Ranwell, 1964; Hubbard and Stebbings, 1968; Carling, 1982; and Stevenson, et al., 1985), and has been justified for studies over a number of years (Ranwell, 1964). The main sources of error with this technique are as follows: (1) the stakes act as an obstacle to flow. As flow moves around this obstacle, small-scale eddies are produced, which may erode the sediment around the base of the stake. (2) Stakes are vulnerable to disturbance and dislocation, either through natural processes (floating debris, wave action), or human interference. (3) Stakes may settle through the sediment, leading to erroneously high readings of sedimentation rates (Ranwell, 1964). (4) The presence of the stake may actually promote deposition (Armentano and Woodwell, 1975). Of these sources of inaccuracy, none were felt large enough to prevent their use in this study or to invalidate results. Where possible, steps were taken to lessen the effects of error which seemed likely to arise during this study.
Initial studies attempted to determine rates of sediment accretion on the west and central deltas. Unfortunately, the deposition stakes on the central delta became disturbed by human interference in early summer 1992, so data collection was only continued on the more isolated west delta.

In total, 42 deposition stakes were placed to record rates of accretion of the west delta over a two year period. Accretion rates were determined from lengths of 3/8 inch diameter rebar, which was driven into the delta surface. The stakes, which were 155 centimetres in length, were driven 75 centimetres into the ground. This depth of burial increased stake stability, making them less susceptible to removal and decreasing the potential for the stakes to settle through the stratum. This settling process, though reported from other sites (Ranwell, 1964) is not considered important here, as all stakes had to be driven into the sediment with a mallet. The delta sediments here are considered too compact to allow any settling. The relatively small diameter of the stakes used, combined with low flow velocities on the delta surface greatly reduced the potential for erosion of sediment at the base of the stake.

The stakes were laid out in a grid system (shown in chapter five) which proved essential, as stakes (even when painted, marked with survey tape, and 80 cm high) proved extremely difficult to find in dense Carex stands. Each was painted and marked twice: once at the base of the stake (the buried section), and once at a fixed height above the delta surface. The buried portion of the stake was painted to enable a rapid assessment of whether erosion was occurring at the site. This lower marker may also be used as a stratigraphic marker if the top of the deposition stake should become bent or corroded. To increase the accuracy of sediment accretion data and to minimise the effects of scour around erosion stakes, four measurements of accretion were taken from each stake, at
each compass point. All measurements of accretion are considered accurate to within 0.1 cm for each particular stake being measured.

The location of every deposition stake was surveyed for its position and height using a total station with electronic distance meter.

**Data Analysis**

**Analysis of Log Data**

Within this study, interpretations of the environment and process of deposition of facies primarily are based on the detailed field analysis of sedimentary structure and size analysis. Interpretation is also greatly influenced by examination of broader-scale features such as facies geometry, size, continuity, bounding surfaces, and associated facies. In this sense, interpretation is also based on an element-scale approach to facies analysis. Interpretation is also aided by an examination of contemporary depositional environments and products of Squamish River, and a knowledge of other published research.

**Sediment Analysis**

The aim of detailed sediment-size analysis is to determine characteristic attributes of the sedimentary deposits under investigation. This information is often used to infer the nature of the depositional process, and the conditions of flow which were operating at the time of deposition. Much research has attempted to differentiate sedimentary depositional environments from detailed sedimentologic evidence alone (Mason and Folk, 1958; Schlee et al., 1965; Visher, 1965, 1969; Friedman, 1967; Hails and Hoyt,
1969; Royse, 1970; Allen, 1971; and Roy and Biswas, 1975), although arguably this is of limited value (Erlich, 1983; McManus, 1988).

Unfortunately, differentiation of depositional environment is not a simple matter, as the grain-size distributions produced by sedimentary analyses may not reflect the actual sediment characteristics at the time of deposition. Sedimentologists cannot always assume that the sediments under investigation were deposited as discreet grains. This is an extremely important point, particularly for researchers working in estuarine environments. In estuaries, the processes of flocculation, agglomeration and pelletization exert a large degree of control on the nature of suspended sediment (Burns, 1965; Postma, 1967; Edzwald and O'Melia, 1975; Krone, 1978; Pejrup, 1988, 1991), and ultimately, the deposition of this sediment (Haven and Morales-Alamo, 1972; Kranck, 1973, 1975, 1981; Syvitski and Murray, 1981).

The fundamental question that sedimentologists must ask, therefore, is whether they wish to examine the sediment textural characteristics at the time of deposition ("apparent" distributions (Postma, 1967)), or whether they wish to examine the true composition of their samples. This will determine the choice of sediment analysis technique, and the methods of sample preparation prior to analysis (Matthews, 1991). If the researcher has access to the appropriate equipment, this will be determined by the primary aim of research and the results which are required from analysis. As has already been noted, however, this choice is not so clear for workers examining the sedimentology of recent estuarine deposits, for a number of reasons. Firstly, once sediment has been deposited, no available technique can analyse the sediment in its originally deposited form. All analyses require some form of agitation of the sample either prior to or during analysis. This agitation will break-up sediment floccules and
agglomerates (Gibbs, 1981; Gibbs and Konwar, 1982), which have been shown to be highly unstable forms (Gibbs, 1982). A further complication is that one cannot be certain that sediments retain their original depositional form. Post-depositional modification of sediments may occur within hours of their being laid down. This modification may take place through a number of processes. During burial and compaction, flocules may begin to break down. This may be enhanced by fluctuating water-table levels and chemical gradients in intertidal zone sediments. As sediments accrete and become drier, soil forming processes may lead to the formation of aggregates. In addition to these processes, flocs, aggregates, and pellets may either be formed or broken by the action of benthic diatoms and burrowing fauna in the substrate. It is unclear what effect organic decay and oxidisation will have on surrounding sediments.

From this discussion it is clear that the sedimentary properties of a sample cannot accurately be examined in its original depositional form. For this reason it is felt that sediment analyses undertaken within this study should aim to determine the true grain-size distributions of samples. The term 'true distribution' is meant to indicate the distribution of the total discreet constituent particles of a sample. It is not meant to indicate any accuracy of the distribution, as the techniques of size analysis chosen in this study only measure some size property which is related to the actual size of the sediment, whether that is a Stokesian diameter or the smallest dimension of a grain. Once this decision is made, it is vital to ensure that all analysed samples are fully dispersed prior to analysis. The methodology employed to disperse samples is discussed in the following sections.
A number of sediment analysis techniques have been chosen in this study. The selection of each technique was made in accordance with the level of information and precision that was required from the analysis. Different techniques were applied to samples obtained from facies-scale and bed-scale sampling. Samples taken during the facies-scale analysis section of this research were analysed with a particle size analysis card. Sediments taken during bed-scale sampling were sieved, and the subsieve fraction analysed with a SediGraph particle size analyser.

**Preparation of Samples for Analysis**

Figure 2.1 shows the main steps that were taken during preparation of samples for bed-scale analysis. The aims of this analysis were to determine the size-distribution of each sample (for the sieve and sub-sieve ranges), the sand, silt, and clay components, and to determine the amount and type of organic matter in each sample.

Samples were oven-dried at 45°C for one week to remove 'free' (all but interstitial) water from the sediment. Once dry, samples were disaggregated by very lightly pressing the sediment in a mortar. Extreme care was taken during this process, as the aim was merely to break up aggregates prior to sieving. Samples were not ground, as this can split larger grains, producing an incorrect shift to a finer distribution.

**Sieving Analysis**

Samples were weighed, then sieved for 15 minutes from 0 to 4 phi in accordance with the technique outlined by Folk (1965). The only difference between the methodology employed in this study and that of Folk is that samples were sieved at full phi rather than
Figure 2.1. Flow-chart showing the methods of sample pre-treatment and analysis, and information provided. * See text for an explanation of lettering
quarter phi intervals. There were several reasons for this choice of sieving interval. Firstly, it was felt that the breakdown of sediment distributions into phi intervals would be adequate for the purpose of this study, and that further resolution was not needed. Perhaps more importantly, it was felt that the increased resolution that would be gained from quarter-phi analysis would be outweighed by the increased error associated with it. The loss of material during sieving is potentially the most important source of error in size analysis, and losses increase with the number of sieves used. The error associated with the loss of material from 17 sieves required for quarter-phi analysis, was judged to be unacceptable in this study. The error was expected to be particularly large, as the material would have to be sieved and removed three times per sample due to the limited number of sieves that can fit on the mechanical shaker.

This sieving process essentially produces two subsamples - the sand range held in the sieves, and the subsieve range held in the pan (Figure 2.1). The analysis of subsieve material is discussed in the proceeding section and the analysis of the sand is discussed below.

Large organic material was carefully removed from the 0 and 1 phi sieves, where all large material was retained. This organic material was set aside, and its weight subtracted from the dry weight of the sample prior to analysis. The material retained on the sieves was then weighed, and examined under a microscope to detect the presence of aggregates. The number of aggregates observed were minimal, and were certainly not present in significant numbers to invalidate results, or to require that the sample be disaggregated and re-sieved (Folk, 1965). Almost all samples collected weighed between 30 and 50 grams (dry weight), to minimise the error introduced by sieves clogging up with sand (Folk, 1965).
After this sieving process, and the removal of large organics, it was felt necessary to remove the remaining (smaller) organics from the sand subsample. This was done for two reasons: to determine the total amount of organics in each sample, and to increase the accuracy of the size-distribution data for the sand range. The inaccuracies in these data arise from the fact that the weights of material retained in each sieve record the weight of sand and small organics. To calculate accurate weights of the inorganic component of samples, weights had to be corrected (reduced) for each sample, at each phi size class. Organics were removed using the loss-on-ignition technique (LOI), described by Thrower and Schmidt (1985). More detailed description of this removal process is given later.

**SediGraph Analysis**

Analysis of the subsieve (silt and clay) material was performed with a model 5000D SediGraph particle size analyser. This technique only requires a small sample size (around 1 gram) for analysis, so the subsieve material had to be subdivided and reduced. Unfortunately, subsampling errors inevitably occur whenever any sample is split for analysis, and these may account for the greatest error of SediGraph analyses (Coates and Hulse, 1985). Because of this, great care must be taken while splitting samples, and sample preparation was standardised as much as possible to guarantee reproducibility of results (Vitturi and Rabitti, 1980). The method employed to obtain this small subsample was that of coning and quartering (Duncan and LaHaie, 1979).

Once the subsample weight had been reduced, it was necessary to test for the presence of carbonates in the sediment. Of the 188 samples tested, however, none tested positive; carbonates clearly are of minor importance in the samples collected here. The lack of
carbonates in these sediments may indicate that carbonate minerals entered into solution and leached out of the sediment (Nelson and Sommers, 1982), or more likely, were broken down into very fine particles which were removed in suspension (Chave, 1964). Land and Hoyt (1966) similarly found little or no carbonates present in estuarine sediments finer than 2 phi.

The remaining organic material had to be removed from the SediGraph subsample before size-analysis could be performed. For this stage of the analysis, a decision was made not to remove organics by the LOI technique, but to chemically remove organics with hydrogen peroxide. The reasoning behind this decision, and more detailed description of the methodology employed are given in the next section.

After chemical burning with hydrogen peroxide was complete, samples were placed in a centrifuge for 15 minutes, decanted, then cleaned with distilled water. This cleaning was repeated three times (one hour total centrifuge time), to ensure that samples were clean. The samples (which were then ready for size-analysis) were dispersed in 0.08% Calgon solution (Sodium Hexametaphosphate), which Tchilingarian (1952) determined to be one of the effective dispersing agents. This concentration (of calgon to sediment dry weight) was chosen because it offers the most effective dispersing capability, without significantly altering the density and viscosity of the fluid (Micromeritics, 1982). This simplifies operational procedure, as the density and viscosity values for water may be used to calculate the running rate of the machine. Samples were prepared to a 3% concentration of sediment (dry weight) to dispersing solution (calgon and distilled water).
In this study, one of the main sources of error likely is the determination of sample particle density. This would lead to an erroneously fast or slow running rate of the machine, causing a shift to a finer or coarser distribution respectively. As such, the average particle density of the samples under investigation was calculated, rather than assuming the particle density of Quartz (2.65 g/cc). This was achieved by analysing the specific gravity of 20 samples from the site under investigation with a pycnometer, following the methodology outlined by Hurlbut and Klein (1977). This analysis determined that the average specific gravity of samples analysed in this study was 3.07, which clearly shows the need to determine the density of samples under study, rather than assuming the density of Quartz.

**Removal of Organic Material**

Previous sections describe how samples were divided and further subdivided (Figure 2.1). This section describes how organic material was removed from three subsamples: the sand, the small subsample required for the SediGraph, and the remaining silt and clay material (marked A, B, and C in Figure 2.1). It was necessary to remove the organics from the sediment for a number of reasons. Firstly, the weight of organics represents an error in the weight retained on each sieve, and equivalent weights determined from SediGraph printouts. If uncorrected, this will lead to incorrect grain-size distributions. More importantly, organic material has to be removed from the SediGraph subsample as it introduces a potentially large error into the results. This error arises because organics can greatly impede the full dispersal of a sample in solution (Coakley and Syvitski, 1991), and may induce flocculation during analysis (Micromeritics, 1982). Also, these organics will not settle through the solution, but
instead will float, causing greater attenuation of the x-ray beam. This will produce an erroneous shift to a finer distribution.

No single standardised procedure of organic removal was employed for all three subsamples. Instead, two different procedures were employed, the choice of which was governed by the nature of the problem in hand and the requirements of the proceeding analysis.

For the removal of organics from the main sand and silt/clay subsamples (labeled A and C in Figure 2.1), the LOI technique of Thrower and Schmidt (1985) was employed. This method was chosen because it enabled rapid and accurate analysis of the large number of samples which were analysed. Samples were weighed, placed in a muffle furnace set at 420° C for 45 minutes, cooled in a dessicator then re-weighed. Mitchell (1932) and Ball (1964) have shown that there is no loss of structural water from the inorganic component of soils at temperatures up to 440° C, so this form of error is not a consideration here. A number of samples were burnt at the same temperature for greater periods of time (up to 12 hours) to determine whether all organic-carbon had been removed after 45 minutes. In no case was any further weight loss recorded.

It is important to note that the LOI technique does not directly measure the organic content of a sample, or the total carbon content. When samples are placed in a furnace at 420° C, the organic carbon component is burnt off. As such, the LOI technique provides an indirect assessment of total organic content, as organic matter comprises around 50% organic carbon (Donahue et al., 1983). The organic matter content may then be determined by multiplying organic carbon values by some conversion factor, such as the Van Bemmelen factor of 1.724. More recent studies, however, have shown this factor to
be unreliable (Nelson and Sommers, 1982) because the ratio of organic to inorganic fractions is different among soils, and variable with depth (Broadbent, 1953). With this in mind, it was felt that too much uncertainty exists to warrant the use of any conversion factor without first testing for its applicability to the samples under study. For this reason, the results of all organic removal techniques will be expressed as organic carbon content only. The accuracy of the LOI technique in determining organic carbon content has been statistically confirmed by Craft et al. (1991).

For the removal of organics from the subsample to be analysed by SediGraph (labeled B in Figure 2.1), a decision was made not to use the LOI technique. This was because it was not certain that the silt and clay-sized sediments would retain their form at these high temperatures. In addition to this, further preparation would be required to remove the ash from sediments after burning. For these reasons, organic removal from the SediGraph subsample was performed by chemical burning with Hydrogen peroxide (H₂O₂), following the methodology of Robinson (1927) and Anderson (1963). This chemical reaction was further induced by placing samples on a hotplate. Hydrogen peroxide was continually added to the sample until reaction visibly ceased, which commonly took up to one week.
CHAPTER THREE
FACIES-SCALE STRATIGRAPHIC ANALYSIS

The following sections present the findings of detailed facies-scale analysis of delta plain and alluvial plain sediments observed in sections along banks of Squamish River. Sections were logged in accordance with the procedures outlined in chapter 2 along a number of reaches, and have been subdivided into distinct facies. The location of bank sections analysed in this study is shown in Figure 3.1. In total, 7 facies have been identified throughout the fluvial-estuarine reach, and these are described below in order of their occurrence in sequence from bottom to top. Facies interpretation and reconstruction of observed estuarine deposits follows these descriptions.

Facies Description

**Facies A: Cross-Bedded Sands and Gravels**

Limited information exists for these sediments, as it was only possible to log and photograph them at two locations. At all other locations, water seepage caused pits to collapse before analysis could begin. For this reason, most of the known depths of this facies were determined by augering.

At those locations where this facies could be logged in detail, it comprised coarse sand and gravel to unknown depths. The gravel and pebbles (which are all well-rounded) occur as discrete layers and lenses within the coarse sand. Contacts between sands and gravels commonly are abrupt, and occasionally erosional. The coarse sands are cross-bedded (bed thicknesses up to 20 cm), and separated by numerous erosional surfaces (Figure 3.2). Erosional surfaces are flat, producing planar cross-bedding. Foreset beds
Figure 3.1. Aerial photograph showing the study reach from Howe Sound fjord to Mamquam River with the locations of study major reaches within which stratigraphic sections were logged. Distance from Howe Sound fjord to Mamquam River is 5 500 m.
Figure 3.2. The internal character of Facies A sands around 1 m below the upper limit of this facies. This site (wd9) is located 450 m upstream from the river mouth, along the west delta. Upstream is to the right. Ruler is 18 cm for scale.
have high angles of dip (up to 35°), indicating bedform migration down-valley; topset beds are preserved in some cases. Within the coarse sands are numerous medium and fine sand beds and flasers. These beds have highly irregular bounding surfaces, and may be laterally and longitudinally discontinuous. Occasionally, the flasers contain coarser material than the surrounding matrix. All contacts are abrupt, and non-parallel. Some large organic debris is found within the sands, but no smaller small organics (termed coffee-ground) were observed in megaripple foresets (terminology of Reineck and Singh, 1980). In a number of cases, small mud clasts of clean silty fine sand were found within this facies. The largest of these was a fist-sized mud clast, which did not appear to contain any internal bedding, though the form collapsed before it could be examined in detail.

Along the lower 3 300 m of Squamish River, these sediments are overlain by the finely-laminated sands and silts of Facies B. The contact between Facies A and B is abrupt and wavy, but from these limited exposures, does not seem erosional. The sand immediately below this contact is heavily oxidised and commonly cemented to a variable depth of 1 to 2 cm. At one site (ox3) the upper limit of this oxidised sand layer appears highly irregular and jagged.

From the longitudinal section of logged sites along the west delta (Figure 3.3), it is clear that the upper surface of Facies A is highly variable with depth. For example, between sites wd 16 and wd 17 (209 m apart) the depth of this facies varies from -2.55 to -1.10 m a.m.s.l.. The lower boundary of Facies A was never reached.
Figure 3.3. Longitudinal section of Squamish estuary stratigraphic sequences, showing a representative number of logged sections along the tidal gradient. See text for an explanation of the letters A and B.
**Facies B: Finely-Laminated Sands and Silts**

This facies, comprised of material ranging in size from clay to medium sand, has a banded appearance produced by alternating beds of predominantly light (silt and clay) and dark (fine to medium sand) material. These parallel wavy beds are of variable thickness within and between sites. This alternating bedding can be seen in Figure 3.4, which shows the uppermost 60 cm of Facies B sediments and the transition to Facies C. These deposits are exposed along the west delta at a site 690 m upstream from the river mouth (between sites wd11 and wd12 in Figure 3.3). Contacts between alternating light and dark beds here appear slightly gradational, and occasionally abrupt. At this location, lighter beds commonly are thicker (up to 7 cm) than darker beds (up to 3 cm). Further upstream (around 3 000 m from the river mouth), Facies B deposits may contain occasional coarse sand interbeds.

When these sediments are dug out or cleaned with a trowel, as seen in Figure 3.4, these beds appear massive. Along sections which have been exposed to subaerial erosional processes, however, beds display very fine laminations. At three locations along the west delta (located 690 m, 1 050 m, and 1 450 m upstream from the river mouth), it was possible to examine this small-scale sedimentary detail, and log small sections of Facies B at a millimetre scale. These three sites revealed considerable within-site variability of Facies B deposits along Squamish estuary. The following discussion will concentrate on those features which generally appear to characterise these deposits, and those features which make each site distinctly different from others.

Small-scale sedimentologic analysis reveals that both dark and light beds contain rhythmically alternating light and dark laminae or groups of laminae. These alternating
Figure 3.4. The banded appearance of Facies B sands and silts seen along the west delta around 690 m upstream from the river mouth. This section, which has been cleaned with a trowel, displays no lamination within beds. A number of small shells and shell fragments may be seen in this photograph. Upstream is to the right.
layers vary in thickness from 0.01 cm to 2.5 cm. Darker laminae are a fine to medium sand, while lighter layers are silts and clays. Generally, the darker, coarser laminae are thicker, and may comprise parallel wavy bedding, ripple foreset beds, climbing ripple lamination (in-drift), and complete ripples preserved in trains. The light silt/clay layers may not exceed one lamina thickness, but commonly comprise a number of parallel laminae. These laminae take on the form of the underlying surface, and so may appear wavy. Despite this wavy appearance, the lighter laminae never display any form of ripple bedding.

A number of these features described above are observed in Figure 3.5, which shows a section of Facies B sediments exposed at the upstream most end of the west delta, around 1450 m upstream from the present river mouth. The section seen here is around 50 cm high, and 70 cm wide. At this location there appear to be 5 or 6 units characterised by dominantly dark, coarse beds, capped by a thinner series of dominantly lighter, finer beds. At this location, darker beds seem thicker than the lighter beds of material, in contrast to those observed in Figure 3.4. Both these lighter and darker beds contain rhythmically alternating fine laminations of sand and silt/clay. In the coarser material, groups of laminae may be produced by the migration of ripples and the deposition of inclined foreset laminae. This thinly interlayered bedding is seen in closer detail in Figure 3.6. From this figure, one is able to determine the nature of contacts between light and dark laminae. The lower contact of lighter layers most often is abrupt, but may occasionally be gradual, while the upper contact (to overlying dark laminae) is always abrupt, and sharply defined. Figures 3.5 and 3.6 also show a ripple train preserved in these deposits. In all, 13 ripples are seen here, and these range in size from 0.2 cm in height by 0.5 cm in length, to 0.5 cm by 2.2 cm. These ripples are not connected (ripple troughs are essentially the upper surface of the
Figure 3.5. Facies B deposits exposed along the upper reaches of the west delta, around 1500 m upstream from the river mouth. There appear to be either 5 or 6 fining-upward units visible here. Midway along the ruler (which is 18 cm long) there is a train of ripples. Upstream is to the right. The location of this site is marked by the letter A in Figure 3.3.
Figure 3.6. A portion of Facies B deposits seen in Figure 3.5 showing the very fine alternating sand and silt/clay laminae which are found throughout the deposits at this location. At least 5 of the ripples from the ripple train can be seen here. The two asymmetrical ripples show that their direction of migration was from left to right in this picture, which is upstream. The location of this site is marked by the letter A in Figure 3.3.
underlying lamina). This gives the ripples the appearance of single thick lenses within lenticular bedding. With two exceptions, these ripple forms are perfectly symmetrical, and display no internal bedding. The exceptions to this are two ripples (shown on Figure 3.6) which are clearly asymmetrical, one of which contains visible foreset laminae. The asymmetry of these forms indicates that this ripple train was migrating from left to right in the photograph - in an up valley direction. This is confirmed by the directional evidence provided by the foreset laminae, which dip up river.

Facies B sediments were also examined at site wd15, a point 1 050 m upstream from the present river mouth. At this site, bedding appeared quite different to that at other sites previously discussed. At this location, deposits display banding very well, as beds of finer, lighter material rhythmically alternate with beds of coarser, darker material (Figure 3.7). This rhythmic bedding is evident throughout Facies B deposits at this site, which are at least 90 cm thick. Bedding is distinctly different to that seen in Figures 3.4 and 3.6, as all beds are parallel, sub-horizontal, and laterally continuous. Beds are only very slightly wavy, and lack all signs of ripple structures. In addition to this, sediments are notably lighter in colour, and appear to be finer than at any other site in the estuary.

Toward the base of Facies B deposits, flaser and lenticular bedding may develop. Toward the top, marsh vegetation can be seen in these sediments, but this does not disturb the bedding patterns reported above. Occasionally, shells and shell fragments were found within the sands of Facies B. A number of these are visible in Figure 3.4. These shells, however, are not present throughout all reaches of Squamish estuary. Shells were only found within sediments along the lower 1 000 m of Squamish west delta, though it is difficult to assess the exact upstream limit of shells because of the limited exposure of these deposits. Shells were similarly found along the seaward edge
Figure 3.7. The uppermost 70 cm of Facies B at site wd15, 1050 m upstream from the river mouth. This highlights the rhythmic alternation of darker, coarser beds and lighter, finer beds, and also shows some of the fine parallel laminae within beds. Ruler is 18 cm for scale. Upstream is to the right.
of both central and east deltas.

Facies B sediments are underlain by Facies A (an abrupt contact), and overlain by the silty rhythmites of Facies C. In exposed sections, the contact between Facies B and C appears abrupt, but on close examination it can be seen to be gradual. With increasing height in Facies B, very fine (sometimes discontinuous) parallel organic layers are seen. At this stage, the sediment does not appear to change character. Within 5 to 10 cm (vertical distance), however, the organic layers become more distinct, and the sediment (now in the transitional region between Facies A and B) rapidly fines upward to a silty fine sand. All evidence indicates that, where Facies B is present, it is underlain by sediments of Facies A. Unlike the deposits of Facies A, Facies B deposits are not observed along the entire study reach, but instead appear to pinch-out between sites ny2 and ny3, around 3 300 m upstream from the river mouth. At their upstream most limit, Facies B sediments are 15 cm thick.

Within the lower estuary, Facies B deposits appear to be present across the entire delta, and are certainly visible at the base of sections along the west, central, and east deltas. Toward the front of the west delta and along tidal channels, Facies B sediments have been exposed and are being reworked. At these sites, sediments show numerous bioturbation structures, and contain small shells. Bioturbators are believed to have colonised these sediments after their re-exposure upon erosion, as bioturbation structures were never noted in Facies B sediments at other locations.

Along west delta sections, where Facies B is most visible, deposits display great variability in thickness over short distances. At site wd16, deposits are 182 cm thick, while 210 m upstream, this same facies is only 48 cm thick.
**Facies C: Silty Rhythmites**

This facies ranges in thickness from 0.25 m to 1.5 m, and consists of alternating thin layers of inorganic material and layers dominated by organics. One pair of these layers forms one bed, which may be considered a couplet. Exposed sections of this facies (along river banks) become differentially eroded, as inorganic layers are scoured-out by Squamish flow. The dominantly organic layers are less easily eroded because of the binding effects of the vegetation matting, and so protrude from the face up to 1 cm. This is clearly seen in Figure 3.8. The organic layers are of fairly constant thickness throughout the facies (around 0.1 cm thick), and comprise a mat of flattened reed stems. At one location, a deciduous leaf was found within these sediments. Inorganic layers are of variable thickness (adjacent layers may vary from 0.8 cm to 2.0 cm), but generally decrease in thickness upsection. This produces the characteristic appearance of this facies, which is seen in Figure 3.9. Inorganic layers are comprised of material ranging in size from silty-clay to sandy-silt, and contain occasional vertical reed stems.

Sedimentary characteristics vary between sites and within beds at one site, but it cannot be said that beds display either fining or coarsening-upward trends. Despite this, beds occasionally display sediment-size heterogeneity. Commonly, no bedding structures are visible within these rhythmite beds, although at some locations, extremely fine parallel laminations are visible within the inorganic layer. Rhythmite beds are parallel, wavy, and may increase or decrease in thickness across the section, sometimes pinching out altogether. In addition to this small-scale variability of rhythmite thickness, rhythmite beds also appear wavy or undulatory over distances ranging from metres to tens of metres. Although sections are sometimes obscured by slump blocks, rhythmite beds may be traced tens of metres both parallel and perpendicular to contemporary flow. Toward
Figure 3.8. A typical view of Facies C rhythmite beds, seen here in section along the west delta. Differential erosion has left organic-rich layers protruding from the face. Beds decrease in thickness upsection. Ruler is 18 cm. Upstream is to the right.
Figure 3.9. Site wd15 showing the typical section appearance along upper reaches of the west delta. Facies B, C, and E are seen here. Shovel is around 1 m high. Upstream is to the right. Also shown is the section diagram for this site, and approximate facies boundaries.
the back of the west delta (approaching the valley wall), rhythmites appear different to those exposed along the river bank. The inorganic component seems finer, and organic layers are less clearly defined.

With increasing height in section, the deposits of Facies C gradually change character. This transition occurs as beds decrease in thickness upsection to the point where distinct alternating organic and inorganic layers can no longer be seen. At this point the sediments appear to contain fine, parallel beds or laminae. Eventually, these laminae become so fine that they can no longer be counted, and deposits take on the form of massive, organic-rich fines.

This facies is observed throughout exposures along the lower 3 200 m of Squamish River, and is visible along the tidal marsh cliffs of the west, central, and east deltas. It is also visible in a few small sections within the Mamquam blind channel, although most of the sections along this watercourse have been destroyed by industrial activity. Along Squamish river sections, Facies C deposits thin with increasing distance upstream. As deposits thin, they also become deeper in section. Toward the upstream limit of this facies the fine sediments may contain sand interbeds, as recorded toward the upper limit of the ox face (Shown in Figure 1.1).

The rhythmite facies overlies Facies B, and along the lower 1 100 m reach of the west delta, it is overlain by Facies D. The contact between Facies C and D is abrupt, as the silty rhythmite sediment immediately changes to medium and coarse sand. This contact is clearly seen because the coarser sands are more easily eroded by Squamish flow. Because of this erosion, the upper surface of Facies C is often marked by a ledge (produced by the removal of overlying sand) which may be up to 1 m wide. At all other
locations, where Facies D is absent from stratigraphic sections, Facies C is overlain by Facies E. The contact between Facies C and E is gradational. Indeed, it is often difficult to determine the location of this transition. It is most easily noted by a gradual increase in grain-size to a fine sandy-silt. Along the ox face, the approximate location of this transition is marked by a dark, wavy layer. On close examination, this dark layer contains at least 5 dark (almost black) beds (around 0.2 cm thick) separated by the light silts and sands observed toward the top of Facies C. Above this dark layer, material coarsens to the sandy-silt of Facies E.

From this discussion, one can see that Facies C is characterised by a great deal of local variability of bed thickness and composition. Despite this variability, however, the rhythmite facies always retains the characteristic appearance shown in Figure 3.9. In addition to these local differences, there is some indication that there exists a more orderly, gradual change in rhythmite character with increasing distance upstream. The nature of this change will be discussed later in this chapter.

**Facies D: Sandy Rhythmites**

This facies is only found within west delta sections along the lower 800 m of Squamish River, where it overlies Facies C. Along this reach, sandy rhythmite sediments form the uppermost deposits of the topset deltaic sequence. They comprise alternating layers of predominantly medium to coarse sand (with some silt), and layers dominated by an organic mass. When these deposits are dug-out or cleaned, they show little detail, but in certain sections, differential erosion exposes organic-rich layers which become undercut and protrude from the face. This differential erosion is clearly shown in Figure 3.10, which shows Facies D and C exposed along the edge of the west delta, 240 m upstream.
from the river mouth. The contact between these facies is shown by the lower limit of the heavily oxidised sand, which forms the lower deposits of Facies D. The form of these deposits after excavation is seen in Figure 3.26 (p. 144).

The organic-rich layers are dominated by a mass of small root systems and marsh Carex stems, which support sands. They have ill-defined boundaries and are much thicker than organic layers previously described in Facies C, although it is almost impossible to determine their exact thicknesses. The coarse inorganic layers of this facies vary between 2 and 15 cm thick, and have numerous thin rootlets running vertically through them. The inorganic material is extremely heterogeneous - material may vary from silt to a very coarse sand, and contains parallel bedding. This bedding is seen in Figure 3.11, which shows sediments of Facies C and D sediments exposed in a block which was dug from the face. One can see from this photograph that unlike the bedding reported within Facies B, changes in sediment size between beds show no cyclicity, or rhythmic alteration from fine to coarse material. Bedding contacts may be either gradual or abrupt, and the sands occasionally contain orange mottles. Figure 3.11 also shows the nature of the contact between Facies C and D. The contact is marked by an immediate change in sediment-size from silts and clays to fine to medium sands.

As organic-rich layers are poorly defined, it is almost impossible to determine the exact number of beds within this facies. Perhaps the best exposures of this facies are seen on the large slump blocks which lie at the base of all sections along both wd and ox faces. Squamish flow differentially erodes the inorganic material, revealing a number of beds which are capped by an organic mass. At one location, 16 organic layers were counted, although this is considered to be an underestimation of the true number as there appeared to be evidence of a number of smaller organic layers within one larger organic
Figure 3.10. Typical section exposed along the lower 500 m of the west delta, showing Facies C and D. The lower limit of Facies D is marked by the lower extent of the coarse sand bed which is heavily oxidised. This site is located between sites wd5 and wd7 (shown in Figure 3.3). The coarse sand layer shown at sites wd5 and wd7 is seen in this figure as a layer which has been highly eroded. Erosion of this layer commonly causes collapse of the upper bank section. Upstream is to the right. Face is 1.5 m high.
Figure 3.11. The distinct and abrupt change in sediment type from a silty-clay to a medium sand marks the transition from Facies C (bottom) to Facies D (top). This block of sediment was cut from the face, but is not from the same location shown in figure 3.10.
mass. Despite this indication, these layers were not distinct enough to be confirmed as individual beds.

In longitudinal section along the channel bank of the west delta, this facies is lenticular in shape. It is thickest (134 cm) near site wd5, a point 200 m upstream from the river mouth. This facies decreases in thickness in both upstream and downstream directions until it is only 14 cm thick at site wd14, a point 960 m upstream. Further upstream from this site, large rhythmites are not seen in section, but the upper few cm of sediment remains slightly coarser than underlying sediment. Generally, however, Facies D sediment-size tends to fine with increasing distance upstream. There is limited information regarding the three-dimensional form of this facies (because of limited exposure), but observations in tidal channels reveal that the large rhythmites do not extend to the back of the west delta (toward the valley wall). In tidal channels near the western edge of the west delta, Facies D was not detected. Instead, sections show the characteristic rhythmic deposits of Facies C throughout.

Deposits of Facies D were noted at one other isolated location (site wd18) at the upper limit of the west delta, as shown in Figure 3.3. These deposits were 60 cm thick, but were neither laterally nor longitudinally extensive.

**Facies E: Parallel-Laminated Sands with Peds**

This facies gradationally overlies Facies C in sections along the upstream 500 m of the west delta, and along the entire length of the ox face. The material within Facies E generally is a fine to medium sand, though locally may contain considerable amounts of silt. The sediments of this facies are characterised by ped development, and so have an
open framework produced by the pore spaces between peds. The formation of peds within this material gives it the appearance of a sandy soil. It is often difficult to detect bedding within these sediments, but from a distance, very fine, wavy parallel bedding can be seen. At a number of locations, the sediments of this facies coarsen-upward from a very silty sand to a pure sand. Sediments often contain orange mottles, particularly toward the top of facies, and roots are found throughout. These roots are not those observed in Facies C, but are larger, more extensive root systems running vertically through the sediments.

This facies is interbedded with deposits of Facies F and G, which are coarser and finer than Facies E sediments respectively. Contacts between Facies E, F, and G are always abrupt. Because of this interbedding, it is difficult to assess the thickness of Facies E. If one measures from the upper and lower limits of these deposits (which includes the thickness of all interbedded facies), Facies E varies from 30 to 73 cm thick.

**Facies F: Ripple-Bedded Sands**

The deposits of Facies F are interbedded with those of Facies E and G, and form the uppermost sediments along the upper 100 metres of the west delta, and all locations upstream from this point. This facies is not observed in sections along the lower 1 400 m of Squamish estuary. All contacts are abrupt, and occasionally, lower contacts are erosional. These deposits are often underlain by a well developed root horizon (as seen at site lb1 in Figure 3.3), which may be up to 10 cm thick. Material ranges in size from fine to coarse sand. Silt is entirely absent from this facies.
This facies, shown in Figure 3.12, is comprised of numerous beds separated by 
reactivation, and more commonly, erosional surfaces. Facies thickness is highly 
variable, and may exceed 115 cm. Beds contain fine parallel bedding which may be near 
horizontal, wavy bedding, and climbing ripple lamination. Climbing ripples are seen to 
develop from near-horizontal parallel beds, and both in-phase and in-drift (McKee, 
1965) forms develop. Beds commonly contain small ground organic litter, which is 
concentrated in ripple foresets, or forms fine laminae (within which sand content is low 
by volume) within the parallel, wavy bedded sands. These organic beds may be traced 
several metres in section. At a number of isolated locations, where these sands form the 
uppermost deposits in section, sediments of this facies have been eroded from the near-
bank zone.

Along the ny face, where deposits of Facies F are thickest and dominate sections, the 
number and thickness of fine, parallel beds increases. One such example is shown in 
Figure 3.13, taken around one metre depth at site ny1. As this example shows, these 
parallel laminae may have a high coffee-ground organic content, or may mostly be clean 
sands.

**Facies G: Parallel-Laminated Silts and Clays**

Deposits of this facies are only found within the middle part of the estuary along the ox 
face. They are not found within sections along the west, central, or east deltas, and are 
not observed in section from a point 2 700 m upstream from the river mouth to a point 
well above Mamquam River confluence. The sediments of Facies G are silts and clays 
which contain very fine, parallel, near-horizontal laminations. This material is often 
detectable by its pink colour, and may contain fine laminae of coffee-ground organics.
Figure 3.12. The upper 60 cm of facies F sands at site ny1. Parallel, sub-horizontal climbing ripple lamination (in-phase and in-drift) forms are present here. Ruler is 18 cm.
Figure 3.13. Fine parallel-bedding at about 1 m depth at site ny1. Note the importance of coffee-ground organics in the bedding toward the lower half of the photograph.
Deposits of this facies are only thin - up to a maximum of 8 cm, and may not be present in section. They are found interbedded with the sands of Facies F, and may be traced (longitudinally) hundreds of metres. The upper and lower contacts of this facies are always abrupt, though never erosional.

In sections upstream from the ox face, thin units of finely-laminated materials are occasionally present, and these resemble deposits of Facies G. The material within these deposits, however, is considerably coarser - fine to medium sands.

**Facies Interpretation**

**Facies A: Cross-Bedded Sands and Gravels**

Facies A deposits likely represent migrating bedform deposits laid down by Squamish flows. This is indicated by the presence of large-scale foreset bedding produced by the down-valley migration of megaripples or dunes. The erosional surfaces within these sands were produced by bedforms migrating over an aggrading bed. As one megaripple migrates over the deposits of the preceding megaripple, it erodes the topset beds, and buries its foreset beds. Gravel lenses and interbeds were laid down during high discharge periods, and these occasionally eroded underlying sands. The contact between Facies A and B has been noted as being of variable depth in section. This variability reflects the uneven upper surface of Facies A, which was produced by large-scale bedforms, individual bars, and small channels. Sonar scans of the present channel bed at this location, shown in Figure 3.14, reveal that contemporary bedforms may exceed 1.4 m in amplitude. This amplitude is in good agreement with the variability of the relative depth of Facies A in section (1.5 m).
Figure 3.14. Sonar scan of the estuarine channel bed in Squamish estuary, toward the upstream limit of the west delta. Numbers represent depth from the water surface in feet. The largest bedform seen here is 1.4 in height. Data courtesy of Babakaiff (1993).
The height of these deposits (relative to mean sea level) indicates that they were laid down along shallowing channel sections, or intertidal sand bars. As flows were clearly competent enough to initiate the transport of gravel and megaripples over 20 cm in height, it is not felt that these deposits were deposited in a fully abandoned distributary channel. These deposits were deposited in a region where Squamish flow was actively transporting large amounts of bed material. This interpretation is also based on the lack of fines within Facies A sediments.

The mud clasts observed in these sediments are likely derived from Facies B. Their presence indicates that the channel was actively eroding sediments locally. Erosion of Facies B deposits is occurring at many locations along channel boundaries at present, and this erosion is seen to form plate-like mud clasts (Figure 3.15). Given their suggested origin, these mud clasts should contain very fine parallel laminae. This material becomes concentrated as lag at the base of channels, and is commonly noted as a characteristic feature of estuarine and tidal channel fills (van Straaten, 1959; Oomkens and Terwindt, 1960; Bosence, 1973). These mud clasts are commonly reported within the sandy foresets of large-scale cross-bedded deposits (as was the case here), within estuarine channel environments (Land and Hoyt, 1966; Van Beek and Koster, 1972; Goldring et al., 1978).

The small-scale irregularities noted on the upper surface of these sands are interpreted as micro load structures. These structures have previously been recorded within intertidal mudflat deposits in the Bay of Fundy (Dalrymple et al., 1991). The lack of evidence of subaerial exposure indicates that these sands and gravels were only exposed for brief periods of time during tidal lows, as their position relative to mean sea-level would indicate.
Figure 3.15. Deposits of Facies B being eroded by contemporary river flow. This erosion forms plate-like mud clasts which contain fine parallel-bedding. Also note the shrinkage cracks caused by the constant wetting and dewatering throughout the tidal cycle, and the colonisation of this surface by marsh grasses. The location of this site is shown by the letter B in Figure 3.3.
**Facies B: Finely-Laminated Sands and Silts**

It is clear from the description of these deposits that interpretation of their genesis cannot rely on a single model. The local variability of these deposits, which has been recorded in Figures 3.4, 3.5, 3.6, and 3.7, indicates dynamic changes in modes of sedimentation. Despite this observation, this fine rhythmically laminated bedding (Reineck, 1967) is interpreted here as tidal in origin. The bedding and sediment-size characteristics of Facies B deposits are very similar to intertidal deposits described by many authors, but particularly the tidal bedding of Reineck and Singh (1980). The silt and clay laminae were deposited during high tide periods. The term tidal standstill has been avoided here, as it is not felt that current velocities were necessarily at or near zero. The coarser laminae were laid down during periods of limited tidal modification of flow. These periods are thought to include the periods of tidal drop, and most of the rising limb of the following high tide. Deposition of silt and clay need not be limited to brief periods of high tide, but may also occur during periods of tidal current activity, as shown by Kirkby and Parker (1983). With the exception of periods of high river stage, deposition will not occur during low tide, as intertidal sand surfaces will be exposed above the water line. While fluctuations in current velocity are capable of producing finely-laminated sand/silt bedding, this is not thought to be the case here. I think that the alternation of sand and silt is too rhythmic, and facies thickness too great (up to 182 cm) to have been produced by current activity alone.

Unfortunately, because of the limited exposure of these sediments, one cannot determine the exact nature of tidal rhythmicity recorded by these silt and sand laminations. It is postulated, however, that these sediments record neap-spring tidal cyclicity. If this is correct, then each of the darker, coarser beds seen in Figure 3.7 represent deposition
over the tidal neap, while dominantly light, finer beds represent deposition over the tidal spring. During tidal neap periods, tidal flux is least, and as such, tides are unable to modify Squamish flow to a great extent. This results in the deposition of thicker, coarser beds of sand produced by Squamish River current activity. During tidal spring periods, tidal flux is greater and exerts more control on river flow. During tidal rise and peak, flow velocities are reduced enough to allow the settling of silt and clay from suspension. The greater degree of tidal modification of flow is reflected in the greater concentration of fines in sediments. Decreased energy conditions are also indicated by the reduction of sedimentation rates over this period (lighter beds are thinner than darker beds). If this interpretation is correct then the annual sedimentation rate at the location shown in Figure 3.7 would be around 100 cm. This in turn would mean that these intertidal sands were deposited over a two year period. These figures assume that sedimentation rates remain roughly constant over the year - an assumption which is not necessarily correct, as will be discussed later. Although this rate of sedimentation appears very large, it is not thought to be an unreasonable rate of deposition from this high energy, sediment-laden river. Lanier et al. (1993) examined tidal bedding with preserved neap-spring rhythmicity in ancient sequence which they interpret as tidal flat influenced by diurnal tidal cyclicities and a high input of sediment from a fluvial source. At one location, approximately 180 cm of tidal flat sands were deposited over a period of one month. In less extreme cases calculated average sedimentation rates are around 380 cm/yr (Lanier et al., 1993). Similarly, Dalrymple et al. (1991) estimate that the maximum sedimentation rates within Bay of Fundy upper tidal flats are approximately 700 cm/yr.

As has been stated, there is insufficient data to test this hypothesis, but there is limited information to suggest that it is correct. At one location, shown in Figure 3.7, it was possible to count the number of alternating sand and silt/clay laminae within one of the
dominantly dark beds. Within this bed (shown more closely in Figure 3.16), 28 layers were counted in total (14 sand and 14 silt/clay layers). This is in excellent accordance with the theoretical number (28.5) of rising and falling tides over a one week period within a semi-diurnal system such as this (Visser, 1980). It is suggested that the dark bed shown here represents deposition over one neap tidal cycle (one week). The nature of the mixed, semi-diurnal tidal system which influences Squamish estuary is shown in Figure 3.17. Over this particular neap-spring period, 29 high tide periods were recorded. Figure 3.18 shows the nature of daily tidal fluctuations, and the proposed depositional product of this flux. This figure shows that silt/clay laminae become deposited at and around daily high-high tide and low-high tide, and that cyclic variations in tidal stage produce rhythmically alternating sand and silt/clay layers. In this manner, the 14 fine, light coloured laminae within the dark layer in Figure 3.16 (which is 2.5 cm thick) were deposited over a 7 day period.

Figure 3.18 also shows the approximate location of the upper surface of Facies B in relation to different tidal heights. One can see from this that sands become exposed over a period of time surrounding low-low tide. This exposure produces a period of time during which no deposition can occur, but when erosion could presumably occur. During the tidal drop and initial stages of tidal rise immediately before and after this period of non-deposition, sand will be transported by Squamish flow. If this supposition is correct, then every other sand layer should comprise material which was deposited over two distinct periods of time separated by a number of hours. Extremely detailed analysis of this fine bedding may be able to detect a reactivation surface, or possibly an erosional surface within alternate sand beds.
Figure 3.16. Close view of Facies B deposits previously seen in Figure 3.7. In the centre of this photograph one of the dominantly dark beds (which is 2.5 cm thick) shows very fine parallel lamination. 28 distinct groups of laminae were counted here.
Figure 3.17. Neap-Spring semidiurnal tidal cycle which influences Howe Sound fjord and Squamish River flow. Within this particular 14 day period from June 12 to June 26 1991, there were 29 tidal highs (both high high tide and low high tide periods).
Figure 3.18. A portion of the semidiurnal tidal system experienced in Squamish estuary. Below the tidal curve are shown the proposed depositional products of daily stage fluctuations. One can see that these fluctuations produce rhythmic alternation of sand and silt/clay laminae.
It is clear from the evidence presented earlier, however, that not all deposits of Facies B appear the same as those shown in Figures 3.7 and 3.16. For example, the light and dark beds exposed in Figures 3.4 and 3.5 are much thicker than those deposits described above. As an example, the deposits shown in Figure 3.5 appear to contain five or six fining-upward units. From the interpretation given above, these sediments would represent deposition over 10 or 12 weeks, representing an annual sedimentation rate between 160 and 190 cm. The work of Lanier et al. (1993) shows that this is not an excessively large sedimentation rate. It is clear from this that the sedimentation rate of Facies B deposits displays a considerable degree of local variability. Local differences in sedimentation rate and the consequent character of tidal bedding may be produced in a number of ways. The thickness of alternating beds within this fine rhythmically laminated bedding most likely will vary with distance from the active channel, and with season. Deposits of Facies B which develop in channel proximal locations will reflect the greater relative importance of fluvial processes. At these locations, sedimentation rates will be higher than in distal channel locations, and tidal modification of flow velocities will be less. This will be evident in sedimentary sequence, as proximal channel deposits will contain coarser sediment within thicker beds. In addition, fine silt/clay laminae will be less evident and less important, and may occasionally be eroded. It is suggested that this was the depositional environment for the deposits of Facies B seen in Figure 3.4. In contrast, deposits which are laid down in distal channel locations will reflect a greater relative importance of tidal flux. These deposits will be finer, contain thinner beds, and include numerous rhythmic fine-grained laminae separating coarser laminae, as seen in Figure 3.7.

These differences in the character of Facies B deposits could also be produced by the seasonality of Squamish flow. During summer months, Squamish flows are greatly
increased by the freshet, and suspended-sediment concentrations are at their highest (with the possible exception of large winter flood events). Given this nature of flow during summer months, one would expect summer deposits of Facies B to differ greatly from those laid down in winter months. During winter, fluvial processes and inputs are far less dominant (except during large floods), and as such, tidal flux should exert a greater degree of sedimentary control. This may be reflected in sedimentary sequence, as winter deposits likely would be characterised by thinner deposits of finer-grained material. This may be enhanced by the presence of salt-wedge intrusion, which may allow the build-up of suspended-sediment concentrations along the turbidity maximum.

It is possible that the deposits of Facies B seen in Figure 3.5 were produced by this seasonality of Squamish flow (or seasonality of sedimentation) described above. These sediments may not display neap-spring tidal cyclicity, but may simply record seasonal changes within an annual deposit. Each of the five or six fining-upward units reported earlier may have been produced over a five or six year period. Under this scenario, high summer discharges deposit thick deposits of coarse material, with fine silt interbeds produced by tidal flux. In winter months when discharges are low, tidal depositional processes become increasingly important, and energy conditions decrease, producing thinner, finer deposits. In my opinion, this interpretation is less appealing than the interpretation of deposition over neap-spring tidal cycles, despite the fact that this would necessarily produce high annual rates of sedimentation. Recorded rates of contemporary sedimentation within estuarine intertidal sand environments give some weight to this assertion. Annual rates of sedimentation (over 25 cm) within the intertidal sand zone along the seaward edge of the west delta are over three times greater than the thickest of the five or six units seen in Figure 3.5.
In conclusion, Facies B deposits display fine rhythmically laminated bedding (Reineck, 1967), which I believe may more accurately be termed tidal bedding (Johnston, 1922). Further examination likely will show that neap-spring tidal cyclicity is preserved in the sediments of Facies B throughout much of the lower estuary. The preceding discussion has shown that these deposits display a great deal of variability throughout Squamish estuary. Similar variability was reported from the intertidal sands and muds of the Cobequid Bay-Salmon River estuary, Bay of Fundy (Dalrymple et al., 1991). These authors logged sections of intertidal deposits on a millimetre scale, and were able to recognise sedimentation produced by individual flood and ebb tides, diurnal inequality of the tides, neap-spring tidal cycles, and annual deposits. It is clear that the deposits of Facies B need to be examined at an extremely fine scale if precise and conclusive proof of the nature of tidal influence is to be found. Such fine scale analyses have previously been employed with good success by Kasse (1986) and Dalrymple et al. (1991).

The presence of preserved ripples within Facies B deposits may be an important indicator of the environment of deposition of these sediments. The presence of ripple forms which were clearly migrating against Squamish River flow may provide clear evidence of the estuarine setting of these deposits. Evidence of reversing current activity is one of the key indicators of tidal environment. Unfortunately, the fact that such features are present is not unequivocal evidence of reversing currents, as these ripples could have been formed by wave action in shallow waters. It is not unreasonable to expect that wind-generated waves could induce sediment-transport over intertidal sands which are inundated by shallow waters during late tidal drop or early tidal rise.

Unfortunately, ripple forms produced by wave and currents are not always simple or even possible to differentiate. While wave ripples are commonly assumed to be typified
by symmetrical crests, they often have an asymmetrical form (Tanner, 1967). Also, while the majority of wave ripples display bi-directional, chevron-like laminae (Reineck and Singh, 1980), Newton (1968) showed that the majority of wave ripples in the nearshore environment displayed unidirectional foreset laminae. These are produced when flow velocities in one direction exceed those in the opposing direction. Where asymmetrical wave ripples form with unidirectional foreset laminae, they will be difficult to distinguish from current ripples with straight crests (Reineck and Singh, 1980). Tanner (1967) developed a number of indices to differentiate asymmetrical wave and current ripples, and these are tested below.

Of the indices reported by Tanner (1967), the ripple index (RI) and the ripple symmetry index (RSI) are probably of best interpretive value, and the simplest to compute. The RI is calculated by dividing the ripple wave length by the ripple height. The RSI is calculated by dividing the straight line distance from the highest point on the ripple crest to the lowest point in the trough, along the lesser slope, by the same distance along the steeper slope. Calculated dimensionless RI values for the two asymmetrical ripples shown in Figure 3.6 are 10.8 and 11.3, while RSI values are 2.3 and 3.95. These values may be compared with those regions which distinguish between wave-formed and current-formed ripples, as highlighted in Figure 3.19.

One can see from this diagram that both the RI values lie within the overlap region between wave and current ripples. Of the RSI values, one is classified as of wave origin, and the other is classified as of current origin. Unfortunately, none of the ripple forms displayed any of the features which Evans (1949) considered unique to either wave or current formation (such as double ripple marks or secondary ridges within ripple troughs). These data are inconclusive, particularly given the limited number of ripple
measurements. The other 11 ripples preserved in the ripple train seen in Figure 3.5 were not measured, as it was felt that their symmetrical forms were not produced by original depositional processes, but by reworking before burial.

In conclusion, it is impossible to determine whether the ripples observed in this study were formed by reversing tidal currents within Squamish estuary, or by wind-generated waves moving across inundated intertidal sands. All that is clear is that these ripples formed under very shallow water conditions (a few centimetres (Singh and Wunderlich, 1978)).
Lastly, the highly variable thickness of this facies likely reflects differences in the relative height of the upper surface of Facies A in section. The upper limit of Facies B is determined by the elevation above mean low tide which allows vegetation colonisation. Therefore, the thickness of this facies is determined entirely by the depth of the lower contact with Facies A.

**Facies C: Silty Rhythmites**

The rhythmite deposits of Facies C clearly are tidal marsh rhythmites. Over the course of one year, these deposits can be seen forming in the tidal marsh environments of the west, central, and east deltas of Squamish estuary. These deposits have previously been described in some detail by Bouma (1963), Evans (1965), and Redfield (1972). In this environment, the inorganic layers are thought to be deposited primarily during combined flood and high tide periods. The majority of this deposition occurs during summer months, when tidal amplitudes are greatest, river flow is greatest on average, suspended-sediment concentrations are highest, and when vegetation cover is greatest. From the nature of bedding within these inorganic layers, deposition appears to have occurred over many flood events. This statement is supported by sediment accretion data collected over the summer of 1992. Sediment deposition was recorded within the west delta using a number of marker beds (orange spray and small plates), over a number of time scales. Marker beds placed within the lower marsh recorded thin, discontinuous films of sediment which was deposited over one tidal cycle. These markers became completely covered after five days.

The flat organic mats are produced by the autumn collapse of dead marsh vegetation cover. This vegetation lies flat on the delta surface, and gradually becomes covered as
the delta accretes. In a number of locations on the central delta these organic layers became exposed due to erosion of the overlying material. On these occasions, *Carex lyngbyei* stems were pointing up estuary (with their root systems down estuary). In this environment, these sedges are less likely to have been laid down by tidal currents than by strong up valley sea-breeze winds, so they should not be used as palaeoflow indicators. This marsh vegetation death and burial, which has been observed in Squamish estuary, produces annual rhythmite beds within tidal marsh deposits. This interpretation of stratigraphic rhythmicity produced by annual sedimentation was also made by Redfield (1972) and Yeo and Risk (1981), although these authors provided different interpretations of the exact processes and seasonality of sedimentation.

The upsection decrease in bed thickness recorded here has previously been described as an upsection increase in organics (Yeo and Risk, 1981). From an examination of the organic-C content of the deposits of Facies C (shown in Figure 3.20), this trend can be seen. The uppermost 16 cm of sediment, which record a decreasing organic content, are associated with deposits of Facies E. I think that it is important to note, however, that any increase in organics is relative, and driven by a decrease of inorganic input to the marsh system, rather than increasing vegetation density or biomass production.

In addition to this, decreasing bed thicknesses can result in an incorrect record of increasing organic content upsection which is induced by sampling technique. If a fixed sampling thickness is used, more organic layers are sampled at the top of the facies than are sampled toward the bottom, resulting in an apparent increase in organic content. The nature of this decreasing sedimentation rate over time likely is produced primarily by the nature of tidal flux within this estuary. In other words, tidal marsh sedimentary sequences record decreasing durations of tidally-driven inundation, and decreasing
Figure 3.20. Changing organic-C content (by dry weight) in marsh sediments with depth below the delta surface at site irs1.

depths of the inundating water column. If this is the case, then tidal marsh deposits may be identified as tidal in origin from an examination of bedding characteristics alone. This hypothesis is tested in chapter four. The decreasing visibility of rhythmite beds upsection may not simply be related to decreasing sedimentation rates, however. It may also reflect an increasing degree of bioturbation with increasing height of the delta, which is caused by increasing durations of subaerial exposure.

The transitional zone between Facies B and C, where the sediment is sandy, but fine organic layers can be seen, can be related to the colonisation phase described by Redfield (1972). Redfield’s model is discussed in greater detail in chapter five, but essentially, this is the stage at which halophyte vegetation spreads onto intertidal flat
surfaces. During these early years, vegetation cover is not yet dense enough to modify flow, so deposits remain coarse. As vegetation density increases, flow becomes increasingly modified, which is seen in sedimentary record as an increase in fines upsection, and an increasing thickness and continuity of organic layers.

The transitional zone toward the top of Facies C, where sediment fines and bedding becomes too fine to count individual beds, represents the transition from low to high marsh (Martini, 1991). This transition is produced by decreasing frequency and depths of tidal inundation, and accordingly, less saline conditions. These changes permit a vegetation succession to less saline-tolerant grass and shrub.

Rhythmites in areas close to tidal channels appear different from those exposed along river banks for two reasons. Firstly, the sediment seems finer throughout the facies because of increased supplies of fine sediment which are brought into the delta environment by tidal channels. In addition to this, with increasing distance from Squamish River, flows are less able to transport sands. Secondly, rhythmites appear different because organic layers differ from those elsewhere as they have less distinct boundaries. This difference is perhaps related to changing vegetation populations as conditions become more saline toward tidal channel margins.

This last point highlights the fact that tidal marsh rhythmite beds display a great deal of local variability which is seen through changes in bed thickness, and internal character (sediment-size characteristics). These noted localised differences are primarily driven by the location of these deposits in relation to Squamish River (at the time sediments were laid down), and also the morphology of the river at different locations. As we have already seen, rhythmites change in appearance with increasing distance from the
channel. Deposits should become finer away from the channel, and with the exception of those areas close to tidal channels, deposits should also thin away from the main channel. Palaeochannel morphology likely is also an important factor which will determine the character of tidal marsh rhythmites. As an example, deposits of tidal marsh which develop along the outer bend of river meanders should be both thicker and coarser than those deposits which form along the inner bend of meander loops.

Finally, it has been noted that rhythmite beds display a wavy or undulatory character over distances ranging from metres to tens of metres. This longitudinal variability of bed depth in section is in part inherited from deposits of Facies B. As we have seen, Facies B deposits are of variable thickness and height relative to mean sea-level. The variable upper-surface form of these deposits was produced by large bedforms, bar complexes, and small tidal channels. As the sediments of Facies C are deposited almost entirely from suspension, these fine sediments become draped over an undulating surface, and they will tend to retain that uneven form. In addition to this, the wavy nature of beds will be further produced within the deposits of Facies C by the local variability of marsh sedimentation rates. These local differences in sedimentation rates are greatly influenced by distance from the river, small-scale topographic variability, and differences in the density of marsh vegetation cover.

The uneven form of the upper surface of the marsh is seen particularly well in sections along the ox face (Figure 3.21). It is highlighted by a dark band (which is either a layer high in organic matter, or is a charcoal layer) which can be traced along the majority of the ox face. It is important to note, however, that despite the fact that these beds are dominantly parallel and wavy, this is not wavy bedding. Wavy bedding is produced by the deposition of ripple-bedded sand, which is then covered by a mud drape (Reineck
Figure 3.21. The characteristic form of sections along the ox face. The undulatory or wavy nature of bedding is emphasised here by the dark layer at around 1 m depth in section. Above this, deposits of Facies E are overlain by a dominant root horizon and deposits of Facies F. This face is over 2 m high.
and Singh, 1980). It is characteristic of tidal environments (though may be found elsewhere), and can be considered a transitional phase between lenticular and flaser bedding (Reineck and Wunderlich, 1968).

The bedding form of Squamish estuarine deposits is often very similar in appearance to climbing ripple lamination in-phase (McKee, 1965). It is clear, however, that although the mechanism of deposition is similar to that of climbing ripple lamination, the bedding in Squamish estuary is different. Firstly, both the wavelength and amplitude of wavy beds are considered too large to be climbing ripple laminae. More importantly, the time scales of deposition between these two forms are quite different. Climbing ripple laminae are formed within one flow event which is characterised by conditions of constant flow and sediment supply. In contrast, the wavy beds in this study may develop over periods exceeding 100 years, under conditions of highly variable flow. Given the character of these beds, it is proposed that the term 'undulatory bedding' be used, as this is explicit, yet does not imply any genetic association which could lead to misinterpretation.

While all the above statements are considered correct, these are not considered the only factors which produce or induce this undulatory bedding. Bedding form is also thought greatly influenced by post depositional modification of these sediments. The tidal marsh deposits likely have settled and slumped considerably in response to the continual wetting and dewatering of sediments throughout the tidal cycle. This settling process is considered particularly important in areas close to river banks, where water flux likely is greatest. Differential slumping may also be induced by gas build-up and escape within these sediments. These gasses are produced in considerable amounts in interdistributary bay environments as organics are continually decomposing.
Facies D: Sandy Rhythmites

Facies D deposits are genetically related to those of Facies C. Each of the beds (capped with an organic mass) within this facies represent deposition over a period of one year. The differences between the rhythmite beds of Facies C and D are their thickness and sediment-size. These two characteristics clearly are related, and indicate an increase in energy environment from Facies C to D. The abrupt contact between these facies, highlighted in Figure 3.11, indicates that the change in depositional environment was rapid. It is proposed that the inferred increased flow velocity along the lower 800 m of Squamish estuary (the extent of this facies), was caused by river training dyke construction. This construction has led to the formation of an anomalous coarse-grained sedimentary sequence within the west delta tidal marsh.

The anomalous sequence (the superposition of Facies D on Facies C) primarily has been produced by the reduction of effective channel widths along the lower 5500 m reach of Squamish River. Toward the southern limit of the west delta, dyke construction has not led to a reduction in channel width since part of the west delta was dredged to maintain the pre-dyke channel width along this section. Despite this, channel widths have essentially been reduced along the lower 3500 m channel reach because the river training dyke prevents the flow of water through the central distributary channel during flood events. This forces all flow through a single channel to the west of Squamish delta. In addition to the direct reduction of channel widths, the river training dyke has greatly reduced floodwater storage capacity, as floodwaters are no longer able to inundate the central and east deltas. Along the lower 1700 m of Squamish River (the extent of the west delta), effective floodplain widths have been reduced by an average of over 68%. The combined effect of this channel and floodplain constriction has been an increase in
water depth, and duration of inundation over the west delta. Presumably, decreasing widths have also led to increased flow velocities through this lower estuarine reach. The combined effects of this constriction have led to an increase in flow velocities of waters entering the marsh system, which also becomes inundated to greater depths than previously. These increasing velocities are evidenced by the rapid coarsening of tidal marsh deposits. This coarsening-upward trend is in contrast to the fining-upward sequence which develops in natural marsh systems. This natural sequence, highlighted in exposures along the ox face and upper west delta, displays a gradual thinning of bedding, and decreasing sediment size with increasing maturity of the marsh system. Sediment sequences only show coarsening-upward characteristics with the transition from delta plain to alluvial plain (transition from Facies C to E - described later).

If the above interpretation is correct, then the contact between Facies C to D should date to 1972. In the description of Facies D deposits, it was stated that at least 16 organic layers were detectable in section and slump blocks, although this was regarded as an underestimation of the true number. If these deposits are genetically related to those of Facies C, then one organic-rich layer should be laid down each year. In the summer of 1992 there should have been a total of 20 organic-rich layers within the deposits of Facies D (no layer would have existed for 1992, as plant death does not occur till late autumn). A test of the age of Facies C and D deposits will be discussed later in this chapter.

From an examination of the internal bedding characteristics of these deposits, shown in Figure 3.11, it is felt that beds form over a considerable period of time, and that sedimentation occurs over a number of events of differing intensity. This interpretation agrees with that made regarding the genesis of the rhythmite beds of Facies C. As the
sediments within Facies D are very sandy, deposits are well-drained and become partially aerated during low tide. This aeration allows partial oxidisation of sediments.

In longitudinal section, Facies D is lenticular, as deposits thin upstream and downstream from point near site wd5. This upstream thinning is related to the height (age) of the marsh surface. Generally, with increasing distance away from the seaward limit of the delta, the elevation of the marsh increases. With increasing elevation, the marsh becomes less frequently inundated by combined flood and high tide flows. This decreasing frequency and duration of inundation is reflected in the upstream thinning and fining of Facies D. This controlling nature of marsh height can be shown from an examination of the small, isolated deposits of Facies D located toward the upstream most end of the west delta. At this location, deposits of this facies are 60 cm thick, and are shown in Figure 3.3 at site wd18. Their presence in part of the estuary where Facies D is otherwise not found, is due to the height of the delta/floodplain surface at this location. These deposits are forming on an emerging bar surface which has largely developed over the last 30 years. As this surface is much younger than surrounding deposits, it is lower (up to one metre), thus allowing the localised formation of Facies D deposits. The downstream thinning of Facies D deposits perhaps reflects decreased riverine influence close to the river mouth. With time, the zone of maximum facies thickness will shift seaward.

In lateral section, Facies D is wedge-shaped, being thickest toward the river, and thinning toward the valley wall. This thinning likely arises from the decreasing flow velocities of floodwaters toward the centre of the marsh. Much of the sand component of the suspended-sediment will settle to the delta surface within a few metres of the river. Toward the back of the delta, flows are only competent to transport small amounts of
fine sand. As such, the sandy rhythmites of Facies D should grade laterally into the silty rhythmites of Facies C.

From the above discussion of facies geometry and sediment-size characteristics, it seems clear that Squamish River is building a levee along the lower 800 m of the west delta. This levee is represented in section as the sediments of Facies D. This rapid increase in vertical marsh growth is a direct response of river training in Squamish estuary. The effects of this river training with regard to channel stability and delta erosion will be discussed in chapter five.

This levee can be seen on the surface of the west delta in Figure 3.22, as marsh surface elevation differences are highlighted by the partial flooding of the marsh. It is important to note from this photograph that there also exists a levee on the central delta, along that portion of the channel which is now artificially removed from Squamish flow. This shows that levee building was operating naturally prior to 1972. The important difference between pre- and post-dyke levees, however, is their sedimentologic character. The levee on the central delta is produced by the streamward thickening of the rhythmite beds of Facies C. The sedimentologic character of these beds is essentially the same throughout - no coarse, thick sedimentary units are found along the upper sequence of the central delta.

**Facies E: Parallel-Laminated Sands with Peds**

The deposits of Facies E represent increasing energy conditions associated with decreasing tidal influence as the delta plain surface accretes. The separation of this facies from Facies C is based on the change in sediment-size characteristics and organic
Figure 3.22. Squamish west and central deltas during flood. Differential submergence of the deltas reveals 2 levees.
content. Facies E deposits essentially represent the initial transition from delta plain to alluvial plain.

The bedding within this facies is the same as the bedding within the upper limits of Facies C. These very fine parallel laminations suggest that the main mechanism of deposition (the settling of material from suspension) remains the same through the Facies C to E transition. This gradual accumulation of material produces bedding which parallels the form of the marsh surface, resulting in the formation of undulatory bedding. With increasing height of the land surface, tides less frequently inundate the marsh surface, and are only able to influence sedimentation during periods of high river flow. These higher discharges introduce coarser material to the marsh/interdistributary bay environment. The increase in sand content, combined with decreasing periods of inundation, permit the establishment of less saline-tolerant (non-halophytes) vegetation cover. These plants (which may include shrubs and trees) do not necessarily die during the winter, so flat organic-rich layers (as seen in Facies C) are not produced. Instead, plants are able to spread their root systems throughout the sandy substrate. As sediments also become more aerated, mottles begin to develop as sediments become partially oxidised. As the land surface is raised further, pedogenic processes begin to operate, resulting in the aggregation of material into soil peds.

**Facies F: Ripple-Bedded Sands**

The sands of Facies F were deposited during large riverine flood events. The presence of climbing ripple lamination indicates that deposition took place from both bed load and suspension from waters which inundated this floodplain surface. The changing character
of bedding within this facies is an indication of increasing and decreasing energy conditions produced by the waxing and waning floodwaters. This is noted as changes from lower flow parallel lamination to climbing ripple lamination of all forms. This may have been partly controlled by tidal flux, though these sands show no conclusive sedimentologic influence of tides. Floodwaters occasionally were powerful enough to erode the upper deposits of Facies E.

Despite the fact that Facies F deposits dominate the upper sequence of sections upstream from the west delta, they are believed to have been produced by a small number of low frequency events. For example, the uppermost unit of Facies F along the ox face is believed to have been deposited during one flood event in October 1984. This flood, which had a recurrence interval of around 35 years, is thought to have exceeded bankfull capacity for at least three consecutive days (Hickin and Sichingabula, 1988). This duration of bankfull exceedance will have been even greater within Squamish estuary, though no data are available. The upper unit of Facies F is up to 40 cm thick in places along the ox face. Further upstream, the thickness of this sand unit (channel bedform deposits) increases to at least 115 cm (at site ny1). Along the ny face (around 3 300 m upstream from the river mouth), the floodplain surface is gently undulating, reflecting the transport and deposition of large sandwaves (dunes) across the floodplain. These sandwaves are up to 115 cm in amplitude, and have wavelengths of around 30 m. Deposits are around 15 cm thick at the dune troughs (as at site ny2). These deposits are underlain by a dominant root horizon which is believed to represent the pre 1984 floodplain surface.

Similar bedforms (though much smaller) are observed on the floodplain surface for another 1 800 m downstream from this point. The decreasing thickness of Facies F
downstream partly reflects the dampening of riverine flood flows by tidal modification. In addition to the decreasing tidal influence with increasing distance upstream, the upstream thickening of Facies F reflects the decreasing effective floodplain width upstream. The term 'effective floodplain width' is used here because overbank flows are laterally confined by the river training dyke, which has effectively reduced floodplain width by an average of 60%. This average value was calculated from 10 cross-sections along the lower 3 500 m of estuarine reach of Squamish River. Pre-dyke widths were measured as the width from western valley wall to the railroad track which services Squamish Forest Products facilities on the east delta. This track, which was completed in the 1920s, is built on a raised dyke to provide flood protection for the town of Squamish. It should be noted that the discussion of floodplain width reduction does not apply to flows which exceed the height of the river training dyke.

Where sand dune heights exceed one metre (3 300 m upstream from the river mouth), effective floodplain widths average 540 m. These dune heights decrease with increasing distance downstream, to the point where they are between five and ten cm high. At this location, effective floodplain widths average 590 m. This increase in width is also responsible for the decrease in sand dune amplitude.

**Facies G: Parallel-Laminated Silts and Clays**

These sediments were deposited from suspension from slow moving shallow waters over the floodplain surface. The depth of the water column over the floodplain surface may only have been a few cm. Such small depths would allow silts and clays to be transported onto the floodplain, but little sand. Bedding form evidence suggests that these deposits formed over a great number of events during high tide periods over
summer months (when riverine flow is higher, as an average, than in winter). On numerous occasions throughout the period of data collection, high tide stage just exceeded the upper limit of these facies along the ox face. If the floodplain elevation had not been raised by deposits of Facies F along this face, the fine sediments of Facies G likely would have continued to accrete. As the floodplain elevation increases, the likelihood that Facies G sediments will be deposited and present in section decreases to zero. This is because as the floodplain is raised, greater stages are required to permit deposition of any material on the surface. Further stage increases, however, will be driven by increased riverine discharges. These increased discharges will reduce the effect of tidal modification of estuarine flow, making the deposition of fines less and less likely.

Despite the fact that the interpretation above states the importance of tidal flux to the deposition of these sediments, there is no unequivocal sedimentary evidence that these are tidal as opposed to fluvial deposits.

Discussion

Depositional Model of Estuarine Deposits

The purpose of this section is to reconstruct the probable depositional environment of Squamish estuarine sediments described above.

Generally, I feel that Squamish estuarine sedimentation primarily is driven by distributary channel abandonment and fill. Continued shifting of distributary channels results in a series of discontinuous fills which may interfinger older channel fill deposits. With the possible exception of the lower 1 000 m of delta deposits, the majority of
estuarine sediments described above were laid down within abandoned distributaries. This abandonment is unlikely to have been immediate, but instead was characterised by a gradual decrease in riverine flow (and an increase in tidal influence) as the channel aggraded. This form of abandonment occurred between the years 1945 to 1972 (documented in chapter one), before total abandonment was artificially forced by dyke construction. The above interpretation is based on evidence provided within a number of sedimentary deposits. This evidence is given below, along with discussion of the sedimentary processes operating at the time of deposition.

Facies A sediments are Squamish River channel lag deposits. The height of these sediments (relative to mean sea-level) indicates that they were laid down within shallowing channel sections, or intertidal sand bars. These deposits can be seen forming in the estuary at present, along channel margins along the eastern edge of Squamish River (highlighted by the letter A in Figure 3.3). They are formed by channel migration and shallowing of the inner bank zone. Deposition along channel margins, however, will not produce laterally or longitudinally extensive deposits of Facies A sands and gravels.

For this reason, the formation of Facies A within shallowing channel margins is considered to be a minor, secondary component of topset delta deposit formation. Instead, the formation of Facies A within aggrading distributary channels is of more sedimentologic importance. Given the dynamic nature of channel shifting within Squamish delta environment, it seems most probable that channel aggradation is initiated by distributary channel shifting. This channel abandonment and filling will produce elongate fills which may be several kilometres long, and (judging from present channel dimensions) over one tenth of the total valley width.
The upper surface of Facies A lies just above mean low tide level. It appears then, that above low tide level, flows are incapable of transporting bed material as megripples or dunes. This decreased competence above this level is presumably caused by tidal modification of flow. The abrupt contact between Facies A and B suggests that an immediate decrease in energy conditions took place, which was perhaps induced by channel avulsion. There is evidence to suggest, however, that immediate changes in sediment-size need not necessarily be produced by channel avulsion. Along the eastern edge of Squamish River, intertidal sand bars are building along the lower 1,700 m of the estuarine reach. In places, these sands have become vegetated, and when they are exposed in trenches, one sees an abrupt contact between coarse sands and overlying finer sands. This sequence is directly analogous to the transition from Facies A to B noted elsewhere, and has clearly not been formed by channel avulsion. Instead, this transition is thought to be produced as the sediment surface continues to accrete above low tide level. Above this level, tidally modified flows are no longer competent to transport large sandy bedforms, and instead, accretion of finer material continues mainly from suspension. The transition from Facies A to B is not always this clear, as large floods may be strong enough to initiate bedform movement, and introduce coarse sand into the finer sediments of Facies B. This process is considered to have occurred at site wd3, shown in Figure 3.23. Due to limited exposure, the extent of this sand interbed is not known.

As has already been noted, few comments may be made regarding the bedding characteristics of Facies A sands and gravels. Despite this, one can conclude that large-scale cross-bedded units (Allen's ECS) will be present within these channel fill deposits, but will not dominate sequences. This is because channel migration and point bar development is not considered to be the dominant process of within-channel
Figure 3.23. The upper surface of Facies A coarse sands seen here (at site wd3) is abrupt, and overlain by the silts and fine sands of Facies B. The coarse sands contain a fine interbed of Facies B sands and silts. This change represents an interruption in the deposition of Facies B produced by a large flood which reinitiated channel bedform movement at this location. This flood deposited the upper unit of coarse sand. Trowel is 25 cm for scale.
sedimentation in this environment. Where such cross-bedded units are present, they almost certainly comprise clean sands with occasional gravel interbeds throughout. Given the flow regime of Squamish River, clay layers and drapes likely will not be present in these sands. Facies A deposits are most likely to be characterised by both trough and planar cross-bedding with numerous erosional surfaces, gravel interbeds and clasts, and occasional mud clasts.

With continued aggradation of these channel sands, the upper surface eventually becomes increasingly exposed to subaerial conditions. Sands eventually become exposed long enough to allow colonisation by halophyte plant species. Once this vegetation becomes fully established it decreases energy conditions at that site, promoting the deposition of finer sediment. In this way the intertidal sands of Facies B grade into the tidal marsh deposits of Facies C. At several locations within Squamish estuary today, one can see the initial stages of vegetation colonisation of intertidal sands, as shown in Figure 3.24. This photograph was taken along the eastern edge of Squamish River, in the location shown by the letter A in Figure 3.3. It documents the early colonisation process, which will eventually be preserved in section as the transitional zone between Facies B and C, where very fine, discontinuous organic layers are present, but where material remains sandy. There are a number of other interesting features in this photograph. Firstly, one can see the highly irregular surface of this young tidal marsh system, which slopes down from left to right and also toward the viewer. Also, to the bottom right one can see evidence of the fine laminae within Facies B sediments which have been partially eroded. It also illustrates how large organic debris may become incorporated into Facies B deposits.
Figure 3.24. Contemporary colonisation of intertidal sands along the eastern flank of Squamish River, at the location indicated by the letter B in Figure 3.3. This colonisation produces the stratigraphic transition from Facies B to C. To the right the picture, large wood debris is being incorporated into the sands. This view is looking up estuary.
During the formation of the tidal marsh, erosion of these marsh sediments by tide- and wind-generated waves is not considered an important process or a common occurrence. Field observation suggests that although large surface disturbances are common within the estuary, these are restricted to estuarine channels, and do not appear to affect the interdistributary bay environment. Also, on no occasion was any evidence of wave reworking or erosion of marsh sediments observed in section. While it is likely that large marine storm events will influence estuarine sedimentation and may induce delta erosion, this study found no sedimentologic evidence of this having occurred within Squamish delta.

This depositional model of Squamish estuarine deposits is not appropriate for the entire estuarine reach. The deposits of the downstream-most 1,000 m may not be estuarine channel fill sediments. Instead, the deposits of Facies A and B within this region may have been produced as Squamish delta prograded into Howe Sound. There is some sedimentologic evidence which indicates that these facies represent extensive intertidal sandflat and sand bar complexes which were laid down within an interdistributary environment at the head of the delta. The main sedimentologic evidence is the presence of shells along the lower reaches of Squamish estuary. These shells were originally deposited in a lower energy environment, away from the river mouth. In these environments, the high energy freshwater input will be less dominant, and tidal influence will be more apparent. Because of this, shells become transported, broken, deposited, and eventually incorporated into the sands away from main channel flow. In addition to this, some of the deposits of Facies B may be too tidal in character to have been deposited within an active river channel. Those deposits shown in Figures 3.7 and 3.16 appear to contain very high amounts of clay, which would not ordinarily be expected in this mostly subtidal environment of Squamish River. This evidence,
however, is by no means conclusive proof of an interdistributary bay environment. As has already been discussed, the different appearance of deposits at this particular location may have been controlled by distance from channel flow, or seasonality of estuarine sedimentation.

With the exception of the abandoned central channel, there are few expanses of intertidal sands in Squamish delta at present. There are a number of reasons for this. Firstly, intertidal sands toward the front of both central and east deltas are regularly dredged to prevent siltation of Squamish harbour. Secondly, artificial constriction of flow to the west of the delta has largely cut-off the supply of sand to the rest of the delta, so sand bars are largely restricted to the western sector (Levings, 1980). Despite this, it is clear that large expanses of intertidal sands have existed along the seaward margin of the delta within recent times. One can get a good idea of the possible environment of deposition of Facies A and B sands along the lower 1 000 m of the present estuary, from Figure 3.25. This figure shows an oblique view of Squamish estuary and delta taken in August 1947. One can see an extensive unvegetated intertidal zone along the front of the east delta, and around the Mamquam Blind channel. Comparison with recent air photographs reveals that much of the intertidal sands toward the front of the central delta have become vegetated in the last 40 years.

This last point highlights the fact that intertidal sandflats within an interdistributary bay environment will eventually become vegetated. They will develop the same sedimentary sequence as those of estuarine channel fills (which has previously been described as a transition to tidal marsh), and with continued deltaic progradation, a transition to alluvial plain dominated by sands (Facies B, C, E, and F).
Figure 3.25. Oblique view of Squamish estuary and Howe Sound fjord taken in 1947. In front of the central delta and the Mamquam blind channel there is an extensive unvegetated intertidal sandflat.
The Age of Tidal Marsh Deposits

A number of assertions have been made within this chapter regarding the timing of specific sedimentary processes. The first of these is that each rhythmite bed within the tidal marsh Facies C is the product of one year's deposition. The second assertion is that the sandy rhythmites of Facies D are the direct depositional response of the partially constrained river system. The purpose of this section is to test whether these assertions are (or are most likely to be) correct. This can be done by detecting (if possible) an easily identifiable and easily dateable marker bed or horizon within sedimentary sequences.

During the summer of 1992, three bottles (two Orange Crush bottles, and one Kik Cola bottle) were excavated from deposits of Squamish west delta. One of these bottles was found within the bank face, 120 m upstream from the river mouth, and two were found within nearby slump blocks. The bottle within the bank face, which is shown in Figure 3.26, lay 23 cm below the upper limit of Facies C, 157 cm below the marsh surface. This depth is that to the bottom of the bottle (the surface on which it was deposited). The origin of the bottles within slump blocks was determined from the location and correlation of a marker bed (a distinctly coarser, thicker rhythmite) in the bank section, and in the slump block. From this, it was determined that one of the slump block bottles lay 25 cm below the upper limit of Facies C, or 159 cm from the delta surface. The second slump block bottle (Kik Cola) was found 10 m downstream, 26 cm from the upper limit of Facies C.

Within one of the slump blocks, 15 rhythmite beds were counted between the base of the Orange Crush bottle and the upper limit of Facies C. Similarly, 15 beds were counted
Figure 3.26. Bank section where three bottles were found within the silty deposits of Facies C. One of these bottles is seen in this photograph. Also shown is the ledge which forms at the contact between Facies C and D as the overlying coarse sands are removed by erosion. This figure also shows the visual character of Facies D deposits when dug out. Shovel is around 1 m for scale. Upstream is to the right.
from the base of the Orange Crush bottle found within the bank face. From this, it is clear that these two bottles became incorporated into the delta sediments at the same time. The small difference in the depth of these two bottles is probably produced by small-scale local variability of marsh accretion. Unfortunately, due to poor preservation, it was impossible to count the number of rhythmite beds above the third bottle (Kik Cola). Despite this, the bottle is known to have originally lain 160 cm below the delta surface (before slumping occurred). Given the known annual sedimentation rate at this location (an average of 1.6 cm/yr), this bottle is thought to have been incorporated at the same time as the other two bottles, or at most, one year previously.

If my assertions regarding the genesis of tidal marsh rhythmites and sandy rhythmites are correct, then the age of these bottles should not post-date 1957. This date was calculated from the supposed age of the lower contact of Facies D (1972), minus 15 years (one year for each tidal marsh rhythmite above the base of the bottles).

Unfortunately, it is not possible to determine a specific year of production for these bottles, only a range of years when the bottle and label design was employed. Fortunately, one of the two Orange Crush bottles (that found in the bank face) has the words "Design Reg'd 1956" printed on the glass. The second Crush bottle lacks this information, but both the bottle and label design are identical. Lastly, the Kik Cola Company was only in operation for a short duration from the 1950s to early 1960s. From this, it is clear that, although no specific dates can be determined, these three bottles are from the same period of time. These bottles have been independently dated between 1956 and 1962 (G. Jacobs, Vancouver bottle collector, personal communication). This range is in perfect agreement with age of these bottles predicted by sedimentologic analysis. This shows that tidal marsh rhythmite beds are indeed
annual deposits, and that Facies D sediments formed after river training dyke construction in 1972.

**Tidal Deposition in Squamish Estuary**

All seven facies described here can strictly be considered tidal in origin as they were laid down within waters subject to tidal flux. However, in many cases these deposits are only known to be tidal from their location within the estuary. There exists a clear difference between deposits which formed in tidal environments, and deposits which can be classified as tidal in origin from an examination of sedimentologic character and stratigraphic sequence. If these deposits were seen out of context (in rock section for example), then only the deposits of Facies B and C could confidently and accurately be interpreted as tidal in origin.

The tidal characteristics of Facies B sediments are the presence of shells and the sedimentary record of tidal cyclicity produced by rhythmically interlayered sands and silts/clays. As has already been discussed, however, these characteristic attributes of Facies B deposits are not constant throughout all sections, but are variable with distance along the tidal gradient. This point will be discussed in greater detail in the proceeding section.

The tidal characteristics of Facies C are the presence of marine and brackish-water floral species, which form discreet layers to produce the characteristic bedding pattern of the tidal marsh deposits. In addition to this, it is possible that tidal control on sedimentation is reflected in the grain-size distributions of tidal marsh sediments. This possibility will be examined in chapter four.
This study has not found enough evidence to prove the tidal origin of Facies A, D, E, F, and G. However, this does not mean that no such evidence exists, as the scope of this study has necessarily been limited. Further analysis, such as the examination of foraminifera species, may provide evidence of marine or brackish-water conditions (Mitchell, 1976). Furthermore, this study has only been able to examine the uppermost deposits of Facies A. These deposits, however, likely are up to 4 m thick in places (the depth of the estuarine channel revealed by sonar plot at low tide). Although it seems highly unlikely that clay drapes would form within this subtidal environment, the lowest deposits of Facies A may reflect their tidal nature in some way - perhaps in the form of shell beds, or a clay pebble conglomerate lag at the base of the channel sediments (Oomkens and Terwindt, 1960; Bosence, 1973; Barclay and Davies, 1989). These sands may also display changes along the tidal gradient, such as a seaward decrease of bed thickness (Campbell and Oaks, 1973).

Despite this lack of concrete evidence of the tidal nature of a number of individual facies, an examination of stratigraphic sequence throughout the estuarine reach may provide a better indication of the nature of tidal deposition in Squamish estuary. This examination should also provide better understanding of the decreasing influence of tides up estuary. The following section presents the results of detailed investigations of the character of estuarine sequences, and the changing nature of these sequences away from the river mouth.

**The Changing Nature of Deposition along the Tidal Gradient**

Observations regarding the changing nature of tidal influence on sedimentation may be made at a number of scales. Firstly, one can examine sedimentary sequences for broad-
scale changes in facies character, position in sequence, facies associations, bounding surfaces etc. Secondly, one can examine up estuary variability within specific facies which are known to be tidal deposits. Both these scales of investigation have been employed here, the results of which are discussed in the following two sections.

**Observations from Sedimentary Sequence**

From an examination of Figure 3.3, a number of observations may be made regarding the tidal depositional product in Squamish estuary, and how that depositional product changes along the tidal gradient. Firstly, sequences along much of the estuary are characterised by thick units of fine-grained material which often contain considerable amounts of silt and clay. This sediment-size characteristic reflects the dominant mechanism of sediment deposition within interdistributary estuarine reaches (Dalrymple *et al.*, 1991). The sediments of Facies C, E, and G may be considered as overbank fines formed by (or primarily by) the settling of sediment from suspension. The great thicknesses of overbank fine units has been stated as characteristic of tidal environments (Barwis, 1978; Thomas *et al.*, 1987). Such thicknesses of overbank fines will be far less evident in fluvial environments because of differences in depth-velocity relationships within fluvial and tidal zones (Barwis, 1978). The differing thicknesses of the overbank cap has been suggested to be an environmental indicator which differentiates fluvial and tidally influenced riverine environments (Smith, 1987).

Thomas *et al.* (1987) stated that the zone of maximum mud deposition within estuarine reaches of rivers is largely controlled by the location of the turbidity maximum (Postma, 1967). This is not considered to be the case in Squamish estuary, because salt wedge intrusion essentially is restricted to winter months, (Levings, 1980) when suspended-
sediment concentrations are lowest. During summer months, the lack of salt wedge intrusion likely will inhibit the development of locally high suspended-sediment concentrations within Squamish River, though these may develop in areas away from the main channel.

With increasing distance away from the river mouth, sedimentary sequences display an increasing importance of sand, and as such, record a gradual increase in mean grain size with distance upstream. This trend has previously been noted by Howard and Frey (1973). The increasing mean grain size of intertidal and supratidal delta and floodplain sequences is caused by two mechanisms: thinning of Facies C and E, and increasing height of the delta/floodplain surface.

Fine-grained overbank deposits of Facies C and E decrease in thickness upstream. This decrease becomes evident along the ox face (in comparison with sections along the upper west delta), though major thinning does not occur for another 500 m upstream from the ox face. The thinning of these facies (and Facies B) reflects the decreasing amplitude of tidal flux away from Howe Sound, and also the decreasing tidal modification (slowing) of flow. As the thickness of fine, overbank deposits decreases, the relative importance of sand increases.

In addition to this up estuary change in sediment-size, there exists a similar change upsection. This is best seen along the ox face, where sections show a coarsening-upward sequence superimposed on a fining-upward sequence. The fining-upward sequence is the transition from intertidal sands to low, and then high tidal marsh, as seen at sites along the upper west delta. As with the up estuary change in grain-size, the upsection changes are driven by changes in the relative importance of tidal influence. The fining-upward
sequence is highlighted in the transition from Facies A to C, and represents an increasing importance of tidal control on sedimentation. This increasing importance is particularly evident as the environment of deposition changes from subtidal to intertidal. This fining-upward sequence reflects changes in depositional process from bedload (Facies A) to mixed bedload-suspended load (Facies B), to deposition from suspension (Facies C). A similar sequence was noted by Yeo and Risk (1981) in the intertidal zone sediments of the Minas Basin, Bay of Fundy.

In contrast to this sequence, the transition from Facies E to F represents a coarsening-upward sequence cap which reflects a decreasing influence of tides in depositing sediment. As the floodplain surface accretes, tidal inundation becomes less frequent, and water depths decrease. This represents a decreasing input of tidally-deposited sediments to the floodplain. With continued accretion, tidally-influenced waters only inundate the floodplain during periods of riverine flood. During riverine flood, however, higher discharges and flow velocities reduce the degree of tidal modification. As such, even though tidally-influenced waters are able to inundate floodplain surfaces above mean high tide level, riverine depositional processes will dominate. This is clear from the lack of fine deposits toward the top of sections along the ox face, and all locations further upstream.

Discussion has not focused on the deposits of Facies D here, as these represent an anomalous sedimentary succession (discussed earlier in this chapter, and in chapter five).

Sedimentary sequences display a number of other important characteristics of estuarine sedimentation. One such characteristic is that sequences along the lower reaches of
Squamish estuary display simple, predictable sequence successions. If the anomalous deposits of Facies D are ignored, sequences show a succession from basal sands and gravels (Facies A) to intertidal sands (B), tidal marsh (C), to an early floodplain surface (E). Along the lower estuarine reach, sequences show a succession from Facies A, B, C, to Facies D. These simplistic sequences, where facies succeed one another in a predictable fashion, are not observed throughout the estuarine reach of Squamish River, however. Around 2 000 m upstream from the river mouth, sequences record facies successions which differ from those further downstream. Facies transitions are no longer predictable, but become broken by what appear to be random injections of coarser material. These coarser deposits form interbeds within facies, as seen at site ox12 in Figure 3.3. In general, sections in reaches which record increasing fluvial dominance (such as the those along the ox face), show sequence successions which appear fragmented or broken by random deposits of coarse-grained material. The increasing complexity of sequences upstream is thought to reflect the interaction of riverine and tidal energy fluctuations which may be either rhythmic or random (Thomas et al., 1987). The coarse sand interbeds are the product of random fluctuations from large river floods. As tidal influence lessens, both rhythmic (snowmelt-induced daily fluctuation in discharge) and random (storm-induced flood) fluctuations in river flow become so large that the depositional product of tidal rhythmic fluctuations (Facies B and C) either become eroded or fail to form at all.

The erosional force of single riverine flood events is seen from the nature of facies transitions along the upstream most estuarine sections. Where Facies F is present in section, its contacts are always abrupt, and most often erosional. In locations where tides exert a greater degree of control, however, facies transitions are gradational. Such transitions are highlighted by those changes from Facies B to C, and C to E. While
abrupt transitions are found along the lower reaches of Squamish, these are associated with Facies A and D, which may both be considered as largely fluvially derived. The abrupt contact between Facies C and D has already been shown to have been artificially produced by an increase in the relative power of Squamish flow. The deposits of Facies A are essentially subtidal, and internal bedding shows no evidence of tidal flux. It seems from this that the nature of contacts between sedimentary facies and beds may be used as an indication of the degree of riverine dominance over tidal processes, or simply the presence of tidal flux within the water column.

Another possible indicator of the presence of tidal flux is the nature of bedding within facies. Estuarine sequences have already been noted as containing facies which display a wavy or undulatory nature, which is produced as sediments settle from suspension onto an uneven surface, and from consequent differential settling through slumping induced by dewatering and gas-escape. As tides exert less influence on sedimentation (visible both up estuary and upsection), facies display less of this characteristic bedding form.

It should be noted that while I suggest that observations of the nature of facies contacts and bedding may be a useful indicator of the relative degree of tidal influence, these data are by no means proof of tidal environments. The presence of gradational contacts within simple sequences which display undulatory or wavy bedding should only be taken as possible evidence of tidal influence. Such evidence should prompt more detailed stratigraphic and sedimentologic investigation within specific areas of interest.
Observations from Tidal Facies B and C

As has already been noted, deposits present within the estuary display variability which appears to reflect local variability of depositional process and product. Such variability of estuarine depositional product has been noted elsewhere (Land and Hoyt, 1966; Terwindt, 1971; Howard and Frey, 1973; Goldring et al., 1978; Jouanneau and Latouche, 1981; Syvitski and Farrow, 1983). In addition to these local influences, there are more orderly changes which appear to be driven by location along the tidal gradient. From this evidence, conclusions may be drawn regarding the nature of estuarine sedimentation, and the changing character of estuarine depositional product with decreasing tidal influence.

The changing character of deposits can be seen on examination of Facies B and C sediments at various locations throughout the estuary. Deposits of Facies B display distinct changes with increasing distance away from the river mouth - changes which are indicative of decreasing tidal influence. One such change is that noted in the character of bedding. In exposures along the lower 1 500 m of the estuary, Facies B deposits display fine rhythmically alternating sand and silt/clay laminae. As one progresses further up estuary, however, deposits appear to show a lesser degree of rhythmicity and regularity throughout the facies. Initially, this change is noted as an increase in the thickness of sand beds, as is highlighted in Figure 3.27, which shows one 2 cm thick sand bed. This photograph was taken around 1 700 m up estuary, and shows a relative increase in the importance of sand in Facies B deposits. Further evidence of increasing energy conditions is seen on examination of the internal character of sand beds. Previous description and figures of Facies B deposits have shown that sand beds further downstream are parallel or slightly wavy, with occasional ripple foresets. In Figure 3.27,
Figure 3.27. Facies B sands exposed 1 700 m upstream from the river mouth. At this location at the upper end of the wd face (Figure 3.1) these deposits contain distinctly less clay than those sections further downstream, and also contains thicker sand beds. One of these beds contains climbing ripple lamination in-drift (type 1). Ruler is 18 cm for scale. Upstream is to the right.
the indicated sand bed contains climbing ripple lamination in-drift (type 1). This bedding indicates that this 2 cm thick sand bed was deposited during one flow event.

This form of bedding typifies Facies B deposits within the upper reaches of estuary (more than 1 500 m upstream). It highlights the fact that riverine processes have an increasing importance with greater distance upstream. With this increasing riverine dominance (shown by thicker sand beds which indicate higher flow velocities), there is an associated decrease in tidal influence. This is evidenced by decreasing amounts of silt and clay in sediments, and a decreasing thickness of silt/clay beds. This is seen in Figure 3.27, where lighter, finer sediments are less evident and less distinct than at any location further downstream (compare, for example, with Figures 3.6 and 3.7). Toward the upstream limit of Facies B, sediments still display thin, parallel bedding, but contain no clay. Instead, sediments comprise medium and fine sands, with occasional silts. In stark contrast to deposits of Facies B within the lower estuary, the bedding within these deposits show no rhythmicity or cyclic variation from coarser to finer sediments. Figure 3.28 shows the thin deposits of Facies B toward the upstream limit of their recorded presence in floodplain sections. Despite the fine parallel bedding, these deposits clearly are very different from those along lower reaches of the estuary.

Deposits of Facies C display similar changes of sediment-size and bedding characteristics with increasing distance away from the river mouth. As with the example above, changes in these characteristics are very gradual (undetectable over distances less than several hundred metres), and are highly disturbed by local variability. As the deposits of Facies C have already been shown to develop within tidal marsh environments, their tidal origin is clear. However, the sediments contained within the rhythmite beds contain evidence of the degree of tidal or riverine influence within the
Figure 3.28. Site nyl, showing the upper 1.5 m of section which is dominated by the sands of Facies F. Toward the base of this section are thin sandy deposits of Facies B. Note the fine parallel-bedding within these deposits which are much coarser than at any other site further downstream. Upstream is to the left.
tidal marsh. As a general rule, the sediments of Facies C increase in size with increasing distance upstream. Within west delta sections, rhythmite beds contain fine sandy silts with clay. In places, clay content exceeds 33% of bed composition (by weight), and beds rarely contain less than 10% clay. Along the ox face, rhythmite beds contain fewer clays, and are considerably more silty. With increasing distance upstream (toward the upstream limit of Facies C), beds are dominated by fine, and even medium sands. Campbell and Oaks (1973) found a similar increase in sediment-size within tidal flat deposits (which included tidal flat proper (Facies B) and tidal marsh (Facies C)).

In addition to these changes in sediment-size composition, tidal marsh rhythmite beds also display variable thicknesses throughout the estuary. Initially, rhythmite beds decrease in thickness with increasing distance upstream. Beds exposed along the ox face are distinctly thinner than those exposed along west delta sections. This decreasing thickness reflects the decreasing tidal amplitude (induced by river slope), and more importantly, the decreasing tidal modification of riverine flow. As tidal flux is less able to slow Squamish flow, particularly close to and during high tide, fewer fines will be able to be deposited from suspension. This is evident from the decreased thicknesses of annual deposits, and the decreasing clay content of beds.

Rhythmite beds do not show a simple thinning trend with increasing distance up estuary, however. Toward the upstream most limit of Facies C, beds increase in thickness again. This increase is associated with a notable increase in sediment-size, as discussed above. The thickness and texture of these beds are indicative of riverine dominance. Although deposits formed within a marsh system which was regularly inundated by tides, sedimentation rates were highly influenced or dominated by riverine input. Flows adjacent to, and over this marsh area were clearly of higher velocity (from the amount of
sand) than flows which inundated the west and central deltas. Apart from these changes in the average thickness of rhythmite beds, these beds display the characteristic thinning upsection at all locations throughout the estuary.

**Upstream Limit of Detectable Tidal Sedimentary Control**

One of the main aims of this research is to determine the extent of tidal control or influence on sedimentation within Squamish estuary. The extent of tidal control can be defined as both the degree of sedimentologic and stratigraphic influence, and the upstream limit of detectable tidal deposition. The term detectable tidal deposition is used to describe those deposits which contain sedimentary evidence of their tidal nature at the scale of investigation employed in this study. The aim of this section is to report on the upstream limit of detectable tidal deposition.

The landward limit of detectable tidal deposits marks an important boundary termed the bayline, as defined by Posamentier and Vail (1988). Upstream from this limit, tides may influence flow (particularly during low flow periods), but will not produce distinct tidal bedding. The bayline is a dynamic feature, and may shift either away from or toward the depositional basin in response to sea-level changes (Posamentier *et al.*, 1988).

Within Squamish estuary, the bayline is marked by the upstream limit of Facies B, and is therefore around 3 300 m upstream from the river mouth (channel centreline distance). This limit is considerably less than the length of Squamish estuary (at least 5 500 m), which is defined as that portion of the river influenced by tidal flux. Similar observations regarding bayline location were made by Allen (1991), who examined the fluvial-tidal transition along the Garonne River and Gironde Estuary in France. Allen
concluded that the bayline marked the landward extent of estuarine facies, and that all locations upstream from this were fluvial channel facies. Dalrymple et al. (1992) similarly defined estuaries as that portion of the system between the landward limit of tidal facies and the seaward limit of coastal facies. I feel, however, that Allen's (1991) interpretation and the definition of Dalrymple et al. (1992) are misleading, as they assume that estuarine reaches of the rivers are characterised by some form of detectable tidal deposits, and that estuaries end at the bayline. I think that it is more accurate to define the estuary from hydraulic, not sedimentologic considerations. These considerations have been employed, correctly I believe, in the definition and interpretation of estuarine sequences by Allen and Posamentier (1993). If the estuary is classified as that part of the river subject to tidal flux, then estuarine sequences will display both fluvial and tidal deposits. The location of the bayline within this estuarine reach may then be used as an indication of the relative fluvial or tidal dominance within a particular system. Within Squamish estuary, for example, the importance and dominance of riverine over tidal depositional processes is indicated by the large distance (at least 2 200 m) between the upper limit of estuary and the bayline.

The importance of this distinction is highlighted in Figure 3.29, which shows a schematic representation of estuaries based on a number of different definitions. One of the most commonly-used definitions of an estuary is that of Pritchard (1967), who defines an estuary in terms of salinity as that portion of the system where salinities range from 0.1‰ to 30-35‰. Dalrymple et al. (1992) argue convincingly that while this definition is of some value, it is of limited use to the sedimentologist because estuarine sedimentation primarily is determined by physical rather than chemical processes. The estuarine limit associated with the definition of Pritchard is shown in Figure 3.29, along
Figure 3.29. Schematic diagram of estuarine limits as determined from a number of different definitions of the term estuary. Modified after Dalrymple et al. (1992).
with that associated with an alternative definition proposed by Dalrymple et al. (1992), employed by Boyd et al. (1992), and that used in this study.

There are a number of important points to consider in regard to the location of the bayline in Squamish estuary, or any other estuary. The first of these is the obvious point that the upstream limit of tidal deposits identified in this study may not give a true indication of the bayline location, as observation is limited by the location of sections within valley-fill sediments. Also, I do not feel that it is strictly correct to state that tides exert sedimentary control for 3300 m along Squamish estuary, as this assumes that the tidal deposits were laid down by an estuarine channel which had the same location and morphology as that today. If the palaeochannel were either straighter, or more meandering, the true upstream limit of the bayline would be either less or more than the 3300 m reported above. Given the nature of flow of Squamish River, and the narrow valley width, this is not a major concern in this study area. It may, however, be of much greater importance in areas where large, slow flowing rivers are able to migrate across extensive floodplains. Because of this, it is perhaps best to report the location of the bayline as a straight line distance from the delta head or shoreline. Under this system, the bayline in Squamish is located 3100 m upstream from the river mouth.

In addition to the above, it is also important to consider the affect that delta progradation will have on the relative location of the bayline. Posamentier et al., (1988) discussed the dynamic nature of the bayline with regard to sea-level changes. In addition to this, it seems highly likely that the bayline location will be influenced by changes in depositional environment of greater periodicity and speed of onset. For example, one would expect to see a seaward translation of the bayline accompanying delta progradation. As the delta-front progrades, the upper delta plain reaches become less
influenced by tidal flux, and more influenced by fluvial depositional and erosional processes. These will tend to re-work delta topset tidal deposits, and lay down alluvial plain deposits with no detectable tidal influence. In this way, the bayline will gradually shift down valley at a rate which should (as a long-term average) keep pace with the rate of delta progradation. Over shorter periods of time (perhaps decades) there will exist a lag between shifts in the location of upstream most tidal deposition (driven by progradation), and shifts in the location of the bayline, as the erosion of tidal deposits will not be immediate.

This may have important consequences for sedimentologists attempting to delineate estuarine zones based on sedimentologic and stratigraphic data, or those attempting to determine the origin of deposits from an examination of flow processes at that particular location. The exact lag time will be specific to a particular system, but will be determined by the energy environment of the fluvial system, the lateral stability of the channel(s), valley width, and possibly the sediment-size characteristics (erosional resistance) of tidal deposits. From this, lag times in Squamish estuary are thought to be relatively small, though no exact figure is known.

Lastly, it is important to consider the location of the bayline in relation to the seasonality of flow in this environment. It seems likely that the upstream limit of tidal deposition will be highly dynamic, and shift upstream or downstream over a number of time scales. These shifts may occur as discharge fluctuates over the course of one day (particularly during summer months when temperature changes induce variable snowmelt), or as discharge fluctuates over seasons. It is unclear, however, how sensitive the depositional system is to these short-term fluctuations of the location of tidal deposition. For instance, the zone of tidal deposition should extend further upstream during winter
months than during summer months when discharges are much higher. It is impossible
to determine, however, whether the bayline will record the upstream location of winter
tidal deposition, or whether summer flows will erode these deposits before they can
become preserved in sequence. Allen (1991) reported that the location of the bayline
coincided with the upstream limit of reversing flows during periods of high river
discharge.

Zonation of Squamish Estuarine Deposits

In the past, a number of authors have attempted to delineate estuarine subenvironments
based on morphologic regions, sedimentary structure, or biologic character. The aim of
this section is to discuss some of these works and to consider the zonation exhibited
within Squamish estuary.

As has already been stated in chapter one, the first attempts to fully investigate the
transitional reaches of estuarine rivers were published in Senckenbergiana Maritima in
1975. The work of Dörges and Howard (1975), Howard et al. (1975), and Greer (1975)
determined that a number of regions exist within the estuary under study, but that these
regions need not necessarily be the same for all variables under consideration. For
example, Howard et al. (1975) divided the estuary into three regions: upper, middle, and
lower. This division was based on the distribution of chemical and hydrographic features
such as salinity, pH, water temperature, and turbidity. Animal distribution characteristics
were distinct within these three regions. In an investigation of sediment texture and
physical and biogenic sedimentary structure, Dörges and Howard (1975) determined that
four estuarine facies were distinguishable. These facies were termed inner, upper-
middle, lower-middle, and outer estuarine facies, which did not correspond with the
regions of distinct animal communities, but did correspond with regions identified from
textural analysis of point bar deposits. Transitions between regions or facies were termed
highly gradational, and no attempt was made to produce any form of model for the
subenvironments identified.

Jouanneau and Latouche (1981) divided the Gironde Estuary into three reaches based on
morpho-sedimentological assemblages, and were the first to use the term "riverine
estuary". Dalrymple et al. (1990) developed a similar tripartate facies zonation of sandy
deposits along reaches of the Cobequid Bay-Salmon River estuary, Bay of Fundy. Allen
(1991) followed these works by proposing a similar morphological estuarine division,
but then made an important further contribution by describing and graphically
presenting the distinct facies types within these morphological zones. This longitudinal
facies sequence, which is similar in form to that proposed meandering point bar deposits
by Smith (1987), is shown in Figure 3.30. Further significance of this longitudinal facies
pattern is given by Allen and Posamentier (1993).

Since publication of these works, a limited number of papers have attempted similar
investigation of estuarine zonation. Nichols et al. (1991) proposed an estuarine facies
model similar to that of Allen (1991), shown in Figure 3.31. This study differs from
those previously undertaken as it presents a schematic representation of facies changes
within a microtidal environment, whereas most environments under study are meso-
or macrotidal (Hayes, 1976). Dalrymple et al. (1992) propose a series of conceptual
estuarine facies models for wave- and tide-dominated environments. In a series of
conceptual schematic diagrams the authors relate sedimentary facies to estuarine
morphological components and to relative energy conditions along the length of the
estuary. As has already been noted, however, this work fails to consider the example of
Figure 3.30. Estuarine division proposed by Allen (1991).

Figure 3.31. Estuarine facies model proposed by Nichols et al. (1991).
fluvial-dominated estuarine systems.

One common theme of these proposed schemes is that they are based (at least initially) on morphologic divisions of the estuary under investigation. Once these morphologic reaches or zones have been identified, authors then look within these zones to identify characteristic sediment texture and structures. This approach is open to the criticism that the prespecification of zones imposes some bias on sedimentologic investigation and interpretation. The approach employed in this study, however, has necessarily been different. Sedimentary sequences within Squamish estuary display gradual changes down estuary, yet these occur without an associated change in channel morphology. The estuarine reach of Squamish estuary is always single channel meandering, and displays only a minimal estuarine funnel. It clearly is more appropriate to consider changes in sedimentary facies in relation to changing energy conditions associated with (in this case) changes in the relative influence of river and tides, than to consider changes in relation to estuarine morphology.

The following paragraphs outline attempts to delineate estuarine zones within Squamish estuary based on sedimentologic and stratigraphic evidence. It is clear from the discussion above that there exist a number of proposed terminologies for estuarine division, which are not necessarily applicable within this system. For example, the term riverine estuary (Jouanneau and Latouche, 1981) cannot be used here as the entire estuarine reach of Squamish River may be considered riverine estuary because of the dominance of the fluvial input to this system. As I have no desire to add to the growing list of terms or add credence to any proposed scheme, Squamish estuary has been divided into four zones labeled I to IV. These zones, which are shown in Figure 3.32,
reflect the decreasing riverine dominance with increasing proximity to Howe Sound fjord.

Zone I extends from the upstream limit of tidal flux to the upstream limit of tidal deposits (the bayline). The upper limit of this zone denotes the extent of Squamish estuary, although as has already been noted, this limit is not static but shifts seasonally and daily. Sedimentary deposits and sequences within this zone display no detectable tidal influence and as such are dominated by coarse sands and gravels separated by numerous erosional surfaces. Sedimentation within this zone appears entirely driven by fluvial processes. While tides clearly influence flow to some degree, this is not enough to induce deposition of sediments which display their tidal nature.

Zone II extends from the bayline to a point around 1,500 m upstream from the river mouth. The downstream boundary of zone II corresponds with that location where the sandy deposits of Facies F are first observed in section. This location is easily detectable in section as the sands of Facies F form a coarse sequence cap which differs considerably from those deposits below. Sedimentary sequences display considerable variability throughout the length of zone II. Toward the downstream end, sections display the simplistic sequence comprised of Facies A, B, C, E, F and G. The fine-grained deposits of Facies B and particularly Facies C are thick at this point. Toward the upstream limit of zone II sections are more complex, as Facies B and C thin and eventually pinch-out, and coarse sand interbeds disturb the simple sequence transitions noted further downstream. Contacts between facies commonly are gradual, though abrupt and erosional contacts increase in importance toward the upstream limit of zone II. This zone may be considered the transitional zone where both fluvial and tidal processes influence sediment deposition and sequence appearance. Tides exert enough
Figure 3.32. Zonation of Squamish estuarine facies. Zones I to III reflect an increasing importance of tidal depositional processes downestuary, as revealed from facies analysis. Zone IV is a reach of increased fluvial influence caused by the construction of a river training dyke. Distance from Howe Sound fjord to Mamquam River is 5 500 m.
influence to allow the settling of fine-grained sediments from suspension during high tide periods, yet do not modify river flow enough to prevent large flood events depositing sands on delta and floodplain surfaces.

Zone III extends from the downstream limit of deposits of Facies F to the point at which Facies D deposits are detectable in section. This downstream boundary is not very clearly defined, as Facies D sediments fine up estuary and become difficult to detect, and may easily be confused with those deposits of Facies E. Sequences within this zone are comprised almost entirely of the fine-grained deposits of Facies C. Only Facies A, B, C, and E are present in sections, which appear similar throughout the length of this zone. Facies transitions most commonly are gradual, and never erosional. Deposits of Facies B display a high degree of tidal influence and record some form of tidal cyclicity, most probably produced by diurnal tidal flux. This zone represents that region where tidal influence is greatest and where river flow is considerably modified and slowed such that the deposition of large sand interbeds or erosion of existing deposits does not occur.

Zone IV extends from the limit of Facies D sands to the seaward limit of intertidal deposits. This zone perhaps is unique to this environment as it records an increasing fluvial influence (which is in contrast to the transition from zones I to II to III which record decreasing fluvial influence) with increasing proximity to the river mouth. This increase in not a natural phenomena, but instead has been induced by dyke construction and the associated reductions in floodwater storage capacity. The decreased channel and floodplain widths have produced anomalous sedimentary sequences characterised by transitions from Facies A to B, to C, and finally to Facies D. If Squamish River training dyke had not been constructed, sedimentary sequences likely would display the same characteristics as those immediately upstream. In this case, zone III would extend
throughout the lower reaches of the estuary to the seaward limit of intertidal deposits, resulting in a tripartate estuarine division.

The limits of all zones identified here are non-static features in the long-term, and they may all be expected to shift seaward as Squamish delta progrades.

**Assumptions of Methodology Employed**

Investigation of stratigraphic sequence from available section evidence (in this case, channel banks) is possibly the most commonly employed methodology of sedimentologists. This technique, however, is not without criticism, as it makes an important assumption which must be considered when interpreting section data. This form of sampling technique assumes that the present section setting can be considered the same as the original depositional setting of those sediments. This assumption is of critical importance in this study, as it could (if proved wrong) invalidate my conclusions regarding the upstream limit of tidal sedimentary control (the bayline).

As an example, if sediments have a high residence time then deposits within the fluvial environment may in fact be 1,000 years old and deltaic in origin, thus bearing no relation to the depositional environment at the present location. When applied directly to Squamish estuarine deposits, this could mean that the upstream-most tidal deposits were deposited at a time when the delta front was 1,000 m further up valley than at present. This would mean that tidal sedimentary control only extends 2,100 m upstream, and not the 3,100 m that I have stated.
While this may invalidate sedimentologic interpretation in certain environments, this is not considered important here, for a number of reasons. Firstly, it is felt that the residence time of sediments within the confined Squamish River valley is too small to record considerable delta progradation. Sediments can reasonably be expected to be representative of the environment within which they are found. This is particularly true of the degree of tidal influence (longitudinal component), though it is more difficult to assess the lateral component - whether sediments were deposited in proximal or distal channel locations. This latter point, however, is of far less importance in this study than that related to tidal influence.

In addition to this question of low residence time, there are a number of pieces of sedimentologic evidence which indicate that sediments are representative of their present locations. The first is the fact that the tidal deposits of Facies B and C thin upstream, and eventually pinch-out. This gradual decrease in thickness is a good indication that the exposed deposits of these facies are contemporary with each other. Facies C deposits were formed within an extensive tidal marsh environment, over the same period of time. Secondly, the lower 1000 m of Facies B sediments contain shell debris, but these are not noted further upstream. These shells are thought to accumulate within intertidal sandflats along the seaward limit of the delta. The fact that these shells are not observed in upstream deposits of Facies B would seem to rule out the possibility that Facies B deposits record progradation of deltaic intertidal sands.

The most important sedimentologic evidence is provided from the recent sediments along the ny face. Within these sediments there is conclusive proof that the tidal bedding furthest upstream is contemporary with the environment of deposition seen at that location today. Examination of Figure 3.1 reveals that the ny face sections are located
toward the upper limit of the abandoned central channel. As such, the deposits seen along this face represent IHS or IS deposits (Thomas et al., 1987) associated with point bar growth, and sediments of the sandy plug which filled this upper meander cutoff. These sediments can only be 60 years old at most. The fact that tidal deposits of Facies B can be seen in this face (in Figure 3.28) indicates that tidal sedimentary control does indeed exist over 3 000 m upstream. The fact that these tidal deposits are so thin at this location (and cannot be traced more than 40 m upstream) indicates that this location is very close to the upstream-limit of tidal sedimentary control.
A number of assertions have been made in chapter three regarding the tidal character of facies within Squamish estuary, particularly the tidal marsh deposits of Facies C. It is important, however, to test whether overbank fines within purely fluvial reaches display some or all of the noted characteristics of tidal deposits. The aim of this chapter is to make comparison between fine-grained deposits in fluvial and estuarine environments to enable clearer definition of the character of tidal deposition. Comparison is based on a number of observations and analyses from bed-scale sedimentologic analysis to element-scale analysis of stratigraphic sequences. Although fine-grained sedimentary deposits may be of significant value as indicators of environments of deposition, Miall (1987) notes that they have received little attention compared with their sandy counterparts.

The following sections describe the possible environmental significance of overbank deposits, and present a description and interpretation of specific deposits exposed along fluvial reaches of Squamish River. The characteristics of these fine deposits will then be compared with those of tidal marsh Facies C. As the specific aim of this study is to determine whether estuarine fine-grained deposits can be accurately and confidently identified as tidal in origin, I examined those fluvial deposits which most closely resembled the deposits of Facies C. This is because these visually similar deposits are most likely to be misinterpreted as tidal in origin. The aim here is not to find those fluvial deposits which are genetically analogous to their estuarine counterparts.
Description of Fine-Grained Units Exposed Along Fluvial Reaches

The fine deposits described here, and shown in section in Figure 4.1, are observed in river bank sections over 13 000 m upstream from Howe Sound fjord. Fines are seen in section at a number of isolated locations indicated on Figure 4.2. Between this location and a point around 4 000 m upstream from Howe Sound fjord, no fine-grained floodplain deposits are observed in channel bank sections of Squamish River. Where fine deposits are present, they are recorded at the base of stratigraphic sequences which are often over 600 cm in height, even during the high flow months of July and August. One such sequence is shown in Figure 4.1, which shows a section diagram for the site indicated in this figure. This particular location was selected for description as it displays the typical Facies assemblage of these upstream sites.

The lowest deposits in section are the finest, ranging in size from a fine sand to a silty fine sand. These deposits are at least 200 cm thick (and at least 370 cm thick in places), though their lower boundary was never observed or reached during augering. The upper boundary is marked by an abrupt change in sediment-size to a medium sand. This contact occasionally appears erosional, as the height of the fine deposits is variable in section over short distances, and bedding within these upper deposits seems discontinuous. Deposits are finely parallel-bedded, and generally consist of alternating thin layers of inorganic material and layers dominated by organics. This alternation produces rhythmite bedding throughout much of the facies. Though beds are parallel, they are not necessarily horizontal. At one location beds appear to dip down out of the face (perpendicular to contemporary flow), and the facies is higher in section 10 m upstream than downstream.
Figure 4.1. Channel bank section exposed along the purely fluvial reach of Squamish River, and section diagram for this location. The site illustrated here is representative of all locations along this channel reach. For the exact location of this site see Figure 4.2. The fines are seen here at the base of a section which is 5 m high at the time the photograph was taken.
Figure 4.2. Location map of fine-grained deposits seen in sections along the fluvial reach of Squamish River. Both general and specific site locations are shown here. The location of sections are shown by a solid black line.
The dominantly inorganic layers are of highly variable thickness both with depth in section and between sections. At one location, an inorganic layer exceeded 9 cm in thickness, while at other locations sediments were only 0.1 cm thick in places. Beds show no pattern of either thinning or thickening upsection; instead, they display variable thickness and adjacent beds may have distinctly contrasting thicknesses. Similarly, the organic-rich layers display variability of both thickness and content. Some of these noted characteristics are seen in Figure 4.3. The organics contained within these layers range from twigs, whole deciduous leaves, pine needles, coffee-ground organics (woodchips, bark etc), and grasses and reeds. Not all of these organics are present in each layer, and individual layers are not of constant composition throughout. At any given location, organic layers may be comprised entirely of twigs, but when traced laterally over distances as little as 200 or 300 cm, these same layers may comprise leaf litter and numerous pine needles. At one site, partial erosion has exposed a number of sedimentary beds and the upper surfaces of the organic-rich material between beds. The organics at this location dominantly comprised flattened grasses and reeds which were very similar in appearance to the Sedge stems observed within Facies C in the tidal marsh.

Unlike the rhythmite bedding of Facies C (produced by the annual death and collapse of vegetation), the bedding form within these fluvial fines is not always a rhythmic alteration of 2 distinct beds. In many places, alternating organic and inorganic layers produce rhythmite bedding, but this is rarely present throughout the entire facies. Instead, there are isolated sections which fail to display any rhythmicity, as organic layers are not present. The sediments in these locations are still parallel-beded, but are not separated by an organic film. Organic layers clearly are laterally discontinuous, and can often be seen pinching-out in section.
Figure 4.3. Fine-grained deposits exposed at the site indicated in Figure 4.1 within the fluvial reach of Squamish River. This photograph shows many of the features characteristic of these upstream fines. In some places, parallel beds are quite distinct, while in others bedding is not evident. At this particular location, the importance of twigs within organic layers is evident. Once again, however, distinct organic layers are not always visible here.
These fine-grained deposits are overlain by up to 400 cm of medium to coarse sand. These sands may be divided into a number of distinct facies which are of highly irregular thickness across the section, and as such are found at variable depths within section. All contacts between Facies are abrupt and occasionally erosional. Finer deposits (medium sand) contain a number of bedding forms ranging from finely laminated horizontal, parallel bedding, climbing ripple lamination (both in-phase and in-drift), flaser bedding with connected lenses, and convolute bedding. Coarse sand deposits show greater internal variability of bedding form. In places, the gradual transition from climbing ripple lamination in-phase to in-drift (type I to type II) is fully preserved within these coarse sands. Elsewhere, Facies Contain numerous reactivation and erosional surfaces. While these often produce abrupt changes in bedding form, no associated rapid changes in sediment-size are recorded. Within one facies, numerous trough-shaped erosional surfaces produce the characteristic bedding of trough cross-stratification (McKee and Weir, 1953). Occasionally, fine laminae become overturned and have a highly irregular form characteristic of convolute bedding. No coarse sand Facies Display detectable fining or coarsening-upward sequences. Toward the top of the section shown in Figure 4.1 there are three organic layers 1 to 2 cm thick. These layers are formed primarily from leaf litter and twigs, and contain occasional roots.

**Interpretation and Reconstruction of the Environment of Deposition**

The fine deposits at the base of the section described above are considered to be backwater deposits which developed within an abandoned channel section. These have since become partly eroded and exposed by the contemporary Squamish River. Channel abandonment probably occurred through meander neck cutoff forming an oxbow, as will
be explained later. Two more-recent oxbow environments can be seen along this stretch of river in Figure 4.2.

During periods of high river stage, this environment would have become ponded, yet removed from the high velocity flows within the main channel. Toward the upstream end of the abandoned meander, coarse sediments entered the channel from the adjacent main channel flow and were deposited, forming a sand plug. This sand plug partially closed the abandoned channel, forming a backwater environment within which fine sediments were deposited from suspension as a series of finely laminated parallel beds. This environment was not constantly ponded, however, as there is evidence of the subaerial exposure of sedimentary beds. As floodwaters receded and stage fell within the cutoff reach, some floating organic debris (transported into the system via the main river) was deposited or laid on the upper surface of the freshly deposited sediment. Contribution of organics to the organic-rich layers by this mechanism, however, is less important than the contribution from the surrounding forested floodplain. The organic layers essentially represent a period of non-deposition, when water levels in the oxbow were low enough for sediment surfaces to be exposed. During these periods, leaf litter and other falling debris (twigs, branches etc) collected at the base of the cutoff. These organics are buried by the sediments introduced by the following flood.

The superposition of medium and coarse sand on these fine-grained deposits results in a coarsening-upward sequence which may not ordinarily be expected in such an environment. This coarsening may have been formed by the reactivation of flow within this abandoned channel section. This may have been produced by an increasing proximity of Squamish River to this abandoned reach as the main channel migrated. An alternative possibility is that the coarsening-upward sequence was produced by the
migration of the sand plug from the upper end of the cutoff. As these sands gradually migrated through the cutoff, they would bury the finer suspended-sediment deposits which formed behind the original sand plug.

The uppermost 260 cm (at least) of deposits shown in Figure 4.1 are coarse overbank flood deposits laid down by a limited number of large flood events. These top 260 cm are thought to have been laid down by a maximum of seven flood events. Within these flood deposits there are a number of convolute beds (Kuenen, 1953) in the medium and coarse sands. While there have been numerous interpretations of this form of deformation structure (discussed in Reineck and Singh, 1980), these are thought to have been produced by rapid sedimentation and consequent loading of near-saturated sands. The organic layers noted between the thick sand facies toward the top of the sequence shown in Figure 4.1 represent leaf litter (primarily) which collected on an older floodplain surface.

This interpretation of channel fill is based on various forms of stratigraphic evidence. Firstly, their location at the base of sections and the superposition of 400 cm of coarse material essentially rules out the possibility that these deposits are either distal overbank fines or backswamp deposits. Secondly, there is limited evidence to suggest that the lateral boundary of these fine deposits (exposed at one location) is abrupt and near-vertical. As this boundary did not appear erosional, it most likely represents the outer bank of the palaeochannel. In addition to this, the limited extent of these thick overbank fines in channel bank sections is characteristic of small yet deep depositional sinks which are restricted in areal extent by the former channel boundaries. The angled nature of some of these beds reflects the uneven form of the underlying deposits. As sediment settles from suspension within flooded abandoned reaches, beds tend to assume the form
of the surface on which they become deposited. Where sediments settle on a point-bar surface, sediments deposited from suspension will parallel this form, producing a series of angled beds.

In addition to the reasoning above, the interpretation of channel fill is also based on the rejection of other possible environments of deposition, as discussed below. It is not implausible that the fine deposits seen at this upstream location formed within a lacustrine environment, or that relict estuarine deposits (equivalent to those of Facies C and E in the present estuary) may remain at isolated locations up valley from the present delta front. There are, however, a number of reasons why I do not feel that these explanations are appropriate.

Fine-grained deposits could have been deposited in a lacustrine environment if Squamish flow had become dammed. The most likely cause of any such possible damming in this region is if avalanche and associated debris-flow deposits entered Squamish valley from the Cheakamus River. The Cheakamus River enters Squamish River 5 700 m downstream from the location of the fine deposits described here, and is seen in Figure 4.2. Such catastrophic damming has previously been recorded in this region. Evans and Brooks (1991) and Brooks (1992) describe a number of backwater deposits produced by 7 or 8 impoundments of Squamish River during the Holocene. Despite the fact that damming could have occurred at this location, the fine deposits observed and recorded in this study do not appear to be lacustrine. Firstly, deposits never display any varves which would be expected in such an environment. In previous recorded dammings along this river system, lacustrine deposits up to 600 cm in thickness contain rhythmite beds 1 to 3 cm thick, all of which fine upwards (Brooks and Hickin, 1991). The rhythmite beds described here never displayed fining-upward trends.
Secondly, these fine deposits are not considered to have enough longitudinal extent to be part of a lacustrine system which extended around 6,000 m downstream to Cheakamus River. These fines were not observed in bank sections at any point between the confluence and the site described in this chapter. Lastly, the organic content of these deposits conflict with a lacustrine setting. At a number of locations, reed stems (similar in appearance to those in west delta sediments) grow vertically through sedimentary beds; a clear indication of subaerial exposure.

It is considered highly unlikely that the fine deposits described here are tidal in origin (recording a former location of Squamish delta), as the preservation potential of estuarine fines (discussed more fully in chapter five) is far too low to permit the preservation of deltaic topset deposits this far upstream. If the short-term rate of delta progradation (3.86 m/yr) identified by Hickin (1989) is representative of the rate of progradation throughout the Holocene, then these deposits must exceed 3,000 years in age. This rate of Holocene progradation has been confirmed by radiocarbon dating performed by Brooks (1992). Even within the lower reaches of Squamish River, the tidal Facies Described in chapter three were produced by contemporary estuarine flow, and do not represent relict deposits which have become essentially abandoned as the delta prograded. In addition to this consideration, the purely fluvial rhythmite deposits can reasonably expected not to be tidal in origin, as they are quite dissimilar from any known tidal deposits exposed in Squamish estuary and delta at present.

**Comparative Analysis of Fluvial and Tidal Rhythmite Overbank Fines**

From the above description of fluvial rhythmite fines, it is clear that there exist a number of characteristic differences between these and overbank deposits in Squamish estuary.
The following sections examine in greater detail the characteristic similarities and differences between the deposits in these two environments. Discussion is divided into sections which deal specifically with facies geometry, organic content, sediment-size characteristics, and bedding.

**Facies Geometry**

Estuarine overbank Facies are characterised in part by their areal extent and geometry. The deposits of tidal marsh Facies C are present across the entire subaerial delta complex (over 2 000 m) and are detectable in river bank sections up to around 3 000 m upstream from the river mouth. Such extensive deposits reflect the deposition of material within a large shallowing open depositional system. In longitudinal section, Facies C displays a lens-like geometry produced by the younger age of the seaward limit of tidal marsh, and the decreasing amplitude of tidal flux with increasing distance up estuary.

In contrast, fluvial overbank fine deposits are limited in longitudinal extent to three isolated exposures, as shown in Figure 4.2. The lateral extent of these deposits cannot be determined in this environment. These deposits seem to have formed in an environment which was partially closed, as this Facies appears to terminate abruptly at a steep contact. This contact marks the location of the outer bank of the palaeochannel.

**Organic Content**

Both fluvial and tidal rhythmite beds contain thin layers of organic material. In places, the rhythmic alternation of parallel organic and inorganic layers within fluvial fines
closely resembles the tidal rhythmites in river bank sections. For example, compare the fluvial rhythmite bedding in Figure 4.4 with the tidal rhythmites seen in Figure 4.12. Despite this occasional visual similarity, however, the organic layers within these two environments differ greatly in their composition. Within Squamish estuary these layers almost exclusively comprise sedges (dominantly Carex lyngbyei, Ian Hutchinson, personal communication) which become flattened into interwoven thin mats. With the exception of organic layers formed during the colonisation phase of marsh development (Redfield, 1972), marsh vegetation densities remain relatively constant. This results in organic layers which are of very similar thickness throughout the facies. On one occasion a deciduous leaf was found within deposits of Facies C, but small coffee-ground organics, twigs and pine needles were never found.

Within fluvial overbank fine deposits, organic content is far more diverse; flattened grass and reed stems, twigs, deciduous leaves, and pine needles were all present. Organic layers are not of constant composition throughout the facies, as one or all of the organics mentioned above may be found within any given organic layer. Also, individual organic layers may not be of constant composition across section, as the layer may be composed entirely of pine needles at one location, yet when traced laterally across the face by only 200 or 300 cm, it may be composed of twigs and leaves.

In addition to this, upstream rhythmite beds differ in appearance because the organic layers protrude from the face to a lesser extent than those within Squamish estuary. This is partly related to the different organic content, which does not form such an interwoven mat of material which resists erosion. It is also related to the fact that upstream deposits are not continually submerged and exposed by tidal stage fluctuations.
Figure 4.4. At this location (shown in Figure 4.2), these fluvial rhythmite beds quite closely resemble those of tidal marsh Facies C. Beds are thin and parallel-bedded, and organic layers protrude slightly from the face. Toward the top of this photograph, one can see the abrupt contact of this facies to the overlying medium sands. The fines exposed here are over 1 m in thickness.
The upstream variability of organic content accords with observations by Barwis (1978), who notes that the organic caps of fluvial point bars contain a wider variety of plant material than similar deposits overlying tidal point bars. Despite this observation, however, Barwis (1978) did not discuss the specific organic forms within fluvial overbank deposits.

The difference in organic content between these two environments is driven primarily by the mode of organic accumulation. Within interdistributary bay environments the organic material preserved in bedding sequences accumulated in-situ. Within abandoned channel sections the organic material is primarily transported into the environment by floodwaters, and delivered directly from the surrounding vegetated floodplain (Fielding, 1985). These different modes of accumulation leave organics which are termed autochthonous and allochthonous respectively (McCabe, 1984). There are three main reasons why allochthonous deposits are of such importance in the fluvial fine deposits but not in estuarine sequences. One is the obvious point that the presence of leaves, twigs and pine needles in the channel cutoff indicates their presence nearby. In addition to this, it reflects the fact that the cutoff was a partially closed system which, unlike the intertidal marsh, was not regularly flooded. Thirdly, the presence of leaves and needles in sediments indicates high rates of sedimentation which rapidly bury the organics. Within Squamish estuary, sedimentation is more gradual and far less rapid during a depositional event. If drifted organics fail to be covered by sediment, they likely will become transported out of the marsh system during the proceeding high tide period.

Despite these observations at this location, however, it should be noted that these forms of organic debris may be found within estuarine sands and muds. For example, Syvitski and Farrow (1983) found pine needles within intertidal mouth bar sands of Bute Inlet,
British Columbia. The presence of these forms of organics, however, likely will be minimal in relation to the amount of autochthonous material present, and in relation to the amounts present within fluvial channel fills.

**Sediment-Size Characteristics**

Detailed sediment-size analysis of overbank fine deposits was performed on samples taken from two sites: one from the fluvial environment and one from Squamish west delta. From these sets of samples, individual rhythmite beds were analysed by sieve and SediGraph particle size analyser in the manner described in chapter two. These data provide a percentage break-down of the sand-silt-clay fractions present within each rhythmite. The percent sand, silt, and clay data are discussed below for both tidal and fluvial rhythmites.

**Tidal Marsh Rhythmites**

Sediment samples of tidal marsh rhythmite deposits were collected at site irs1, located 1100 m upstream from the river mouth. In total, 146 samples were analysed, which represent the whole thickness of Facies C at this site (188 cm). The beds at the top of this Facies Could not be individually sampled because of their thinness, so the upper 98 cm of deposits were sampled at 1 and 2 cm intervals (as described in chapter two). The results of initial sediment analyses, given in appendix A, show the percent sand-silt-clay within each sample, sample thickness, and sample depth at site irs1. If the three sediment components of individual samples are plotted against depth from the delta surface, a number of interesting trends are revealed. Figures 4.5a, b, and c show the sand, silt, and clay content of samples from site irs1 against depth.
Figure 4.5. Plots of the variable content of a) sand, b) silt, and c) clay against depth from the tidal marsh surface at site irs 1.
Figure 4.5a shows that the amount of sand within Facies C deposits increases to 68% with increasing height in section (decreasing depths), to a point 24 cm below the delta surface. Above this point the amount of sand rapidly declines to a minimum of 40% at the surface. The plot of silt content in Figure 4.5b essentially mirrors the trend noted for sand content. Silt content decreases with increasing height in section from a maximum of 86% to 17% at a point 24 cm below the delta surface. At this depth, the trend reverses as silt content climbs to around 60% content (by weight) at the surface. The plot of clay against depth in section (Figure 4.5c) shows a similar reversal in trend to those discussed above, though the location of the turning point is different. The percent clay in each rhythmite increases with increasing height in section to a point 82 cm from the delta surface. At this point clay content exceeds 32%, but rapidly decreases in content with greater height in section. At the present delta surface clay content is less than 3% of the total sample. The variability in clay content from sample to sample is greater than similar small-scale variability recorded in the sand and silt subsamples.

It is clear from these three diagrams that there exist very distinct sediment-size trends at this estuarine location. It remains unclear, however, what mechanisms led to the abrupt change in sediment characteristics at these two recorded depths (76.5 and 18.5 cm above mean sea-level). I am unaware of any other studies which have performed such detailed size-analysis throughout the full depth of tidal marsh deposits. The work of Dalrymple et al. (1991) is of some relevance, however; these authors presented a depositional model of the vertical variations in sedimentation rate of sand and mud within the intertidal zone of Bay of Fundy unvegetated mud flats. They produced a conceptual model of sedimentation rate against elevation, which is shown in Figure 4.6. In this model, the changing rates of both sand and mud (combined silt and clay fractions) are shown relative to one another. If the known sedimentation rates of sand and mud fractions
Figure 4.6. Conceptual model of the sand, mud, and total accumulation within unvegetated mudflats of the Bay of Fundy. Modified from Dalrymple et al. (1991).
calculated from this study are plotted against known elevation, the pattern revealed (in Figure 4.7) appears quite different from that in Figure 4.6. If one ignores (for now) the reversing trend recorded at a depth of 24 cm, then data from this study oppose that of the conceptual model of Dalrymple et al. (1991). The disagreement may indicate that the underlying assumptions of the conceptual model are limited, and not necessarily applicable within other tidal systems, particularly those which are fluvially dominated. For example, Dalrymple and co-workers state that the concentration of sand decreases upward through the flow so that higher surfaces receive less sand than lower surfaces. The research of Rood and Hickin (1989) and Babakaiff (1993) has shown that this assumption is not valid within Squamish estuarine environment, where macroturbulent eddies are able to transport coarse channel bed sands to the upper surface of flow. In addition to this, it is unclear whether one conceptual model of sedimentation can be applied within vegetated and unvegetated environments.

Despite the dissimilarity of these two diagrams, I would agree with the observation of Dalrymple et al. (1991) that "the contrasting behaviour of the sand and mud fractions is a direct consequence of their different settling rates" (p. 155). As sand rapidly settles from suspension, the deposition of this material is controlled by its availability within the marsh system, which is in turn dependent on flow velocities. As the settling rates of mud are so much lower, however, their deposition is not primarily controlled by its availability (as one can assume that silt and clay are always present within estuarine waters), but by the length of time that these materials are allowed to settle through the water column.

The sediment-size trends recorded in this study reflect the changing relative influence of tides over time. These relative changes are presumably driven by the increasing height
Figure 4.7. Plot of the variability of individual sand and mud fractions with depth in section at site irs1.
of the marsh surface, and the decreasing depth and duration of tidal inundation. During
the colonisation or pioneer stage of marsh development, when vegetation is first
becoming established on intertidal sandflats, newly deposited sediment is dominantly
silt with an admixture of sand and clay. Silt dominates these early marsh deposits for a
number of reasons. Firstly, the marsh vegetation induces the sedimentation of fines by
decreasing flow velocities near the bed (marsh surface) and by directly sequestering fine
sediment from the flow. Secondly, fines become deposited on the surface through the
processes of flocculation and pelletisation which produce groups of small particles with
a greater weight and higher settling velocity. Thirdly, and perhaps most importantly, the
sedimentation of silts is directly related to the depth and duration of tidal inundation. As
tidal amplitude decreases, the sediment-laden water column overlying intertidal surfaces
becomes shallower, and surfaces become inundated for a lesser period of time. The
deposition of fines will be more highly influenced by tidal amplitude and duration than
will the deposition of sands. This is because sands will become deposited relatively
quickly from the decreasing velocity flows which enter the marsh system. Because of
this, the deposition of sand is not highly dependent on duration of inundation. Once all
sand-sized material falls from suspension, increased sedimentation of sand cannot
continue, regardless of whether the delta remains inundated or not.

This argument is of course a little simplistic, as sands may be transported continually
into the marsh environment throughout the period of inundation. If we assume, however,
that flood velocities are too low to transport sands over the majority of the rising tide (all
but the early stage), then increasing tidal depths and durations can only influence the
sedimentation of fines. As the duration of inundation increases, the effective
sedimentation depth of the water column becomes greater. The effective sedimentation
depth is defined here as that portion of the water column which is able to contribute to
deposition. Clearly, for a given water depth the effective sedimentation depth decreases with decreasing sediment size. Thus, when the elevation of the marsh surface is low, fines will be preferentially deposited in relation to sands. As the marsh surface is raised, however, the deposition of fines will decrease while the deposition of sands will remain fairly constant. This will result in a decreasing amount of fines present in beds, and a relative increase in the amount of sand. Under this proposed scheme, the increasing percentage of sand with increasing delta height does not necessarily reflect an absolute increase in the amount of sand but a relative increase induced by decreasing amounts of fines. While this proposed explanation for the increasing sand content upsection is favoured, it is likely also that there is a real (not simply relative) increase in sand content upsection produced by the greater flood discharges required to inundate the rising delta surface.

The change in trends which occurs at a depth of 24 cm may also be driven by the relative proportions of sand and fines. It is suggested that the decrease in the amount of sand at this elevation is physically real, and caused by decreasing floodwater depths and velocities. Eventually, the delta will build to a height above which floodwater depths and velocities are too low to allow the transport of large amounts of sand onto and over the delta. This decrease in the amount of sand will lead to a relative increase in the amount of silt recorded in sediment samples. This is thought to account for the increase in the importance of silt within samples above 24 cm depth. This increased silt content is not considered to have been caused by dyke construction and the associated increased water depths.

Figure 4.5c shows that the pattern of clay content within each sample does not agree with the pattern shown by mud content (Figure 4.7). This highlights the fact that the
hydrodynamic factors that govern the deposition of clay are not the same as those which
govern the deposition of silt. Comparison of these combined fractions (termed 'mud')
requires far less laboratory analyses, but will result in decreased precision of data, and
may lead to loss of accuracy of interpretations based on these data. When clay content is
examined in isolation, there appears to be a specific height at which clay is preferentially
deposited. This height lies between 70 and 110 cm below the delta surface (30.5 and -9.5
cm a.m.s.l.), although there is a considerable degree of variability within this region.
Both above and below this region the amount of clay in samples decreases. The decrease
with depth (lower elevations) probably reflects the higher flow velocities within these
deep waters over the delta surface. These velocities are high enough to prevent the
deposition of considerable amounts of clay. As water depths decrease (as the delta is
raised), flow velocities decrease, permitting the deposition of greater amounts of clay.
This increase in clay content may also be related to increased vegetation establishment
and density. Above a certain height, however, clay content begins to decrease. This
decrease is presumably driven by decreasing durations of flood inundation, and related
to the settling lag effect (Van Straaten and Kuenen, 1957, 1958). The settling lag is that
time between the moment at which flow is no longer able to suspend particles of a given
size, and the time these particles become deposited. If the duration of floodwater
inundation is less that this settling lag time, clay will not be deposited as individual
grains. As clay has a far slower settling rate than silt, the critical sedimentation duration
(the period of time before which no deposition can occur during the tidal cycle) is
reached more rapidly than that of silt. The critical sedimentation duration of clay will
occur at lower marsh elevations than that of silt or sand. This process explains why the

1Despite this, some clays will become deposited in the form of flocs and faecal pellets, or may
adhere to grass stems.
maximum clay content peaks lower in section than that of silt, and why the maximum sand content peak is highest in section.

Figures 4.5a, b, c, and 4.7 all record greater scatter of data at greater depths in section. The decrease in scatter toward the delta surface is not a physically real decrease in variability of these data, but is produced by the sampling technique. Toward the top of the section, samples incorporate a number of beds (perhaps 10 or more). This causes an averaging of data, thus decreasing bed-scale variability of sediment characteristics, creating a smoother line.

**Fluvial Rhythmites**

Sediment samples of fluvial overbank rhythmite deposits were collected at site ol1, near the location of Figure 4.1, which is shown in Figure 4.2. In total, 42 samples have been analysed, which does not include all the rhythmite beds observed at this site, but does represent the majority of this facies thickness between 463 cm and 622 cm depth. Where deposits displayed no rhythmicity, sediments within these zones were sampled every 1 cm, but these samples have not been chosen for detailed sediment-size analysis here.

The sand, silt, and clay content of these 42 rhythmite beds have been analysed, and these data are presented in tabular form in appendix B. These data have been plotted against depth from the floodplain surface in Figures 4.8a, b, and c. These plots differ markedly from those of Squamish tidal marsh rhythmites which show clear, distinct trends with depth. Figures 4.8a, b, and c show that both sand and silt content show great variability with depth, but show no systematic changes such as those recorded within estuarine deposits. The sand data (Figure 4.8a) do suggest a general trend of an increasing sand
Figure 4.8. Changing content of a) sand, b) silt, and c) clay content with depth in section. Samples taken from the fluvial fine deposits at site 011.
content upsection to a depth of around 510 cm and decreasing sand content above this point. This general trend appears mirrored in the plot of silt content with depth (Figure 4.8b). These trends may not necessarily be real, but may in fact be apparent trends produced by the absence of data at three locations in section. These missing data represent those places where rhythmite bedding was not present in the fine-grained channel-fill deposits. Despite this, it is clear that there is a region toward the centre of these deposits where sand content is far greater than at any other depth in section at this site. These higher values are recorded over 18 different rhythmite beds, where sand content averages 98%. Differences in sand content are produced by differing magnitudes of riverine flood events, though it is unclear why so many high values are recorded at this particular depth.

Perhaps the most interesting result of these analyses is the difference in total sand and clay content. The sand content of fluvial overbank deposits is both higher (maximum 99%) and more variable (range of 75%) than the deposits analysed in Squamish estuary, where maximum sand content is 68% and the range of values is 64%. In contrast, the clay content of fluvial fines are far lower (average 2%) than the clay content (average 16%) within tidal fines.

The difference in clay content between tidal and fluvial deposits is considered to reflect the differences in the characteristics of flow within these two environments, primarily differences in discharge-stage relationships. The low clay content of fluvial overbank fines is most unlikely to reflect a lack of clay-sized sediment within Squamish flow at this location. Given this fact, the lack of clays at this site indicates that this material failed to be deposited or has since been eroded or winnowed-out. Given the semi-closed nature of this depositional environment, one would expect to find a far greater
concentration of clays than has been found here. Given this lack of clays, one can conclude that these deposits did not form within a lacustrine environment, so the cutoff was not constantly ponded. Within the more open depositional environment of the interdistributary bay, clay content has been found to be much higher (to a maximum of over 32%). This higher content primarily reflects the tidal modification of Squamish flow, which becomes more marked away from the main channel. In addition to this slowing of flow caused by tidal flux, the sedimentation of fines is enhanced by the baffling effect of vegetation, which further retards flow. This process likely is of greater importance in the estuarine environment, as stratigraphic evidence suggests that vegetation densities are far greater here than those in the channel cutoff. Although the cutoff was not constantly ponded, floodwater inundation was perhaps too frequent to allow the colonisation by terrestrial vegetation.

**Bedding**

The bedding characteristics within tidal marsh and fluvial overbank fines have been examined in detail, and analysed in relation to the sediment-size characteristics of these deposits. Results of this analysis are presented below.

Figure 4.9 shows the changing character of rhythmite bed thickness throughout much the depth of tidal marsh Facies C. In this figure only those individual rhythmite beds that could be detected in section have been included. The thickness data presented in this figure are not those values determined during sediment-sampling (presented in appendix A) but have been determined from close-range photographs from site 1rs1. Rhythmite bed thickness and depth data determined in this way differ slightly from those data presented earlier because the same location in section could not be analysed. These data
Figure 4.9. Decreasing thickness of individual rhythmite beds with depth in section. Data taken from photographs near irs1.
(presented in appendix C) are considered of greater accuracy than those determined during sediment-sampling. It is clear from this figure that rhythmite thickness decreases with increasing elevation of the marsh surface. This trend is clearly seen in exposed sections of this marsh facies (Figure 4.10). There is a strong negative correlation between sedimentation rate and elevation. Toward the base of Facies C (the transitional zone to Facies B) rhythmite thicknesses are more variable. The decreasing thickness of rhythmites upsection reflects the decreasing tidal influence on sedimentation with increased elevation of this intertidal marsh. As surfaces are raised, both the depth and duration of inundation decreases, thus reducing the sediment supply to the marsh.

Rhythmite bed thickness against depth in section are shown at two locations within the fluvial overbank fine deposits in Figures 4.11a and 11b. As before, the full thickness of these deposits are not represented here, as rhythmic alternation of inorganic and organic layers was not always present at these locations. In contrast to the simple, predictable decrease in bed thickness noted in tidal deposits, bed thicknesses in the fluvial fines show no relationship with depth. Instead, bed thicknesses at both locations appear random, and are highly variable over small depths. This bedding form is thought to develop because sedimentation is driven primarily by random (Thomas et al. 1987) riverine flood events rather than from regular cyclic tidal flood events. These differences in bedding form are considered a potentially important environmental indicator, and so will be discussed in greater detail later in this chapter.

The above section has shown the importance of Facies Changes in bed thickness. It is also important to consider these changes in thickness in relation to the changing sediment-size characteristics of beds at a particular location. Syvitski and Murray (1981) report that sedimentation rates of suspended-sediments within Howe Sound fjord are a
Figure 4.10. Tidal marsh rhythmites beds exposed in bank sections along the west delta between sites wd13 and wd14. With increasing elevation in section, bed thicknesses decrease to the point where individual beds are no longer visible. This is not the same location as that sampled for sediment analysis. Shovel is 1 m in height.
Figure 4.11. Changing thickness of individual rhythmite beds with depth in section within the fluvial environment. Data from a) site 011, and b) cb2.
function of the coarse end of the particle-size distribution. Dalrymple et al. (1991) similarly comment that differences in sedimentation rates are driven primarily by the amount of sands deposited, as there exists a strong positive correlation between the sedimentation rate and the sand content of deposits. Dalrymple et al. (1991) determined sedimentation rate from vertical and horizontal variations in bed thickness within deposits of the intertidal mudflats.

The sedimentation rates (or at least relative sedimentation rates) of Squamish tidal marsh rhythmites can similarly be determined from an examination of variations in bed thickness. The changes in bed thickness record a gradual decrease in sedimentation rate with increased height. Figure 4.12a shows a plot of the percent sand content of each rhythmite against rhythmite thickness (data not including the 1 and 2 cm sampling). This figure shows that, contrary to the findings of Dalrymple et al. (1991), bed thickness is not dependent on sand content. Figures 4.12b and 12c show that no relationships exist between bed thickness and any of the constituent fractions of sediment content. The same results were found on comparison of rhythmite thickness and sediment content within fluvial rhythmites.

In places, organic layers are seen to pinch-out altogether, thus breaking the rhythmicity produced by alternating inorganic and organic layers. At these locations, fluvial fines lose all resemblance to tidal marsh deposits. This lateral discontinuity of organic layers is most probably produced by partial erosion of the litter layer. This could easily have been achieved by flows entering the cutoff which were not flowing fast enough to erode the underlying sediments. If litter were dry it could also have floated off the surface and transported away by flows.
Figure 4.12. Relationships between the varying thicknesses of tidal rhythmite beds and their a) sand, b) silt, and c) clay content at site IRS1.
Discussion of the Genesis of Fluvial Rhythmite Beds

In conclusion, the fine grained deposits seen at the base of certain sections in the fluvial reach of Squamish River likely formed within an abandoned channel section, most probably an oxbow cutoff. Sedimentation occurs within this environment when floodwaters enter the channel section, carrying fine sediment in suspension. The deposition of fines from suspension occurs as the channel reach is temporarily ponded. As sedimentation is driven by input from the adjacent main river flow, these random flood events control bed thickness and grain-size. As such, adjacent beds display considerable variability in thickness and sediment-size characteristics, and appear to lack any order (such as thinning bed thickness upsection).

During periods of low river flow, much of the abandoned channel bed will be above water level. Sedimentation during these between-flood periods will be minimal, and, given a long enough period of non-deposition, a litter layer of leaves, needles and twigs will build-up on exposed surfaces. This litter layer may be removed during the proceeding flood period, but is more likely to be incorporated into the deposit as rapid sedimentation buries the organics. Given this proposed origin of organic beds, one rhythmite bed may not necessarily represent deposition over one year, as is the case for tidal rhythmite beds. Instead, a number of beds may be laid down in any given year. The absence of organic layers could indicate that litter layers either failed to form, or were eroded so that the inorganic layer represents a number of depositional events. Such an occurrence could also have been produced by rapid sedimentation from one large flood event. One can see from this discussion that fine-grained rhythmite deposits may form within purely fluvial environments, but that they may not necessarily do so.
This depositional model is quite different than that proposed for the rhythmite beds of tidal marsh sequences (presented in chapter three). These differences account for the dissimilarities in bedding form and grain-size characteristics of tidal and fluvial rhythmite beds described here.

**Sediment Accumulation in the Tidal Zone:**

**Observations from Bedding Form**

The upsection decrease in tidal marsh rhythmite bed thickness (or sedimentation rate) recorded in this study has previously been reported in other tidal marsh environments (Richards, 1934; Chapman, 1938; Bouma, 1963; Redfield, 1972; Richard, 1978), but the possible significance of this trend has never been fully investigated. Redfield (1972) presented a graph showing the cumulative number of rhythmite beds (his strata) observed in a marsh bank, against depth from the marsh surface. Redfield's data (taken from a New England salt marsh with a tidal range of around 3 m) are shown in Figure 4.13. This plot of the cumulative number of beds at different depths essentially records the changing rate of sedimentation over time. Higher sedimentation rates will produce a steeper line and constant sedimentation rates will produce a straight line. Figure 4.13 shows a clear non-linear relationship between the rate of sedimentation and marsh elevation. Sedimentation rates decrease with increasing age (elevation) of the marsh system. These data provide the opportunity to make comparison of the changing nature of sedimentation with elevation between Redfield's study and data from this study.

Because of differences in the total age of deposits under investigation, and differences in sedimentation rates, both sets of data need to be transformed in a number of ways to enable valid comparison. The tidal marsh rhythmite deposits in Redfield's (1972) study
Figure 4.13. The cumulative number of rhythmite beds (strata) against depth from the marsh surface (in feet). Data from a New England salt marsh (Redfield, 1972).

Figure 4.14. Comparative plot of the changing rate of sedimentation through time between data provided by Redfield (1972) and data from this study. Both the rate of sedimentation and time of marsh development are expressed in relative, dimensionless terms.
are 184.0 cm thick, and represent deposition over 92 years (total number of beds), which is an average rate of accretion of 2.00 cm/yr. By comparison, the thickness of detectable rhythmite beds in this study is 117.4 cm. These have accumulated over 121 years at an average rate of 0.97 cm/yr. These depth and age data have been transformed into dimensionless expressions of sedimentation rate and time. Sedimentation is expressed as a proportion of total deposit thickness, and time is expressed as a proportion of the total age of deposits. These two relative scales yield a dimensionless expression of the changing nature of sedimentation over time, enabling a simpler, more meaningful comparison of deposits of different age and thickness.

The results of this comparison are shown in Figure 4.14, which displays an almost perfect agreement of the transformed curves derived from both studies. These curves show that sedimentation rates are greatest during the early stages of marsh development, and that with increasing time, this sedimentation rate gradually slows. The close agreement of these curves suggests that the changing nature of sedimentation over time may be the same for a number of different marsh environments, irrespective of local differences in sediment supply and energy conditions. The nature of the decreasing sedimentation rate likely is produced by the nature of tidal flux. Tidal marsh sedimentary sequences record decreasing durations of tidally-driven inundation, and decreasing depths of the inundating water column. Decreasing depth and duration is driven by the increasing elevation of intertidal surfaces, the effect of which is illustrated in Figure 4.15. This figure shows the percentage of tides which equal or exceed given heights above mean sea-level. From this, one can see that as the marsh surface is raised from 0 m to 1 m a.m.s.l., the total number of tides which will inundate this surface decreases by 39% (from 62 to 23%). This represents a dramatic decrease in the number of events which are able to contribute to continued marsh accretion.
Figure 4.15. Distribution curve of tidal inundation within Squamish estuary. This curve shows the decreasing percentage of tides which are able to flood marsh surfaces of increasing elevations above mean sea-level.
This assertion may be tested by comparing sedimentation rates to marsh surface inundation rates at different elevations. This comparison is shown in Figure 4.16, which shows the relationship between the relative rate of sedimentation of Squamish tidal marsh, and the percentage of tides which inundate the marsh surface. These percentages have been calculated from combined hourly tide data at Point Atkinson for the years 1991-1993 (data provided by Rick Thomson of the Institute of Ocean Sciences). These data were also used to plot the tidal height distribution of Figure 4.15. The cumulative percentages of inundating tides were calculated for the known heights (in cm above mean sea-level) of rhythmite beds. Figure 4.16 shows that there exists a very strong correlation between the number of tides which inundate the marsh surface, and the rate of marsh accretion. Harrison and Bloom (1977) presented similar evidence of this relationship. These authors observed that average rates of accretion at five sites correlated well ($r=0.90$, $p=0.05$) with the mean tidal range at each site.

The correlation shown above, however, is not linear, as relative sedimentation rates decrease slightly toward the top of the tidal marsh deposits (as is shown by the lack of fit to the straight line). There are thought to be at least two explanations for the non-linearity displayed here. Firstly, sedimentation rate is not simply a function of the number of tidal events which flood the marsh, as suspended-sediment concentrations are not always constant with depth. Below a certain depth, flow velocities decrease to the point that sand is no longer transported onto and across the delta. The specific depth will presumably be dependent on the relative roughness of the marsh surface (vegetation density and height) and the turbidity of flows adjacent to the marsh. This is only likely to become important toward the upper limit of the tidal marsh deposits. In addition to this, as water depths decrease, inundation occurs closer to high tide. This results in decreasing durations of inundation (as well as depths) with increasing height of the
Figure 4.16. The decreasing percentage of tides which inundate the marsh surface as that surface accretes. With decreasing tidal influence the sedimentation rate decreases, though not at a uniform rate.
marsh. As flood durations decrease, lesser amounts of sediment are able to fall from suspension, resulting in decreasing sedimentation rates. The tide data used here (cumulative number of observed tides (recorded hourly) which exceed given heights above mean sea-level) are not flexible enough to record these changing durations of inundation.

This last point raises an important limitation of the data presented here. One would expect the tide data presented in Figure 4.15 to be slightly inaccurate in Squamish estuary, as inundation of marsh surfaces is not a simple function of tidal flux. Both the depths and durations of floodwaters over Squamish marsh surfaces are influenced by discharge fluctuations of Squamish River and wind set-up in Howe Sound, which operate over a number of time scales. If discharge is low then certain tides may fail to inundate the surface, but if discharge is high enough the surface may be inundated throughout most or all of the tidal cycle. For data to be fully comprehensive, one would have to determine the actual duration of inundation at given tidal heights, over a large enough period to record the effects of the majority of discharges experienced by Squamish River.

If the above interpretation is valid then tidal marsh deposits may be identified as tidal in origin from an examination of bedding characteristics alone. While it certainly is not true to state that decreasing bed thickness upsection is a tidal characteristic, it may be shown that the rate of that decrease with height is a characteristic of tidal environments. For this reason it is important to examine the character of bedding produced by differential sedimentation within a number of different environments. Unfortunately, published data of the kind required for this purpose are very limited, though Nanson (1977) provides some useful data. Nanson (1977) produced an average sedimentation
curve for sandy floodplain deposits of the Beatton River, British Columbia. This curve, which is shown in Figure 4.17, was determined by combining sedimentation rates determined from changes in poplar wood anatomy, with sedimentation rates determined from changes in floodplain elevation. Changes in floodplain elevation with time were determined from averaged data from 10 locations. These combined data provide average rates of accumulation of sediments of known depths, over a period of 400 years. The average annual rate of sediment-accumulation for Beatton River floodplain deposits is 1.70 cm/yr.

Nanson's (1977) average sediment-accumulation curve has been used here for comparative analysis of the character of sedimentation rate with time, between fluvial and tidal environments. The entire data set has not been used here: only the first 235 years (400 cm depth) of floodplain development are used, for two reasons. This point represents an inflection point in the sediment-accumulation curve (identified by Nanson (1977)) beyond which floodplain-building occurs at a negligible rate. This negligible rate of sedimentation is comparable to the very slow rates of tidal marsh accumulation as the marsh approaches mean high-water, as evidenced by the lack of visible rhythmite beds toward the top of marsh facies. The 235 year range of data used in this analysis was also chosen to ensure that the deposits within this fluvial environment reflect almost the entire range of depositional events experienced by this river system. This is considered essential if fluvial deposits are to be compared with deposits in a tidal environment, which are known to have developed under the full range of tidal flows.

The results of this comparative analysis are shown in Figure 4.18. All data have been transformed into dimensionless relative values. The curves produced in this way show a distinct difference between the nature of sedimentation rates within the fluvial and tidal
Figure 4.17. Changing rate of sedimentation of Beatton River floodplain over a period of 400 years. Sedimentation rates (determined from changes in poplar wood anatomy and floodplain elevation) decline rapidly after 50 and 235 years. From Nanson (1977).
Figure 4.18. Comparative plot of the changing sedimentation rates of tidal marsh and fluvial floodplain deposits over time. Data from this study, Redfield (1972), and Nanson (1977).
environments investigated here. Within the fluvial environment, sedimentation rates are extremely high during the early stages of formation of the floodplain, in comparison with a far more gradual rate of tidal sedimentation. To highlight this difference, half the total thickness of Beatton River floodplain deposits formed in under 12% of the time taken for the entire floodplain deposits to form. In Squamish estuary, half the tidal marsh rhythmites formed in 33% of the total time taken for these marsh deposits to form. After around 20% of the total period of Beatton floodplain formation, sedimentation rates rapidly decrease. Toward the last stages of floodplain formation, relative sedimentation rates follow the same curve as that toward the last stages of tidal marsh formation.

From the data available, there exist clear differences between the nature of sedimentation within fluvial and tidal environments. If the relationships highlighted here are shown to be representative of floodplain and delta marsh deposits in a number of environments then these results provide a further tool for environmental reconstruction. The following paragraphs outline a number of mechanisms which may be responsible for the different sedimentation rate patterns observed here.

I suggest that differences in the relative sedimentation between the fluvial and tidal environments examined here are driven by characteristic differences in effective discharge relationships, flood frequencies, and sediment transport mechanisms between these environments. The most effective flood discharges in rivers commonly are high frequency, low magnitude events. Whether or not the effective discharge equates with bankfull conditions is not important here, it is merely important that effective discharges have recurrence intervals of a small number of years (perhaps 2 or 3). Sedimentation rates within the fluvial environment primarily are driven by flow velocity. In contrast, sedimentation rates within the intertidal zone primarily are determined by duration of
inundation, as flow velocities become too moderated by tides to allow the transport and deposition of bedform sands. This could account for the lack of relationship between sand content and bed thickness of tidal marsh rhythmites. Because tidal sedimentation is driven by flood duration, the most effective tidal discharges are low frequency, high magnitude events. During these periods water depths are greatest, and the durations of marsh inundation are the longest, allowing greater sedimentation from suspension.

The differing sedimentation rates between fluvial and tidal regimes are not simply a function of flow velocities, however, but are also a function of differences in relative flood frequency. Although sedimentation within the intertidal environment is greatest during flood events which occur least frequently, in relative terms these events occur very frequently over highly predictable time scales. Because of the nature of tidal flux, the highest tidal events occur fortnightly during the Spring tide. Within fluvial environments, however, flood discharges of similar relative magnitude have recurrence intervals which likely are in excess of 200 years. It is this regularity of tidal flows which accounts for the characteristic decline in sedimentation rates with increasing age of the tidal marshes shown in this study. During the early development of the marsh, regular inundation by the full range of tidal flows ensures rapid rates of sedimentation. As the marsh elevation is raised, however, fewer tides are able to inundate the surface, and the largest tidal floods become less effective depositional events, as flood depths and durations decrease. Because of the regularity of tides, the decreasing rate of marsh inundation is gradual, producing a gradual decrease in sedimentation rate, as seen in Figure 4.16. In those tidal environments where sediment input from other sources is minimal and relative sea-levels are stable, sedimentation will slowly continue to the elevation of maximum tidal amplitude. Without a change in the elevation of the marsh in relation to mean sea-level, continued marsh accretion cannot occur above this point.
Within the fluvial environment examined here, flows clearly are competent to initiate considerable bedform transport. Because of this, flow velocities are considered to exert greater influence on sedimentation rates than are flood durations. Because of this and the fact that the most effective flood discharges are those of relatively high recurrence, sedimentation rates are very high during the early stages of floodplain development (when floodplain elevations are low). During these early stages, floodplain deposits are produced by within-channel flows. In contrast to the tidal marsh environment, the most important flows (for floodplain formation) in fluvial environments are not the deepest flows experienced by the river. Although these flows deposit considerable amounts of material, they occur too infrequently to exert major influence. Instead, the most important flows are those of intermediate depth (high recurrence interval events). As floodplain surfaces continue to accrete, the surface will eventually exceed the depth of flow associated with these dominant high frequency events. Above this elevation, the dominant mode of sedimentation will shift from bedload to suspended-sediment deposition. This shift in sedimentation mode essentially reflects the transition from within-channel to out-of-channel deposition. This transition is expected to occur rather abruptly, and most likely accounts for the rapid decrease in sedimentation rate recorded in Figures 4.17 and 4.18. As the floodplain continues to accrete beyond this point, sedimentation rates will continue to decline (though at a more gradual rate) because of the decreasing frequency of flooding at these higher elevations.

Nanson (1980) presents some evidence which agrees with the fluvial depositional model outlined above. He states that the lowest sediments in floodplain sequences are not overbank deposits, but are the vertical component of lateral accretion which occurs within the channel during periods of high flow. This sedimentation forms the lowest deposits in floodplain sequences, and accounts for the rapid sedimentation rates
observed for the first 50 years of floodplain formation (Figure 4.17). Above this point, the sediments in floodplain sequences are overbank deposits, which as Nanson (1977) shows, develop at a far slower rate (Figure 4.17). Nanson's (1980) interpretation of this sedimentologic evidence and sedimentation rate data, however, differs from that given here. He suggests that the rapid decline of sedimentation rate after 50 years of floodplain formation (indicated in Figure 4.17) is attributable to a lateral shift in the location of maximum within-channel deposition. Nanson argues that this shift is caused by the establishment of a new scroll bar closer to channel flow. This scroll bar captures the majority of sediment, so that continued accretion of the floodplain occurs through suspended-sediment deposition. The formation of a new scroll bar closer to flow records channel migration away from this region of floodplain accumulation.

I have suggested here that the mechanism responsible for the decrease in sedimentation rate need not necessarily have been a lateral shift in the zone of maximum deposition associated with channel migration. Such a trend could have been produced by a shift in the dominant mechanism of sediment deposition (bedload to suspended-load) associated with an increase in floodplain elevations above the limit of most effective flood depths.

The models presented above clearly are limited by their failure to consider changes in suspended-sediment concentrations, and by consideration of the flood flow velocities and durations in isolation. Despite these limitations, the models presented above provide a theoretical framework to aid interpretation of the relationships between flow (in fluvial and tidal regions) and bedding form (used here as a surrogate for sedimentation rate). To test the validity and accuracy of these models, there is need for examination of bedding form within a number of different tidal marsh systems, and other depositional environments. For example, it would be interesting to examine the relative
sedimentation curve produced by lacustrine fill deposits, where annual beds commonly are easily detectable. Lacustrine deposits may be quite similar to those of tidal origin, as the decreasing depths of lake water (with continued sedimentation) may be similar in form to the decreasing depths of tidal inundation (with higher elevation of the marsh). It would also be interesting to determine whether the nature of sedimentation of intertidal unvegetated sand and mud flats agrees with the findings from the vegetated tidal marshes investigated here.
CHAPTER FIVE
ESTUARINE CHANNEL EROSION
AND TIDAL MARSH ACCRETION

This chapter discusses a number of erosional and depositional processes operating within the estuarine reach of Squamish River. Discussion will focus on observations made in the field, data collected during the field season, and also on data extracted from historic photographic evidence. Squamish estuary is characterised by a history of rapid distributary channel shifting and lateral instability of the river. The purposes of the following sections are to determine the rate and nature of channel bank erosion within the estuarine reach of Squamish River.

Analysis of Historic Rates of Bank Erosion

Historic rates of estuarine bank erosion have been determined from aerial photographs dating from 1957 to 1990. In an attempt to determine the influence of dyke construction on rates of erosion, pre- and post-dyke data are treated separately. A total of 13 photographs were traced then scanned as described in chapter 2, although only 8 of these could be used in the analysis due to poor fit of the scans (excess error). The results of the analysis of these scans are discussed below, and have been summarised for the periods 1957-72, and 1972-1990.

Erosion along the lower 4 000 m of Squamish estuary is analysed within 4 zones. These zone boundaries were determined from examination of aerial photographs over the 33 year period of study, and reflect the importance of channel morphology as a determining factor for both the location and rate of erosion. As such, the 4 zones identified here
follow the length of individual meander arcs, which have been further subdivided into zones along concave and convex banks. The locations of these zones may be seen in Figure 5.1. These zone boundaries are dynamic features because they must shift as the channel migrates. Despite this, however, little to no boundary shifting has taken place since 1957. Examination of aerial photographs from the years 1957, 1969, 1973\(^2\) and 1990 (shown in Figures 5.2 and 5.3) yields qualitative evidence of the amount of erosion which has taken place within the 4 identified zones over this 33 year period.

The areal loss of floodplain and delta material over the periods before and after dyke construction have been determined within each of these reaches. Although determination of the volume of eroded material would be of greater significance than areal loss, it was not felt that these data could accurately be determined. As no detailed channel cross-sections could be made at the time of data collection, it was felt that the estimation of channel depth (the equivalent depth of erosion) would introduce too much inaccuracy to be of real use. Also, eroded bank material is not all completely removed from the near-bank region, but may form a wide submerged ledge. Although bank erosion is known to occur very rapidly, it is not known what time scales are required to remove slumped material from the near-bank zone.

Table 5.1 shows the areal loss of material within each specified zone over a number of time-spans before and after dyke completion. Yearly average losses were first calculated as monthly averages for the periods between photographs, and these data were then changed to yearly averages. The data presented in this table in and Figures 5.2, and 5.3

\(^2\) Although erosion rates were analysed over the 1969-72 period, the aerial photograph for 1973 is shown here because of technical difficulties involved in the presentation of the scanned image for 1972.
Figure 5.1. Location of zones within which erosion rates are assessed for the period 1957 to 1990. Over this 33 year period of study these zone boundaries have remained quite stable, though they can be expected to be dynamic features in the longer term.
Figure 5.2. Photographs of Squamish estuary and delta taken in 1957 and 1969, showing the general changes in channel form and associated regions of bank erosion.
Figure 5.3. Photographs of Squamish estuary and delta taken in 1973 and 1990, showing the changing channel morphology and the associated areas of bank erosion.
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| PRE-DYKE |         |         |         |         |
| 1957-64  | 21 826  | 3 214   | -        | -        |
| 1964-69  | 11 709  | 2 342   | 13 090   | 2 618    |
| 1969-72  | 74 163  | 25 427  | -        | -        |

| POST-DYKE |         |         |         |         |
| 1972-73  | 2 189   | 1 812   | -        | -        |
| 1973-78  | 13 090  | 2 586   | -        | -        |
| 1978-82  | 5 993   | 1 563   | -        | -        |
| 1982-90  | 22 376  | 2 783   | 14 726   | 1 831    |

| PRE-DYKE average 57-72 | 107 695 | 7 322 | 13 090 | 890 |
| 1957-69 | 33 535 | 2 844 |

| POST-DYKE average 72-90 | 43 648 | 2 405 | 14 726 | 812 |
|                        | 40 906 | 2 254 | 137 764 | 7 592 |

| WHOLE PERIOD average 57-90 | 151 343 | 4 607 | 27 816 | 847 |
| av. (57-69 + 72-90) | 77 180 | 2 578 |

Table 5.1. Rates of erosion along four zones within Squamish estuary over a 33 year period. Data are expressed as both total areal loss and yearly average loss over a number of timespans. The symbol (-) indicates that no erosion was detected from scanned images, though it is seems highly unlikely that erosion did not occur.
show that Squamish River has been laterally shifting and eroding banks at a considerable rate over the 33 year period of record. The following is a summary of trends observed in the data for each of the 4 zones.

1) Squamish west delta (zone 1) is decreasing in size as the lower reach of Squamish River migrates westward. Rates of erosion within the time-spans under study record considerable variability which primarily is caused by the very high rates of erosion (over 25 000 m$^2$/yr) over the period from 1969 to 1972. This high value, however, is not considered comparable to data from other periods because it not only records possible natural erosion but also records the direct loss of delta material through dredging. Levings (1980) reports that some loss of west delta sediments occurred because of engineered attempts to straighten the lower channel. As no erosion was recorded within any of the other 3 zones, the material eroded from zone 4 most likely represents that portion of the delta which was removed during channel straightening. If the erosion value for the 1969-72 period is not included then pre-dyke rates of erosion average 2 844 m$^2$/yr.

As an average over the 33 year period of study, the west delta has been eroding at a rate of over 2 578 m$^2$/yr. This value does not include that area of material artificially removed in 1972. Comparison of pre- and post-dyke rates of erosion reveals that erosion rates have remained essentially constant throughout all periods of study, indicating that erosion of the west delta has not been exacerbated by dyke construction, though it clearly has been greatly increased because of dredging. Erosion rates are in fact seen to decrease somewhat from 2 844 to 2 405 m$^2$/yr. Figures 5.2 and 5.3 reveal that erosion has taken place along the entire 1 700 m length of the west delta.
2) **Zone 2** has experienced little erosion over the last 33 years, as would be expected at this inner meander bank. The erosion that has been recorded along this reach has been quite localised, and as an average, has been decreasing over the last 2 decades. Erosion in this zone is unchanged by dyke construction. In fact, photographic evidence indicates that deposition and floodplain growth is actually occurring within much of this zone, as can be seen on comparison of Figures 5.2 and 5.3. The area of deposition along this reach records the colonisation of a large point-bar surface. Unfortunately, the true rate of deposition cannot accurately be determined because of differential exposure of the unvegetated bar surface.

3) **Zone 3** has undergone considerable erosion (an average of 3 138 m²/yr) over the last 33 years, although data in Table 5.1 indicate that the rate of erosion has decreased considerably since dyke construction. The erosion recorded within this zone is produced as the meander arc tightens at this location. The recent slowing of this rate of erosion may indicate that the meander has exceeded some critical $r_m/w_m$ value as identified by Hickin (1974). This possibility has not been examined in detail, however, as there is no reason to assume that the character of estuarine meander growth is similar to that of fluvial meanders.

4) **Zone 4** is perhaps the most important study area, as it provides information regarding the stability of the river training dyke. One can see from the proximity of the dyke to the river along this zone (seen in Figure 5.4), that this structure may be threatened by severe bank erosion and eastward migration of this meander. This zone includes the areas of intense erosion identified by Bell (1975), Zrymiak and Durette (1979) and Levings (1980). Over the 33 years of study, almost 170 000 m³ of material has been eroded from this zone, which is an average of over 5 000 m²/yr. This average
Figure 5.4. Showing the proximity of Squamish River to the river training dyke and the abandoned central channel. The outer bank of this meander experiences considerable erosion as this meander migrates both across and down valley. Note the unvegetated point-bar surface at this location.
value masks considerable temporal variability which is particularly evident upon comparison of pre- and post-dyke rates of erosion. This comparison reveals that rates have more than trebled since 1972, increasing from $2151 \text{ m}^2/\text{yr}$ over the period 1957-72 to $7592 \text{ m}^2/\text{yr}$ between 1972 and 1990.

While no attempt was made to calculate rates of erosion over smaller zones than those described here, it is evident that erosion (particularly within zone 4) is not constant throughout the meander arc. Throughout this meander, erosion is variable with location along the arc, as noted elsewhere by Hughes (1977), and the location of maximum erosion has not remained fixed throughout the period of study. Instead, the point of maximum erosion has shifted from the upstream end of the meander to a point several hundred metres downstream. This indicates that the meander is migrating across valley and is also shifting down valley. The zone of maximum erosion identified from aerial photographs (up to the year 1990) is labeled in Figure 5.4. This figure also shows the extent of the point-bar which is building along the inner bank of this meander.

When the average rates of erosion within each of the 4 zones are presented diagrammatically (Figure 5.5), a number of general trends both within and between zones are revealed. Although this figure is necessarily simplistic because of the averaging process, it is clear that individual meanders respond to changes in flow in different ways, and at different times. The 3 outer meander bends within this study reach (represented by zones 1, 3 and 4) do not necessarily all experience accelerated (or decelerated) erosion over a given time span. For example, as an average over the period from 1978 to 1982, zones 3 and 4 experienced greater erosion than over the preceding years, whereas zone 2 remained unchanged and zone 1 experienced a slower rate of erosion than previously. This highlights the fact that system response to fluctuating
Figure 5.5. Diagrammatic representation of erosion data presented in Table 5.1. Straight lines indicate an average rate of erosion over periods between aerial photograph coverage.
discharge occurs as a series of local hydraulic and planform changes, and that erosion rates determined for one meander need not be applicable to adjacent meanders (Ferguson, 1977). There is some limited evidence to suggest that changes in rates of erosion within zones 1 and 3 are negatively related, as the trends through these data sets generally mirror each other.

Figure 5.5 also shows that meander migration does not occur linearly, but occurs as a series of rapid shifts evidenced by peaks in the average erosion data, followed by periods of limited or greatly reduced channel activity.

Unfortunately, the aerial photograph for September 1991 was not usable in these analyses because the small scale of the photograph introduced an unacceptably large error. Despite this, detailed examination of this photograph reveals that considerable erosion took place within zone 4 over the 13 months since the 1990 photograph. From an examination of this channel bank reach during the field season early in May 1992, it was also clear that considerable loss of floodplain material had occurred in the previous 8 months. At its closest point, the basal section of the river training dyke lay within 5 m of the river bank in mid-1992. In addition to these qualitative observations, direct measurements of bank erosion within this zone were taken over the latter half of 1992 and on a number of occasions throughout 1993. These data are presented later in this chapter.

**Error in Erosion Data**

The errors intrinsic to aerial photographs primarily are those that occur as a result of the differential scaling across the ground extent of the photograph. Because of the
decreasing angle of view away from the centre, the surfaces recorded toward the edges of photographs become distorted or stretched relative to those surfaces in the centre. This results in a photograph scale which is not constant across the entire image. Unfortunately, it is extremely difficult to determine the order of error within data derived from an overplot of two superimposed images which were originally taken at different elevations and along different flight paths. However, those images which did record detectable distortion were not used in analysis. In 5 of the 13 scanned images, the reference points used to establish fit between different images could not be matched, and so were rejected.

The error which arises when data are extracted from photographs is more easily controlled and may be determined with ease. The main error is produced by the differential scaling of air photograph traces; even though traces are scanned then rescaled to some common size, trace scales are not exactly the same. Of the 8 scans which were used in erosion analysis, 4 different scales were represented (determined by comparison of areas on all scans with the known areas from Squamish area map). For the calculation of areal loss of delta and floodplain material, an average of these scales was used (mean scale = 17 535, S.D. = 404, n= 4). The maximum error which may arise from this use of an average scale is calculated as 10.8 %. This value is the difference in area calculated from the smallest and largest represented scales, expressed as a percentage of the largest value (Hooke and Perry, 1976). The calculated areas of erosion presented in Table 5.1 may then be expressed as accurate to within ±5.4 %. This value includes the error which results from the different scales of scanned images used, and also some of the error incorporated by the differential scaling (stretching) within each of the photographs. Because areal measurements of scale were determined as opposed to single length measurements, this latter error is recorded in two directions. This is
preferential because stretch may be greater in one direction than another. For example, because of plane roll, distances perpendicular to the flight path may exceed those parallel with the flight path.

Despite this inclusion, the error value of $\pm 5.4\%$ is considered something of an underestimate as it fails to consider scale differences across the whole extent of all photographs, and it also takes no account of minor errors incorporated when the original image was traced. Because of these factors, it is considered more realistic to report the erosion data as accurate to the order of $\pm 10\%$.

**Analysis of Short-Term Rates of Bank Erosion**

Erosion rates within zone 4 were determined over the period from June 1992 to August 1993. Bank retreat was recorded from a series of 64 erosion pins placed on the floodplain surface along that section of the meander marked in Figure 5.4. After their initial placement in June 1992, these pins were resurveyed in August, November, December 1992, and February, April, and August 1993. During this period of study, considerable erosion was noted within this zone, which (as previously noted from historic data) was highly discontinuous. The discontinuous nature of these data suggest a seasonal variability which likely is driven by the seasonal nature of flow of Squamish River.

Erosion was recorded over each of the 6 periods of study, though not all pins recorded floodplain loss. Over the summer months of 1992 and 1993, bank retreat was measured in decimetres when averaged along the entire length of this monitoring site. During the autumn of 1992, however, substantial bank retreat occurred within this zone. The time of
initiation of this erosion or its duration are not known, but by 11 November, 42 of the 64 erosion pins had been eroded and lost. Estimates of the amount of bank retreat over this 3 month period were able to be made from a series of accretion stakes which had previously been placed at 100 m intervals, 15 m back from the river bank. Two of these stakes had been eroded, while a third was found within a partially-slumped block of floodplain material by the river bank. From this evidence it is clear that bank retreat at these 3 locations exceeded 15 m, and this rate of retreat is considered representative of this 300 m section of bank. Erosion was also recorded at all pins both upstream and downstream from this section, though these recorded a bank retreat averaging 3.5 m. This material is considered to have eroded rapidly over a short period of time, in the manner described in the proceeding section.

Observations of Channel Bank Erosion

Field observation and erosion pin data suggest that the majority of channel bank erosion occurs during high discharge flood events in winter months. For this reason the process of bank erosion was rarely observed, with one notable exception. Late in June 1992, considerable bank erosion occurred at the upstream-most end of the west delta, as indicated in Figure 5.4. This erosion began abruptly and continued at a considerable rate throughout the months of July and August, along a section of river bank 100 m in length. Because of the speed of erosion within this localised region, the erosional process (shown dramatically in Figure 5.6) was easily observed. The following is a discussion of these field observations.

Prior to the initiation of erosion, the bank in this location resembled all others in this study reach, and in fact at most locations along the lower 2 500 m of Squamish estuary.
Figure 5.6. Showing the sudden collapse of marsh deposits toward the upper limit of the west delta. The location of this observed erosion is shown in Figure 5.4.
The upper two-thirds of the bank were vertical to the approximate depth of the marsh rhythms. The lower third of the bank (at low tide) was obscured by a large amount of slump block debris. Figure 5.7 shows this characteristic bank profile which is observed here immediately downstream from the eroding section of bank.

Although the point at which erosion began is not exactly known, by the morning of June 28, the bank face at this single location was rapidly eroding. At this location, the channel thalweg crosses over the channel and impinges on this outer concave bank section. The erosion is thought to have been initiated by high discharges which removed the protective toe of the bank. In this respect, the removal of intertidal marsh deposits appears dependent on basal endpoint control, as described by Carson and Kirkby (1972). Once erosion became initiated, the bank face had a very different appearance (Figure 5.8), as the slump material at the toe of the bank had been completely removed, leaving a vertical face. Note that the coarse angular material to the centre and left of the photograph is not naturally occurring, but material from an abandoned logging road.

While it appears accurate to state that erosion is initiated by the removal of the basal region of the bank (during the state of *excess basal capacity* (Thorne, 1982)), the nature and rate of that erosion is determined by a number of factors. For example, Figure 5.8 (taken in late July 1992) shows how erosion, or at least the degree of erosion, was highly influenced by stratigraphy. Once the slump material had been removed from the base of this section (that above low tide), a thick, very coarse sand layer became exposed. These coarse sands, shown in greater detail in Figure 5.9, are deposits of Facies A. Once this sand layer became exposed, it controlled the rate of erosion along this whole face. As these unconsolidated sands were eroded, the composite face became increasingly undercut until the weight of overlying material could no longer support itself. The result
Figure 5.7. A down-valley view taken at the upper limit of the west delta, downstream from the bank erosion along this reach. Note the blocky appearance of the slump material at the toe of the exposed bank.
Figure 5.8. Same location as that shown in Figure 5.6, taken one day after the initiation of erosion, showing that the protective toe of the bank has been removed, exposing an uncohesive sand layer (Facies A) near the water-level. A decaying boil feature may be seen in the near-bank region. Face is around 1.5 m high.
Figure 5.9. Close-up view of the coarse sand of Facies A exposed at the base of the eroding bank, overlain by fine sands and silts of Facies B.
of this lack of support was the sudden collapse (as a cantilever failure) of bank sections up to 5 m$^3$ in size. In this way, the cohesive nature of the overlying sediments controlled the nature of bank collapse (as noted elsewhere by Turnbull et al., 1966, and Hughes, 1977). Thorne (1982) similarly notes that in cases where cohesive sediments overlie uncohesive sediments, the mechanism of failure is determined by the thickness of overlying deposits. These large coherent blocks collapsed as a single unit, thus retaining their internal character and also often removing trees. In this way the size of these coherent blocks also may be determined by the size and character (number, depth, diameter etc) of the root systems of these trees.

Because of the location of this unit toward the base of the exposed face, erosion was also controlled by tidal cycle. No erosion was observed during the majority of the tidal cycle when stage was high, or at least higher than the upper limit of sand. All of the erosion that was observed over a number of weeks, took place within 1 or 2 hours of low tide. At these low stage periods the sand layer became exposed, and then became susceptible to erosion. Tidal stage may have further influenced erosion rates because of increased pore pressures within the bank during falling stage, and because of the exfiltration of water from the bank face (Barwis, 1978).

Another important point to note is that the erosion did not seem to be controlled by mean flow velocities of Squamish flow, as the river was not moving rapidly past the bank at this point at those times when the basal sands were exposed. Instead, most erosion appeared to be controlled by turbulent conditions within flow and by wind-generated waves on the surface of flow. Erosion was greatly increased during periods of turbulent activity at the bank face, where small eddies (bursting boil features) caused disruption of the water surface. One of these boil features may be seen in Figure 5.8, though at this
time the feature is in a state of decay. These periods of localised velocity fluctuations lasted for 5 to 10 seconds. Without more detailed measurement of flow characteristics, it is impossible to state the cause of this flow turbulence. It seems likely, however, that it is caused by secondary circulation along this concave outer bend, rather than from the shedding of turbulent eddies from the lee of channel bedforms or obstacles to flow. Detailed discussion of the causative mechanisms of boil production within Squamish estuary is given by Babakaiff (1993).

The most intense periods of undercutting and collapse took place during periods of wave activity in Squamish flow. The greatest wave activity in Squamish estuary (with the exception of storm events) occurs when up valley sea-breeze winds combine with an incoming tide. These conditions produce waves up to 2 m in height, which may travel up to 2500 m upstream from the river mouth. During the early stages of tidal rise (within the first hour) the sand unit is exposed and becomes intensely eroded by wave activity at the base of the bank. Within minutes of the river level exceeding the upper limit of sand, however, almost all erosion ceases. Bhowmik (1982) similarly noted the importance of wind (and boat) generated waves on bank erosion.

Over the period of study from late June to late August, the bank retreated up to a maximum of 4.9 m at one location. A number of erosion pins were washed out, and the largest bank retreat recorded from one erosion pin in one day was over 2 m. Re-stabilisation of the river bank took place within a number of weeks, and was essentially achieved when the toe of the bank became protected by collapsed bank material. At this stage the bank was in the state of impeded removal (Thorne, 1982). Figure 5.10 shows the appearance of the re-stabilised bank taken on July 14, over two weeks after erosion began. At the time this photograph was taken, the coarse sand unit at the base of the
Figure 5.10. The eroding section of the west delta taken from the same location as Figure 5.8, seen here less than 3 weeks after the erosion began (visibly). The toe of the bank is again protected to a large degree by slump block material.
bank was covered and so unable to be eroded by flow. While erosion did continue at this site for a number of weeks, it was limited to a number of small sites along the bank, and never equaled the rates of erosion recorded in the first week of observation.

Discussion

It is clear from the above discussion that certain reaches of Squamish estuarine river are laterally unstable, and are eroding floodplain and delta deposits at a fairly rapid, though discontinuous rate. The most important conclusion from these analyses is that the channel meander closest to the river training dyke is migrating eastward. The rapid rates of bank erosion at this location (recorded from photographs and erosion pins) pose a serious threat to the dyke structure. One can see from examination of Figure 5.4 that the river is very close to the dyke structure and central channel at this location. Without remedial engineering, Squamish River will switch distributary channels and once again flow through the central channel. If this channel bank section continues to erode at a rate similar to that determined in this study, dyke stability will soon be threatened. Although the dyke structure is itself a stable feature, it will be eroded through the removal of large blocks underlying tidal marsh deposits. For engineering attempts to be successful therefore, the near-bank zone along the concave meander bend needs to be protected at the low-water mark and the subtidal bank region. This may require that any protective structure extend between 5 and 10 metres away from the channel bank. Given the difficulties and expense of such engineering measures, and the likelihood that they will still fail to entirely stabilise this bank section, additional or alternative protective measures likely will be necessary. The alternative measure which is considered most appropriate here is artificial channel widening, which can be achieved by dredging the large point bar which is building along the inner bank of this channel meander.
If the river is successfully prevented from shifting eastward then it is clear that the west delta will continue to decrease in size as the lower 1700 m reach of the river migrates westward. Substantial loss of land has already occurred over the period of study, though there is no indication that rates of erosion have increased since dyke construction. This may have important implications for Squamish delta landuse management.

From the analysis of erosion within zone 4, it is clear that rates of erosion have increased dramatically in recent times. It is important to assess the possible effects that the river training dyke may have had on Squamish flow along its lower reaches, and whether these have exacerbated the natural process of erosion within this identified zone. Dyke construction has modified Squamish flow by effectively reducing channel width during flood conditions. Prior to dyke construction, floodwaters inundated all of the central delta, and water was able to flow through the cutoff and central channel of Squamish delta. With dyke construction, the central channel, east delta, and most of the central delta have become isolated from flow, resulting in a reduction of floodwater storage capacity. Although it is clear that dyke construction has modified Squamish River flow, it is impossible to determine from existing data, the exact nature of the rivers response to the new flow regime. One may postulate, however, that in order to maintain continuity, the decrease in flow width during periods of flood must be accommodated by an increase in flow depth and velocity. These increased flow depths and velocities have formed an anomalous stratigraphic sequence along lower reaches of Squamish River, as discussed in chapter three. These changes also likely are responsible for increased basal scour along the near-bank region, resulting in a more rapid rate of meander migration (Wiebe, 1976; Zrymiak and Durette, 1979; Government of Canada, 1981).
There are problems in any study of this kind, which attempts to determine the effects of some artificial flow modification from a limited period of record. These primarily arise from difficulties in quantitative distinction between allogenic and autogenic change (Lewin, 1977). Despite this, the order of magnitude of change in rates of erosion recorded around 1972 indicates that the loss of floodplain material is not through natural processes alone. This observation, in combination with the observations and recordings of accelerated and rapid erosion immediately after dyke construction by Bell (1975), Wiebe (1976), Zrymiak and Durette (1979), and Levings (1980), lends strong weight to the argument that dyke construction has exacerbated erosion, at least within this particular reach (zone 4) of Squamish estuary. It still remains unclear why this particular zone has experienced increased erosion yet other reaches have not.

Despite these comments, it is important to attempt some form of distinction between allogenic and autogenic change within this environment, even if this is necessarily limited to a qualitative assessment of the effects of discharge fluctuations. For this reason, maximum daily discharge data from 1957 to 1990 have been examined, and these are shown in Figure 5.11. These data are considered of greater value than mean annual or mean monthly discharge data because the monthly or yearly averaging process will mask discharge extremes. Figure 5.11 indicates that the largest daily discharge on record occurred in 1957, and the second largest (calculated as a 35 year recurrence flood) occurred in 1984. Unfortunately, it is almost impossible to detect any relationship between these discharge data and rates of erosion presented in Table 5.1. Despite this, it is evident that the post-dyke period has experienced a greater number of large floods than the pre-dyke period. For example, between 1957 and 1972 there was one recorded maximum daily discharge in excess of 2,000 cumecs, whereas three such discharge magnitudes have been recorded over the 1972 to 1990 period.
Figure 5.1. Maximum daily discharge data for the period 1957 to 1990. Lines indicate the periods over which erosion was determined from aerial photographs. These lines do not exactly match the periods of study identified in Table 5.1 because aerial photographs generally are taken in summer months, while maximum daily discharges mostly occur during autumn and winter months.
From this discussion it appears likely that river training dyke construction has served to increase rates of bank erosion along certain estuarine reaches through the decrease in channel and effective floodplain widths. This erosion has perhaps been exacerbated by an apparent increase in the frequency of the highest magnitude floods over the 33 year period of discharge record.

**Preservation Potential of Squamish Estuarine Deposits**

In Chapter 3, I conclude that within this high energy fjord setting, the facies which most clearly identify these deposits as tidal in origin are those of Facies B and C. We must consider, however, the relative importance of these deposits to the sedimentologist or geologist who wishes to determine the origin of sedimentary deposits. For a deposit to be of any use as an environmental indicator, it must first be preserved in some recognisable form. For this reason it is important to assess the preservation potential of marsh rhythmtes at this location.

Nio (1984) suggests that tidal sediments have a very low preservation rate, as they are usually destroyed by waves and storms. Similar conclusions may be made regarding tidal sediments in Squamish estuary, although the process of erosion differs from those mentioned by Nio (1984). The combined effects of a high-energy glacially-fed river migrating within a steep, laterally confined valley mean that estuarine deposits have a very low preservation potential. It is clear from the analysis of a limited record of erosion in Squamish estuary that Squamish River and delta distributary channels are highly unstable, and are constantly re-working sediments. It seems unlikely that any of the delta topset beds and floodplain deposits within the lower estuary exceed 400 years in age.
Thomas et al. (1987) state that the preservation potential of intertidal deposits is greatly increased if the sediment is cohesive, and particularly if these sediments were deposited as floccules, aggregates, or as faecal pellets. While marsh sediments possess all of these qualities, observations regarding the nature of bank erosion show that this is not important in Squamish estuary. The internal characteristics of the marsh rhythmite facies has little effect on its erosion because these deposits become eroded by removal of underlying uncohesive sediments. These uncohesive sediments underlie marsh deposits across the whole delta.

From the analyses of short-term and longer-term erosion rates, one must conclude that marsh rhythmite deposits will not be preserved (intact) in stratigraphic sequence in this fjord environment. From this, it also follows that these deposits will not become preserved in ancient and recent bay-head delta sequences such as those described by Bartsch-Winkler et al. (1983), and Postma and Cruickshank (1988). Despite this statement, it does not necessarily follow that marsh sediments will be completely eroded, or will fail to be preserved in some form. It is proposed that marsh rhythmite deposits become incorporated into the fjord deltaic sequence as cohesive blocks which are produced during bank erosion. These blocks (shown in Figure 5.7) are of considerable size (occasionally over 4 m in length, 2 m wide, and 1.5 m thick), and retain their internal form because of their cohesive nature, and because of the binding effect of the organic matting. Barwis (1978) states that slump blocks in South Carolina marshes become incorporated into the lower bank deposits, where they often become preserved in ancient sequences (e.g. Horne and Ferm, 1976). Howard et al. (1973) report that blocks of these cohesive materials up to 1 m in diameter may be found along the channel thalweg.
Once these blocks become eroded from the edge of the marsh, they collapse into Squamish River and rest on the channel bed. These become incorporated into the delta foreset beds if buried by an aggrading channel, or by migration and point bar growth. It is important to note, however, that this erosional process and production of large intact slump blocks is not unique to tidal environments. As such, it is important that these blocks can be determined as tidal in origin. If these deposits are to be of any value in environmental reconstruction, it must be shown that the blocks are able to retain both their form and internal character prior to burial, and upon compaction and diagenesis. In the case of these tidal marsh deposits, it is important that the organic bedding be seen in rock form, and that it is possible to identify the type or size of organic matter. Thomas et al. (1987) report that the organic criterion is of limited value to environmental reconstruction because of the lack of preservation of organic forms into rock. However, the processes of compaction and diagenesis are not fully understood, or constant from one site to another (Bouma, 1963). Although it is true that preservation will not always occur, a number of studies indicate that organic material is often well preserved. Zaitlin and Schultz (1984) reported a Lower Cretaceous coal facies up to 5 m thick which contains pyritised rootlets, interbedded with finely laminated carbonaceous shales. This facies is interpreted as salt marsh or peat swamp deposits. A very similar description of ancient deltaic coal deposits was given by Selley (1985). Okazaki and Masuda (1989) describe two Late Pleistocene delta sequences, and base their interpretations (in part) on differences in organic matter. The authors describe two deposits as "peaty layers with rootlets" and "muddy layers with roots and plant debris", and use this to distinguish between marine and fluvially dominated environments. If slump blocks remain relatively undisturbed these deposits may be identified as tidal in origin from an examination of bedding form. As has been shown in chapter four, the nature and pattern
of sedimentation within tidal environments appears to differ from that within fluvial environments, and this may be detectable in bedding form.

There are a few recorded exposures of marsh rhythmite blocks preserved within the sandy fill of ancient estuarine channels. Smith (1989) described blocks of material up to 1 m thick and 5 m long within the ECS of point bar deposits. An example of one of these blocks enclosed within sandy inclined strata, is shown in Figure 5.12. Smith based his interpretation of estuarine channel on these blocks of material (again in part), which were reported as having eroded from the tidal marsh and slid part way down the point bar surface.

![Figure 5.12. A block of marsh sediments on the sandy point bar surface. From Smith (1989)](image-url)
West Delta Accretion

The purpose of this section is to investigate the rates and spatial variability of west delta accretion. These data will be used to determine the changing rates of accretion with time (marsh maturity), the dominant source of sediment, and any seasonality of accretion. After this, the stratigraphic implications of these results will be discussed.

Spatial and Seasonal Variability of Accretion

The spatial variability of marsh accretion may be shown in two ways: by recording short-term (yearly) accretion, and by examining differences in marsh surface elevations. This latter approach gives information regarding longer-term accretion rates, and an indication of the relative age of the marsh. Both approaches are used here, the results of which are presented below.

Rates of marsh accretion were determined from a number of deposition stakes, the location of which can be seen in Figure 5.13. For a number of reasons, however, the short-term rates of marsh accretion provide limited data regarding temporal variability of accretion rates. A large number of deposition stakes were disturbed or removed over the winter period (as will be discussed later in this chapter), and a combination of time constraints and poor field conditions prevented the collection of some data. Despite these problems, accretion was measured at most stakes in August 1992, and limited data were collected in June and November 1993.

Figures 5.14, 5.15, and 5.16 show the amount of accretion over periods of 3, 13 and 18 months respectively. Although the data for June and November 1993 are incomplete, a
Figure 5.13. Showing the location of deposition stakes on the west delta. Stakes are in 12 transects, which are 100 m apart. XS-2 shows the location of a detailed survey of the marsh surface in 1977.
Figure 5.14. Showing depths of accretion on the west delta after 3 months of measurement. Data collected on August 16 1992. Accretion is in cm.
Figure 5.15. Showing depths of accretion on the west delta after 13 months of measurement. Data collected on June 20, 1993. Accretion is in cm.
Figure 5.16. Showing depths of accretion on the west delta after 18 months of measurement. Data collected on November 26 1993. Accretion is in cm.
number of observations may be made regarding accretion patterns. A number of trends may be noted within each of the 3 data sets; these trends are discussed below.

August 1992 (3 months after placement): 1) Along 10 of the 12 transects, recorded accretion is greater toward the centre of the marsh than at the marsh edge (at the river bank). 2) In three cases (stakes D2, A7, and Ab11) stakes record erosion, although in two of these cases erosion is minimal. 3) In one specific location, stakes record extremely high rates of accretion, but these seemingly anomalous values were very localised. Stakes A8, A9, and Ab9 measure accretion up to 3.7 cm, yet surrounding stakes measure little to no accretion.

June 1993 (13 months after placement): 1) From the limited data for this period it appears that accretion rates are highest at the river bank (along 3 of the 4 transects), and generally decrease toward the centre of the marsh. This is a reversal of the trend noted after 3 months of accretion. 2) Over the 10 month period from August to June, the delta has accreted rapidly, and at a much faster rate than initial results (in August 1992) would have predicted.

November 93 (18 months after placement): 1) Along 6 of the 8 transects, accretion rates decrease with distance away from the river bank. 2) As before, extremely high accretion rates are recorded at a number of locations along the river bank (stakes A3 and A4), and these rates decline rapidly with increasing distance inland. 3) Stake A7 recorded 0.7 cm of erosion while surrounding stakes recorded considerable accretion. This pattern was also noted at this location in August 1992, although erosion has increased since then.
From the observations above, one can say that west delta accretion is highly variable, both spatially and temporally. Data from August 1992 and June 1993 indicate that accretion is highly seasonal. During summer months, marsh accretion generally is minimal, and tends to be greatest away from the channel bank. Over the winter months of 1992 and 1993, accretion increases dramatically, and the location of maximum accretion shifts toward Squamish River. These results indicate that Squamish west delta accretes during the regular inundation by tides over the summer, and during the larger winter flood events, but that winter deposition provides the dominant supply of sediment required for the vertical accretion of the delta. In addition to this, the spatial variability of accretion shows that different depositional processes operate within summer and winter months. Over summer months, accretion takes place during combined high discharge and high tide periods, when suspended sediment concentrations are greatest on average. During winter months, sedimentation seems to be dominated by riverine processes over marine because winter accretion data show a shift in the location of maximum accretion toward the eastern marsh edge. This indicates that the river is the dominant source of sediment, which is deposited from suspension as floodwaters pass onto the smooth, largely unvegetated delta. The greater importance of winter over summer accretion becomes more apparent with increasing proximity to the river, where accretion rates generally are greatest over the 18 month period of study.

There is an obvious need for some caution with regard to these accretion data, because of the inability to determine the true representativeness of the necessarily short observation period. It is difficult to assess the importance of high magnitude events (of either fluvial or marine origin) to tidal marsh sedimentation. However, the short-term accretion data are in good accordance with the average annual sedimentation rate of post-dyke (Facies D) deposits. If this short period of measurement reflects the true long-
term post-dyke pattern of west delta accretion, some important points arise regarding the seasonality of accretion, and the changing nature of deposition from summer to winter seasons. The depositional process in winter months is different in at least one important respect - as most vegetation lies dead on the delta surface, it cannot promote the deposition of fines. This process is often said to be responsible for the greater proportion of marsh accretion. In Squamish delta, however, accretion data indicate that the majority of marsh accretion occurs when the delta is essentially unvegetated. In addition to this lack of vegetation, winter flood flow velocities are greater than those associated with the freshet. Given these facts, it seems probable that depositional product will vary seasonally. Without the presence of vegetation and marine organisms, which both promote the deposition of fines (Hubbard and Stebbings, 1968), winter deposits should be coarser than those deposited over summer months. Despite this likelihood, examination of the internal character and variability of marsh rhythmite beds of Facies C and D (chapter three) fails to record distinct seasonally induced differences. Nevertheless, the rhythmite deposits of Facies D do display distinct and abrupt changes in sediment character, though these clearly do not simply record seasonal changes.

These measurements of contemporary marsh accretion cannot be discussed in relation to marsh rhythmite development prior to 1972. As has already been discussed in chapter three, flow constriction has resulted in an increased riverine influence on sedimentation along lower reaches of Squamish estuary. This has increased both the size and amount of sediment deposited on the west delta. As such, accretion data may only be discussed in relation to the deposits of Facies D (treated in this chapter). Prior to dyke construction, the majority of Facies C rhythmite bed accretion is believed to have occurred over summer months.
The above discussion of spatial and temporal variability of accretion has implications for the sampling of delta surficial sediments. Great care should be taken to ensure that sediment sampling is both meaningful and provides accurate information required to answer the research question. Before sampling is performed, it is essential to be aware of any seasonality of deposition, and more importantly, aware of the average depths of deposition at each sampling location. For example, Bell (1975) analysed surficial sediment samples taken from a north-south transect along Squamish west delta. Samples were collected to a depth of 2 cm at each site, and were analysed for sand, silt, and clay content. There are a number of problems with this sampling technique. Firstly, the results of size-analysis may vary greatly depending on the time of data collection. Samples collected in August are likely to contain greater amounts of fines than samples collected from the same site in March. Secondly, it is important to know the rate of yearly accretion before sampling, to ensure that the sampling interval does not exceed the depth of one year's deposition. Stratigraphic analysis of marsh rhythmites (chapter three) indicates that at certain stages of marsh development, a depth of 2 cm may represent up to 15 rhythmite marsh beds (15 years of deposition). The importance of this fact will depend on the aim of sampling, but if the aim is to produce a sediment-size distribution which is representative of the marsh unit, sampling must include the whole depth of marsh deposits.

**Sedimentologic Implications of Accretion Data**

The contemporary short-term rates of west delta accretion may be used to test conclusions drawn from stratigraphic analysis of the salt marsh facies. Figure 5.13 shows the location of a transect (XS-2) surveyed in detail by the Water Investigations Branch of the Ministry of the Environment (Pronk, 1977). This transect closely follows
transect 4 in this study, so an examination of these data will show average accretion rates over the 15 year period from 1977 to 1992. This value may then be compared to the known accretion rates over the 1992-93 year.

Transect XS-2 follows almost exactly the same line as transect 4, and as the original survey (Pronk, 1977) was quite detailed, it was possible to identify points which were close to the location of the deposition stakes surveyed in 1992. Stakes A4, Ab4, and B4 lie within 3.5, 17, and 12 m of survey points along transect XS-2 respectively. Table 5.2 shows the average yearly accretion over the years 1977-92, and a comparison with accretion measured over one year.

<table>
<thead>
<tr>
<th>STAKE</th>
<th>marsh height in cm, 1977</th>
<th>marsh height in cm, 1992</th>
<th>annual average accretion, cm</th>
<th>accretion from May 92 to June 93, cm</th>
</tr>
</thead>
<tbody>
<tr>
<td>A4</td>
<td>41</td>
<td>90</td>
<td>3.27</td>
<td>6.0</td>
</tr>
<tr>
<td>Ab4</td>
<td>46</td>
<td>80</td>
<td>2.27</td>
<td>3.1</td>
</tr>
<tr>
<td>B4</td>
<td>46</td>
<td>59</td>
<td>0.87</td>
<td>1.1</td>
</tr>
</tbody>
</table>

Table 5.2. Comparison of short-term (1 year) and longer-term (15 year) accretion rates along a transect across the west delta. Average annual accretion is determined from elevation differences between 1977 and 1992, while accretion over the 1992-93 period was determined from accretion stakes. Marsh heights are in cm a.m.s.l.

The table above shows that 2 of the 3 measured accretion depths quite closely agree with the 15-year average accretion depths, while the third value is almost twice the average. Error may be induced by the lack of fit of one years accretion to a general trend, indicating high annual variability of accretion. Error may also arise from the local
variability of accretion (bearing in mind that survey points are several metres apart), and due to the compaction of sediments upon burial.

It is interesting to note that the poorest correlation between short and longer-term accretion is for the stake closest to the river bank (10 m in 1992), and that the best correlation is for the stake furthest from the river (110 m in 1992). This agrees with observations made with regard to the spatial variability of accretion rates measured in 1992 and 93. Figures 5.17a and b show that accretion rates generally are highest near the river bank, and that the variability of accretion decreases away from the channel.

![Graphs showing variability of west delta accretion with distance from Squamish River](image)

Figures 5.17a and b. Variability of west delta accretion with distance from Squamish River.
There is another possible explanation for the lack of fit of the 1992-93 accretion data to the averaged values determined for accretion stakes A4 and Ab4. The 1977 survey along transect XS-2 shows that around 86 m of bank retreat (along this transect line) have occurred between 1977 and 1992. This means that over the past 15 years, the location of accretion stake A4 has been getting closer to the main source of sediment supply to this delta. Because of this, the rates of accretion at this site will have been increasing since 1977. This may explain the poor fit of the short and longer-term values of accretion. This process is also likely to have led to some change in rates of accretion at all stakes, although this change is expected to be less pronounced with increasing distance away from Squamish River.

**Model of Squamish Delta Growth**

The purpose of this section is to draw comparisons between the lateral and vertical growth of Squamish delta, and the stages of salt marsh development outlined by Redfield (1972). Redfield undertook a detailed study of the development and characteristics of a New England salt marsh. This landmark work has come to be accepted as the blueprint model for salt marsh development for those systems which experience rising sea-levels. The following sections describe the stages of salt marsh development as observed by Redfield (1972), and a comparison with observations from Squamish delta. Discussion then focuses on the sedimentologic implications of different stages of marsh development, with specific reference to Squamish delta stratigraphic sequences.
Redfield's (1972) Model of Salt Marsh Development

Redfield stated that five distinct stages of salt marsh development may be distinguished, which are related to the elevation of the marsh. These stages will fully develop where the salt marsh is building onto a flat or nearly flat surface.

The first stage is that of colonization, which is the initial spread of marsh vegetation onto previously unvegetated sand flats and bars. These intertidal sands will be preserved as the "basement" of inorganic deposits underlying peat. This colonization occurs by the lateral spreading of the edge of vegetation by the spreading of rhizomes, and by seeding. Seeding produces clumps of vegetation beyond the extent of full marsh cover, which eventually spread and merge to form complete cover within a number of years. This stage is followed by the juvenile marsh stage, which is that of an even vegetation cover which is dominated by a single species. These young marshes usually do not develop distinctive drainage channels. This early marsh stage progresses into panne marsh, which is prompted by local differences in marsh accretion. These differences create localised depressions which are unable to drain fully at low tide. This poor drainage creates small pools of saline water which soon kill off patches of vegetation. Over time, these pannes fill with soft fine material containing little organic matter. This stage of marsh next develops in to slough marsh, where the vegetated marsh accretes at a faster rate than the pannes, creating raised vegetated surfaces surrounding lower ponds. The final stage of development is the transition to high marsh. During this stage, pannes become drier, allowing the invasion of marsh vegetation. With increased marsh maturity, all pannes may disappear, leaving a relatively flat fully-vegetated delta surface.
The above model is applicable to marshes developing on intertidal sands with low relief. Where marshes develop on sloping surfaces, however, sediments are better drained, and fewer pools of water remain at low tide. This results in a marsh surface which is less frequently broken by panne formation.

Redfield states that the vegetated marsh will develop a characteristic stratigraphy related to the seasonality of accretion and vegetation growth. This stratigraphy is revealed in eroding marsh banks, and is described as: "...layers of fibrous material alternating with layers of nearly pure sand and silt." (Redfield, 1972 p. 223). The layers of fibrous material are thought by Redfield to originate during summer months from the algal mat which grows on the surface of the delta, and from the concentration of roots at this level. The dominantly inorganic layers are thought to be deposited during winter storm events, when marsh vegetation has died-off. This interpretation differs markedly from that made within this study. Redfield suggests that each pair of organic and inorganic layers represents one year's vertical growth of the delta, an interpretation which agrees with the conclusions of this study.

Applicability of Redfield's Model to Squamish Delta

Comparison of the development of Squamish delta with the model described by Redfield (1972) has been undertaken by an examination of contemporary marsh environments, and by careful study of aerial photographs dating back to 1957. In general, the model does appear to fit the example of Squamish marsh growth. The specific stages of development of Squamish marshes (with reference to Redfield's observations) are discussed below.
The colonization of intertidal sand flats is clearly occurring at the southern tip of the west delta. At present, this colonization is driven by the spreading rhizome mat of established marsh vegetation, though there is photographic evidence (provided by Bell, 1975) that spreading by seeding has previously occurred.

After the initial colonisation of sand flats, the marsh passes through the juvenile stage to the panne marsh. Across Squamish delta, however, the development of the panne marsh is not complete, and few pannes are observed. There is photographic evidence of pannes on west and central deltas of Squamish estuary (Figure 5.18), though these clearly are isolated and uncommon features. Aerial photographs also reveal the transition from intertidal marsh with pannes to high marsh. By 1977, the pannes observed in Figure 5.18 (10 years previously) had become vegetated.

Figure 5.18. Pannes on the surface of the west delta in 1957
The small number of pannes observed within Squamish marshes is to be expected, and was predicted by Redfield (1972). The marshes in this environment are spreading onto intertidal sandflats and bars which are far from flat. The surface relief of this sandy basement is reflected in the surface relief of the salt marsh (at least in the early stages of development), which is easily drained at low tide. For this reason, pannes are not dominant features on Squamish marshes, which never develop to the slough marsh phase.

From this discussion, it can be seen that growth of marshes in Squamish estuary quite closely agrees with the development described by Redfield (1972). The Redfield model is limited, however, in that it does not describe the final stage of marsh development - that from high marsh to floodplain. This is a particularly important transition within Squamish estuary, where accretion rates exceed the rate of compaction, and short-term (hundreds of years) sea-level rise is essentially zero. The change from delta to floodplain is not an immediate process, but initially a gradual one which will produce a transitional sequence. The nature of this stratigraphic sequence has been discussed in chapter three as the transition from Facies C to E. This discussion is only applicable to those marshes which become increasingly influenced by riverine depositional processes over time, such as those which develop within prograding delta systems.

Sedimentologic Implications of Marsh Development

While Redfield (1972) did describe different stages of marsh development, he did not fully discuss the stratigraphic implications of these different stages. This section is an attempt to describe the characteristic sedimentologic attributes of each of the stages
operating in Squamish marshes. This will be related to the results of stratigraphic analyses of delta sediments presented in chapter three.

During the early colonisation of sand flats, vegetation cover is too sparse to have any effect upon flow processes, and as such, on sedimentation processes. After colonisation, however, the growth of Spartina and other marsh vegetation is rapid, growing to maturity within 5 years (Redfield, 1972). As this vegetation becomes more established and taller, it is able to modify flow, largely through the baffling effect of stems. This flow modification promotes the deposition of fines, which is aided by the removal of fines from the water body as material adheres to the stems of plants. Some of this material will inevitably reach the delta surface during low tide. The processes described can clearly be seen in stratigraphic sequence. The colonisation of intertidal sands produces a gradual transition from the basement sands to the overlying marsh rhythmites (Facies B to C). Toward the bottom of this transition, no change in grain-size is observed, but traces of organics in very fine layers can be seen. Further up-section (increasing maturity of the marsh), the amount of organics increases within clear parallel bands, and grain-size decreases. The gradual decrease in grain-size continues with decreasing depth in section, until the layers become the distinctive silty marsh rhythmites of Facies C. This whole transition takes place over a vertical distance of a few cm.

During the colonisation process, the formation of small islands or patches of vegetation through seeding is an important one, as it can determine the internal characteristics of marsh rhythmite bedding. Redfield (1972) noted that the islands of vegetation appeared on mounds of material elevated above the intertidal sandflats, produced by the promotion of deposition by the vegetation itself. When these vegetated patches become
fully incorporated into the marsh cover, the marsh bedding will retain an uneven form characterised by gentle undulations. As fine material is continually draped over this undulating surface, the form of that surface will be retained (to some degree) in overlying beds. This is partly responsible for the wavy, nodular or 'crinkly' nature of beds reported by van Straaten (1954) and Martini (1991).

As is noted above, the panne marsh is not well developed in Squamish estuary, but a number of isolated pannes are visible on all delta surfaces. As pannes infill and become vegetated, a unique sedimentary deposit will be preserved in the marsh sediments. Relict panne deposits will be preserved as isolated lenses of material which contain little to no organics, and which likely will be finer than surrounding sediments. Panne deposits will display very fine bedding (much thinner than surrounding marsh rhythmites) which commonly is undisturbed by bioturbators due to the anoxic conditions in the pools (Yeo and Risk, 1981). Panne deposits will be capped by marsh rhythmite beds as vegetation gradually colonises the sediment surface.

While such depositional features can be expected in isolated locations throughout all three deltas in Squamish estuary, none were found in section. This should not be unexpected, for even though marsh rhythmite deposits are exposed along a 3 200 m reach of the lower Squamish River, bank sections only represent a two-dimensional view of a single line through delta sediments.

Specific Deltaic Depositional Process and Product: Rafted Deposits

During the collection of accretion data on the west delta in the winter of 1993, a number of large stones were found on the surface of the west delta. These large deposits were
not observed in summer months, as they were obscured by the tall, dense Carex lyngbyei stands on the delta surface. During winter months, however, when almost all delta growth has died and collapsed, these seemingly anomalous sedimentary deposits are clearly visible. The following two sections describe these deposits in detail, and discuss the nature of their deposition and transportation onto the delta.

Description of Deposits

The stones were found on the surface, or partially buried within the sediments of the west delta. In all, over 50 stones were found (though this is by no means the total number of stones present on the delta), ranging in size from gravel to cobble. Table 5.3 shows the size of a representative number of stones found on the surface of Squamish west delta. All of the stones were either well-rounded or subrounded. In no instance were any striations found on the stones, which were all igneous (predominantly granodiorite).

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Table 5.3. Sizes of some of the stones found on the west delta in 1993 and 1994
The stones lay on the surface of the delta, as observed in Figure 5.19 giving the appearance that they had been placed or dropped there. This observation is supported by the lack of disturbance of the surrounding delta surface. In no case were the stones associated with any other material (large organic debris, coarse sand, or pebbles), and the delta surface was quite undisturbed; there were no roll or drag marks in the surrounding soft fine sediments.

The stones were not spread across the whole delta, but instead were concentrated toward the streamward edge of the lower (seaward) half of the delta. Over 80% of the stones were found within 10 m of the river bank, and with greater distance upstream, an increasing proportion of stones (over 90%) were found within 5 m of the river bank. Many stones appeared not to be randomly placed, but appeared to be clustered in discrete groups, as seen in Figure 5.20. On a number of occasions, many stones were found clustered together within a few square metres, with no other stones for 30 m in any direction. In the largest cluster, 10 stones ranging in size from gravel to cobble were found within 1 m². At five locations, stones were lying within one metre of the rebar deposition stakes placed to measure delta accretion. In all five cases, the stakes were severely bent down estuary (Figure 5.21).

Stones were found on the sandy intertidal zone at the southern limit of the west delta, to a point 600 m upstream. One stone was found 110 m in from the river bank (shortest distance), which was 210 m from the valley wall. It should be stated that time constraints prevented an extremely detailed search across the whole extent of the delta, so one may expect to find stones both further upstream and further inland than those reported above. Despite this, it is clear that these observations reflect the general pattern of stone location on the west delta.
Figure 5.19. One of many stones on the surface of Squamish west delta. The stone sits atop silt and fine sand and appears to have been dropped or placed there. The stone lies directly over the dead marsh grasses, indicating that it was deposited after the late autumn death and collapse of vegetation. Scale is 10 cm.
Figure 5.20. Stones on the surface of Squamish west delta are often found in discrete clusters as seen here. Scale is 10 cm.
Figure 5.21. One of many bent rebar accretion stakes on the surface of the west delta. In several cases (as above), a cobble is seen within 1 m of the stake. The stake in this figure points down valley. Scale is 10 cm.
Interpretation

There are a number of possible processes which could have deposited this large material on the delta. Each of these will be discussed in turn.

One possible explanation is that the stones originated on the steep-sided valley wall, and were transported to their present position by a rockfall. However, if this material were brought to rest on the delta by this process, the number of stones would increase toward the back of the delta (the western edge), and some form of debris cone or lobate form may be expected to develop. In addition to this, the stones do not display the characteristic angular and fractured appearance of rockfall debris, and there are no roll marks visible on the delta surface. For these reasons the stones on the west delta cannot have been transported from the valley-side during a rockfall event.

A more plausible explanation for the origin of these stones is that they were transported onto the delta surface during a large flood. There are several pieces of evidence which seem to substantiate this theory. Firstly, all stones display rounded or well-rounded forms characteristic of fluvial origin. Secondly, the location and concentration of stones toward the river bank clearly indicates that the river was the source of this coarse sediment. Despite these points, however, there are several observations which are extremely difficult to explain with this theory. For example, if the stones were deposited during a large flood, why are they not present by the bank along the entire reach of the delta? One would expect to find a number of stones along the upstream bank of the west delta, as the channel thalweg crosses over and impinges on the delta at this location.
Perhaps the most convincing evidence against the flood mechanism is the size of the material transported. While it is certainly possible for Squamish River to transport material of such a size, it is not possible that the velocities required to keep the stones in suspension could have been maintained over the delta surface. Some of the material located within a few metres of the river bank could have been deposited there from the fast within-channel flows. However, it is not possible that flow velocities over the delta were great enough to transport a cobble 110 m from the river. Flow velocities of this magnitude would have been sufficient to induce considerable erosion of delta sediments, yet such erosion was neither observed nor recorded. Even if the delta surface sediments were cohesive and compact enough to prevent their erosion, such flows would have transported very coarse sand, pebbles and gravel onto the delta, along with the larger material. This sand would have been deposited as crevasse splays or extensive sand sheets. Not only were such large sand deposits not observed, there was a clear gap in sediment-size distributions of delta sediments. Other than the recorded gravels and cobbles, the coarsest material visible either on the delta surface or in section is medium to coarse-grained sand.

For the reasons stated above, it is felt that direct fluvial action could not be responsible for the deposition of the greater number of stones on the delta, though it is possible that some of the smaller stones within one or two metres of the bank could have been deposited in this fashion. Despite this statement, it seems clear from the character of the stones, and particularly from their location, that the deposition of this material was in some way controlled by fluvial action. This indicates that the river is the source of the material but it is not strictly the transporting agent, or directly responsible for the deposition of these stones.
This raises the possibility that the stones were rafted into the delta environment by river ice. Once over the delta, coarse-grained sediment could have either rained-out, or the ice could have become stranded on the delta, releasing any sediment held within it upon melt. This latter process is an extremely common occurrence in more northerly fjord deltas (Gilbert, 1978, 1983). Many of the observations made in the field seem to substantiate this theory. This process would certainly explain how such large material came to rest so far inland, and would explain why the stones so closely resemble dropstones sitting atop much finer sediments.

Once again, however, it must be stated that despite these observed similarities between the proposed mechanism of deposition and the deposits, ice has to be ruled-out as the origin of the stones on the west delta. The simple reason for this is that Squamish climate does not generate low enough temperatures to produce ice flows. Temperatures certainly drop enough to prevent the measurement of stage (Water Survey of Canada), but this is merely due to the failure of stage recorders (freezing) or the local build-up of ice at the bank. It is not possible that this ice could reach the size necessary to support and effectively transport cobbles.

Having established which processes are not considered responsible for the deposition of stones on the west delta, it is now possible to describe the depositional process which is believed to be occurring.
Proposed Mechanism of Deposition

Transportation of Stones onto the Delta

It is proposed that the coarse material is transported onto the delta by large floating debris - trees and root wads. This material may be held within the soil mass in the root wad, or more commonly, is held in place by the roots themselves. The stones are not thought to be transported into the existing root system, but become incorporated as the tree roots grow into the soil. Therefore, the existence of stones in tree roots is dependent on tree maturity (root number and size), and more importantly, on the presence of stones within the soil in which the tree becomes established. Presumably, this process may also be species specific, depending on the characteristics of root development (depth and density of roots).

When these trees fall, become eroded or washed out, the roots pull up the material held within them. If the tree enters a river course then the finer material will become washed out quite quickly, but larger material may be more effectively trapped. This material then floats downstream within the root system of the transporting agent. Studies by Fritz and Harrison (1985) have shown that stones may remain trapped within roots even during very high energy flows. These authors observed stones held in the root systems of trees transported by highly turbid flows produced by the March 1982 eruption of Mount St. Helens (Figure 5.22). A number of these trees had been sheared by the force of the volcanic blast.

Trees (and any sediment they may be carrying) are transported downstream during floods. These trees are mobilised if bank erosion occurs, or as flow velocity and stage become great enough to shift trees caught on channel bars. This process may occur at
any time of the year, though larger winter flood events are thought to erode and transport the greatest number of trees. If flood discharges are great enough and are combined with high tide, some of these trees are moved onto and across the delta surface. While many of these float over the delta and out into Howe Sound, some become stranded on the delta by a falling tide or waning flood. Once here, they remain until floodwaters and/or tides create a body of water deep enough to allow the tree to float off the delta. Aerial photographic evidence and field observation indicates that the residence time of these trees varies from days to years. Indeed, there is stratigraphic evidence to show that some
of this very large organic material is never removed, but instead becomes incorporated into the sediments of tidal marsh Facies C.

An examination of the root systems of a number of trees and wads found along the lower estuarine reach of Squamish River in June and November 1993, and April 1994 revealed that large debris is often held within root systems. In almost all recorded cases the stones were held between two or three roots (Figures 5.23 and 5.24, and 5.25), and were not found within a matrix of soil. The sizes of some of the stones found held within the root systems of trees within Squamish delta are shown in Table 5.4.

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Table 5.4. Sizes of some of the stones found within root systems of trees on the west delta and along lower reaches of Squamish estuary in 1993 and 1994.

Stones were found within the root systems of large intact trees and in the stumps of trees which previously had been logged. Most of these trees had clearly been dead for many years, and showed signs of having been transported in Squamish flow over considerable distances. Considerable amounts of organic debris were found along the lower 1700 m of Squamish River on the eastern side of the channel, and on the west delta.
Figure 5.23. Large stone held tightly by the roots of a tree located on the eastern edge of Squamish River, around 1500 m upstream from the river mouth. Scale is 10 cm.
Figure 5.24. Tree and root system on the eastern side of Squamish River opposite the west delta. Seven stones were found in this root system, which faces upstream. Book is 20 cm.
Figure 5.25. Close-up of the root system shown in Figure 5.24. Four stones are seen here, although one of these is not thought to have been transported with the tree. The stone closest to the camera is too clean and is sitting loosely on top of the roots. The other stones are covered with green algae, indicating that they have been within the roots for a considerable length of time. Book is 20 cm.
Figures 5.26a and b give an indication of the size of root systems found in this environment. Figure 5.26a shows a densely-rooted wad found to the east of Squamish River, on the surface of a large bar building opposite the upper end of the west delta. This wad is approximately 3 m in diameter, and contains one stone. In the background, a number of trees can be seen on the surface of the west delta. Figure 5.26b shows the amount of debris at the southern limit of the west delta. This material is resting on the small sandy intertidal zone and on sand bars to the south of the delta. The root system of the tree closest to the camera is approximately 4 m in diameter. No stones were found in the roots of the trees shown here, though the sandy intertidal zone was scattered with half buried stones (Figure 5.27). In addition to this stone, Figure 5.27 shows dead Carex lyngbyei stems which may become incorporated into the sediment, and cracks in the surface. These cracks are shrinkage cracks produced by the constant wetting and drying of these intertidal sediments. There also is clear evidence of geese prints, and small circular marks which resemble rain imprints or gas-escape features, but which are almost certainly made by geese feeding.

The best example of a root system carrying stones was found on the edge of the west delta, 500 m upstream from the southern limit of the delta. The tree and root system were bleached, clean, smooth, and largely stripped of bark. This, in combination with the rounded and fairly smooth root stumps, gave the impression that this tree had been dead and in transport for a considerable length of time. Despite this, the roots were found to contain 22 stones of varying size ranging from gravel to cobble, some of which can be seen in Figure 5.28. This tree (photographed in early April 1994) lay parallel with, and 2 m from the river bank, and was not at this location at the end of November 1993. Around the base of this root system, 5 stones were found on the delta surface.
Figure 5.26. a) Stone held in root wad on the eastern edge of Squamish River. Scale by stone is 10 cm. b) Looking south into Howe Sound fjord, showing the number of trees stranded at the seaward tip of the west delta.
Figure 5.27. Close-up of stone partially buried in the intertidal sands at the southern tip of the west delta. Numerous other markings are present in the sands, and these are described in the text. Scale is 10 cm.
Figure 5.28. Close-up of the root system of a tree stranded on the edge of the west delta. This root system points downstream. In total, 22 stones were held by these roots. Scale is 10 cm.
Removal of Stones from Root Systems

It has already been established that once these trees and root wads become stranded on the delta surface, they may remain over a number of years. Over this period of time, a process or number of processes operate to remove the stones from the roots. It is not thought, however, that currents (either tidal or riverine) would generate the turbulence required to remove the stones. Instead, it is felt that the stones are loosened and removed by the action of waves.

As has already been discussed in chapter one, Howe Sound fjord and estuary is highly influenced by wind action which can act to increase the wave height of tide-generated surface instabilities. These waves may spill onto and over delta surfaces if the delta is inundated by high tide or flood. Figure 5.29, taken on 31 December 1980, shows waves on the surface of the central delta. In this particular example, however, the waves are travelling in a southwesterly direction (down estuary). The direction of wave travel most likely is determined by wind conditions (whether sea-breeze or land-breeze conditions prevail), which may vary throughout the course of one day. If the delta is partially inundated, these waves will strike the trees and their root systems with some force. On a number of occasions, deep (7 cm) scour marks and imprints were observed around the base of stranded root wads. These clearly indicate that the trees had been moved (by up to 1 m) a number of times since their original placement on the delta. While it cannot be proven (without direct observation) that these objects were moved by wave action, waves are the only action thought powerful enough to shift objects of such size. It should also be noted that scour marks indicated that the trees had moved upstream from their original locations. Presumably, this process was aided by the decrease in downward pressure exerted on the delta by the partial floating of the trees in standing water.
Figure 5.29. During this flood in 1980, two levees became exposed on west and central deltas. Stranded trees are seen on both these delta surfaces, and there are waves breaking over the surface of the central delta. These are travelling down estuary.
As it seems clear that the waves strike stranded trees with enough force to move them, the next question is whether the force of these waves is great enough to dislodge stones held within roots. Field testing of the force required to remove stones (all stones were tested), indicated that in many cases very little force was required. Some stones would obviously never be removed by natural forces exerted on the tree roots. In these examples the stones would only drop out if the tree remained in place long enough for the roots to rot to the point where they could no longer support the weight of the stone. An example of this is shown in Figure 5.23. This stone could only be removed by breaking away roots. In other cases, however, stones required extremely little force to remove - they just required that that force be applied in a specific direction.

**Distribution of Stones on the West Delta**

It has already been noted, the stones are dominantly found toward the lower portion of the west delta, and most of these occur alongside the river bank. Observations made during field visits, and close examination of aerial photographs reveal a close relationship between the location of stones and the occurrence of stranded trees on the delta. Aerial photographs show that trees with root systems are never recorded more than 700 m upstream from the seaward edge of the delta. Similarly, far fewer trees are observed in the centre of the delta than by the river bank. Some trees are seen toward the very back of the delta (by the valley wall), which must have been transported up the large tidal channel. This section of the delta was inaccessible at the time of data collection, so no check for stones could be made.

The above discussion is with specific reference to trees with intact root systems, or root wads. No mention has been made of tree stumps and logs, as these cannot transport and
deposit stones. These were observed over much of the delta in locations further upstream than the larger rooted-trees.

The large number of trees toward the river bank is obviously not by chance, nor is it simply that the large floating debris is 'deposited' rapidly after removal from the main flow. The trees commonly are found within a few metres of the bank because the large root systems (1 to 3 m or more in diameter) get caught on the levee identified in chapter three, then stranded as waters fall. Figure 5.29 shows a view of the west and central deltas during flood. Levees can be seen on both of these deltas, which have been built by flows in the western channel (now the only open channel). If water depths are great enough to allow the whole tree and root system to pass over the levee then the tree likely will float across the delta and out into Howe Sound.

This proposed mechanism of stone transport and deposition can be used to explain the bent rebar deposition stakes noted earlier. The stakes would very easily have been bent if struck by trees floating across the delta in a general downstream direction. It follows that the impact of this collision would jar the tree, and could dislodge one or more stones held in the roots. This would explain why stones were found within 1 m of two-thirds of the bent stakes, as seen in Figure 5.21. This also highlights the fact that human interference (the positioning of deposition stakes) has artificially induced the deposition of some of this 'float-in' or rafted material.

**Sedimentologic Importance of Rafted Clasts**

From the discussion above, it is clear that within this environment, the transport of coarse material by trees is a common occurrence, and that a large number of stones
become deposited on the delta. These rafted deposits are important sedimentologic features because their origin may not readily be apparent. The coarse material is brought into the deltaic environment by larger objects (trees) which are then removed, leaving little or no evidence of their presence. This lack of evidence leads to the danger of misinterpretation of the origin of rafted deposits (or clasts) seen in section. For example, in more northerly deltas where ice transport occurs, these stones may be misinterpreted as dropstones or ice-carried debris.

If these large deposits are to be interpreted or misinterpreted, it is clear that they must first be observed in section, so it is important to determine the preservation potential of these deposits. It is interesting to note that despite the fact that so many stones were observed on the delta surface, none were ever observed in section. This is considered to be pure chance rather than an indication that rafted clasts are a recent phenomena. The processes of trees transporting and depositing stones in that delta environment today are considered essentially the same as those occurring during the early stages of development of Squamish delta. This statement is substantiated by pre-dyke photographic evidence which shows a number of trees on all delta surfaces.

If these processes have always been in operation in this environment then rafted clasts should be present throughout the delta topset beds from intertidal sands to marsh rhythmites. Despite this statement, it is felt that the majority of rafted deposits have a short residence time (poor preservation potential), at least in west delta sediments. This is because the majority of stones appear to be deposited within a few metres of the channel bank. As has already been shown, the entire length of this bank has been eroding at a considerable rate over the last 33 years.
The preservation of rafted deposits is dependent on the burial and preservation of intertidal sediments. Within this high-energy environment where the sedimentary basin is laterally restricted, it is felt that these deposits have a poor preservation potential and a residence period not exceeding 400 years. This statement appears substantiated by contemporary rates of channel bank erosion, and the local history of channel avulsion and switching.

**Previous Findings and Observations**

Despite the low preservation potential of gravels and cobbles within the marsh sediments of this fjord environment, a number of similar deposits have been recorded elsewhere. Orton (1892) described a number of boulders (one up to 400 lbs) found within a thick coal seam in Ohio. In no case did any of these boulders disturb the internal structure of the coal. While most of these boulders were well-rounded and smooth, one was described as being of different composition, and "...angular, as if freshly broken from the parent mass." (Orton, 1892 p. 63). Orton could offer no explanation for the mode of transport or deposition of this stone, but felt that it was different in origin from the other, well-rounded boulders.

Twenhofel (1917) described a number of distinct clusters of boulders at the top of what is interpreted as a deltaic sequence in Kansas. In all, around 100 boulders were found, the largest being over 2 m in diameter. All boulders were noted on the surface of a unit of Pennsylvanian shale, although Twenhofel speculated that these boulders originally lay within the shale. Four possible methods of origin of these deposits are discussed by Twenhofel (1917), who dismisses the possibility of a fluvial source because the boulders lay within sediments which were obviously deposited in a low energy environment.
Flows in such an environment would have been incapable of transporting boulders of this size. Instead, Twenhofel (1917) believed the boulders to have been transported from ice which broke-off from glaciers and was transported downstream. Melting blocks of stranded ice in the intertidal environment deposited the boulders within what is now the shale deposits.

In 1932, Price attempted to determine the origin of what he termed erratics in coal beds, with reference to a number of previous recordings. All recorded erratics were well-rounded, and a few were reported to be slightly etched or grooved. There were no signs of site disturbance or destruction of bedding surrounding the stones. These stones, which represented a number of different rock types, were found throughout the coal seam. Price (1932) discussed the possible origin of these erratics with reference to transportation by ice, water, and trees. Without offering any definite answer to this problem, Price (1932) argued that the material was incorporated into the marsh by trees. Price argued, however, that these trees were deposited in a stream before marsh development, and that these trees eventually became overturned, elevating boulders (held within tree roots) into the marsh sediments. No explanation was given for the driving force for this overturning, or for the lack of bed disturbance, which would surely have occurred during such a process. Price did not believe that the material could have been rafted-in directly by trees because he felt that a stream large enough to carry trees would have eroded the peat bog. In addition to this, Price felt that tree-rafting could not account for the deposition of erratics throughout the coal seam.
Discussion of Previous Research

The early work discussed above raises a number of interesting points which are worth mentioning here. The first point relates to the shape of the stones found within a matrix of finer sediment (or coal). Orton (1892) believed that the angular boulder was significantly different from other examples, as it must have a different mode of origin than well-rounded examples. It is clear from this study that the shape of stones deposited by tree-rafting is determined largely by the type of material within which the tree grows. It is highly possible that tree-rafting could deposit fluvially rounded stones, angular rockfall material, or glacial till, so the shape of float-in clasts should not influence interpretation. The only information which shape may reveal is the distance that the tree has been transported in the river, or the length of time that the tree has been in the river system.

Similar warnings must be made with regard to the rock type of rafted deposits. Differing rock types need not imply different processes of deposition or formation (a point examined by Twenhofel (1917)), but may certainly be used to give an indication of source. In this way, Price (1932) determined that trees had transported boulders some 60 miles from their source. It should be stated, however, that the parent rock may not be the direct source of rafted material, as trees may grow in deposits which have previously been transported considerable distances by fluvial action, ice, or mass movement.

Finally, Price’s (1932) thoughts regarding the inability of trees to deposit boulders throughout a marsh or peat deposit is clearly incorrect. Within the specific Squamish example, however, the number of these deposits should decrease with increased height of the marsh surface, as trees are unable to move onto the delta. The ideal location for
this process of float-in deposition is the interdistributary bay environment, although it should not be restricted to these areas. The ideal location is one where floodwaters rise and fall on a fairly frequent basis, yet one in which energy conditions are low enough to permit the accumulation of fine sediments. If Squamish estuary can be regarded as being typical, the environment must also be open enough to allow the generation of wind waves during high water stages.
CHAPTER SIX
CONCLUSIONS

The initial aim of this case study was to examine the contemporary depositional and erosional processes operating within the high energy, fjord-head Squamish delta. More specifically, the different components of this study aimed to determine the degree of tidal control on sedimentation, the changing nature of that control along the tidal gradient, and to determine the nature and rate of bank erosion within the estuary. The following pages present a summary of the findings.

Depositional Model of Squamish Delta and Floodplain Formation

Floodplain and delta stratigraphic sequences were examined within bank sections along the estuarine reach of Squamish River. Seven distinct facies have been identified within these sequences (labeled Facies A to G), which have been given purely descriptive terms to avoid bias or confusion. Facies are labeled as follows: Facies A - cross-bedded sands and gravels; B - finely-laminated sands and silts; C - silty rhythmites; D - sandy rhythmites; E - parallel-laminated sands with peds; F - ripple-bedded sands; G - parallel-laminated silts and clays.

With the possible exception of the deposits along the lower 1 000 m of Squamish River, these deposits were laid down along shallowing channel margins and within aggrading channel sections, and later, within an interdistributary bay environment. Channel aggradation is believed to have been induced by gradual channel abandonment. Within this environment, channel lag sands and gravels (Facies A) were deposited to an elevation just above mean low tide level. Above this point, tidal modification of flow
prevents the transport of sandy bedforms. This is reflected in sequence as a distinct change from the coarse sands of Facies A to finer intertidal sands of Facies B. As this intertidal surface accretes, halophyte vegetation begins to colonise the bare sands. This vegetation rapidly spreads and becomes more dense, so that it acts to preferentially deposit fine-grained material. This stage represents the formation of tidal marsh deposits (Facies C). As this marsh surface accretes, both sediment and vegetation type change as the delta plain gradually changes to an alluvial plain (Facies E). This gradual transition is abruptly interrupted by riverine flood coarse overbank deposits of Facies F, which generally form the uppermost deposits in section. In places, thin deposits of finely laminated silts and clays (facies G) deposited by tidal overbank flows are present toward the top of estuarine sections. Deposits of Facies D are coarse-grained, crude rhythmite beds forming a levee along the lower 800 m reach of Squamish River. At those isolated locations where these deposits are found, they produce an anomalous coarse-grained stratigraphic sequence which is discussed later in this chapter.

There is some sedimentologic evidence to suggest that the lower 1 000 m of estuarine deposits are the direct response of deltaic progradation. The presence of shells and distinct tidal bedding suggest that the sands of Facies B formed within an intertidal sandflat environment along the front of the delta rather than within an aggrading channel reach.

**Tidal Controls on Sedimentation Within Squamish Delta**

All facies identified above are tidal deposits by strict definition, as they formed within an environment subject to tidal flux. There exists a distinction, however, between those deposits which are known to be tidal from their present location, and those deposits
which display sedimentologic evidence of their tidal character. Of the 7 facies identified in this study, only the intertidal sands (B) and tidal marsh facies (C) display distinct tidal characteristics. These characteristics are discussed below.

The intertidal sands of Facies B comprise fine rhythmically-laminated bedding produced by the alternation of sand and silt/clay laminae. This rhythmic alternation is thought produced by tidal flux. Silts and clays are laid down during high tide periods, while sand layers are laid down during periods of tidal drop and rise, and those low tide periods when river discharges are high enough to inundate the intertidal sands. As a consequence, the tidal sands of Facies B may be classified as tidal bedding, as described by Reineck and Singh (1980). These deposits most likely display neap-spring tidal cyclicity. Despite this statement, these deposits are characterised by considerable local variability, and do not always display such distinct tidal bedding. Further evidence of the tidal nature of these deposits is provided by the presence of shells at a number of locations along the lower reaches of the estuary.

The tidal characteristics of Facies C are the presence of marine and brackish-water floral species which form discrete layers to produce the characteristic bedding pattern of these tidal marsh deposits. In turn, the bedding characteristics of these tidal marsh deposits appear to yield evidence of their tidal character. Bed thickness have been used as a surrogate for sedimentation rate, thus yielding evidence of the changing rate of marsh sedimentation with age. Comparative analysis of the nature of this changing rate of sedimentation between tidal and fluvial environments suggests that characteristic differences exist between the two. These differences are related to differences in effective discharge relationships, flood frequencies, and sediment transport mechanisms within tidal and fluvial environments.
Preservation Potential of Tidal Deposits

This study has identified a number of deposits which display sedimentologic and stratigraphic evidence of their tidal origin. It is important to determine, however, whether these deposits are likely to be of significance to geologists investigating ancient estuarine sequences. This prompted discussion of the preservation potential of the tidal deposits identified here. This discussion concludes that these intertidal deposits are most unlikely to be fully preserved in sequence within this high-energy, laterally confined system, agreeing with the earlier comments of Nio (1984). Despite this, the tidal marsh deposits of Facies C likely will be preserved as large intact blocks of slump debris at the base of estuarine channel sequences. The cohesive nature of these sediments enables these blocks (which currently litter the near-bank zone) to retain their bedding form and resist erosion. If buried by channel sands, these likely will become preserved, as has been noted elsewhere.

While the process of erosion and eventual preservation of large cohesive blocks of sediment is not limited to tidal environments, these deposits may still be important indicators of depositional environment. This is because the slump blocks retain their original bedding form, (which appears indicative of a tidal environment) and halophyte organic material.

Comparative Analysis of Fluvial and Estuarine Fine-Grained Deposits

A limited number of previous studies have suggested that distinct differences exist between fine-grained sedimentary deposits within fluvial and tidal regions. This
prompted a comparative analysis of tidal deposits of Facies C (tidal marsh) with fine-grained sedimentary deposits along the riverine reach of Squamish River. The fine deposits within the fluvial environment investigated here, formed within an abandoned channel section, probably an oxbow produced by meander neck-cutoff. The results of sedimentologic and stratigraphic analyses of these deposits are summarised in Table 6.1. These differences highlight a number of important tidal characteristics of fine-grained deposits within Squamish estuary.

**Sedimentologic Evidence of the Fluvial/Tidal Transition Along the Estuarine Reach**

Bank sections exposed along Squamish River record the changing nature of sedimentary sequences along the tidal gradient. These sequence changes in turn record the decreasing influence of tidal control on sedimentation with increasing distance upstream. An examination of broad-scale changes in sedimentary sequence, and a more detailed examination of changes within facies of known tidal origin (Facies B and C) has yielded considerable information regarding the nature of the fluvial/tidal transition. The main conclusions from these analyses are given below.

1) Estuarine sequences are characterised by thick, extensive deposits of fine-grained overbank deposits, primarily those of Facies C and E. This reflects the dominant mechanism of deposition within tidal environments - deposition from suspension. With increasing distance upstream these fine-grained deposits become thinner and eventually pinch-out. This results in an increase in the mean grain-size of sequences up estuary. This decreasing facies thickness reflects higher energy conditions (less tidal modification of flow) and less sedimentation from suspension.
<table>
<thead>
<tr>
<th>CHARACTERISTIC UNDER INVESTIGATION</th>
<th>TIDAL DEPOSITS</th>
<th>FLUVIAL DEPOSITS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Facies Geometry</td>
<td>Sheet-like. Over 3 200 m in longitudinal extent, and over 2 000 m in lateral extent. In longitudinal section, these deposits are lens-shaped.</td>
<td>Very limited in extent to a few exposures in section. Lateral contact to coarse sands appears abrupt, but not erosional.</td>
</tr>
<tr>
<td>Organic Content</td>
<td>Almost exclusively halophyte marsh vegetation. These form thin, parallel layers of flattened organics between inorganic beds. Organics are almost entirely autochthonous. Layers are of constant thickness.</td>
<td>Twigs, leaves, pine needles, and non-halophyte grasses. These form thin, sometimes discontinuous layers between inorganic beds. Organics are almost entirely allochthonous. Layers are of variable thickness.</td>
</tr>
<tr>
<td>Bedding Form</td>
<td>Bed thickness decreases with increasing height in section. Rhythmite bedding produced by constant alternation of organic and inorganic layers. The nature of decreasing bed thickness upsection may be characteristic of tidal environments.</td>
<td>Bed thicknesses are highly variable, and show no distinct trends upsection. Rhythmitic bedding may develop, but need not necessarily do so, as organic layers are not always present in section.</td>
</tr>
<tr>
<td>Sediment-Size Characteristics</td>
<td>Beds contain considerable amounts of clay (average 16%). Sand, silt, and clay content of beds in section show distinct trends with increasing marsh elevation. No relationships exist between bed thickness and sand, silt, or clay content.</td>
<td>Beds contain few clays (average 2%), and are dominated by sand. No trends exist between sand, silt, and clay content with elevation. No relationships exist between bed thickness and sand, silt, or clay content.</td>
</tr>
</tbody>
</table>

Table 6.1. Characteristics of fine-grained sedimentary deposits within tidal and fluvial environments of Squamish River.
2) Bank sections along the lower reaches of Squamish estuary reveal simplistic, predictable sequence successions between a small number of distinct facies. With increasing distance upstream, however, facies successions are no longer simple. Instead, sections comprise a number of facies which may not be of constant position in sequence. Facies often are interbedded with coarse sand deposits which may erosionally overlie fine-grained overbank deposits. The simplistic sequences become broken by injections of coarser material from random riverine flood events.

3) Bedding form changes up estuary from fine parallel wavy (or undulatory) bedding, to a dominance of ripple bedding forms such as climbing ripple lamination. As before, this change in bedding form reflects the decreasing importance of suspended-sediment deposition as energy conditions increase up estuary.

4) Estuarine sequences display fining-upward trends produced by the transition from Facies A through B to C. Mid-estuarine sequences show a reverse of this trend as sequences become capped by deposits of fine to medium sand (Facies E and F). The riverine flood deposits of Facies F are produced as tidal influence decreases with the increasing elevation of the delta surface.

5) At those locations where tides exert considerable control on sedimentation, Facies Contacts most commonly are gradational, as highlighted by the transitions from Facies B to C, and C to E. When fluvial depositional processes dominate, as with the coarser deposits of Facies D and F, contacts are always abrupt and may be erosional. This simply reflects the higher energy conditions associated with decreasing modification (slowing) of flow by tides.
6) The deposits of Facies B display considerable changes up estuary, as sediments gradually lose the characteristic tidal bedding form. Initially, sand bed thickness increases and display an increasing amount of ripple bedding forms as opposed to fine parallel bedding. In addition to this increasing sand bed thickness, silt/clay beds decrease in both thickness and number within section. This results in a coarsening of Facies B deposits up estuary, and a lessening degree of rhythmicity and regularity produced by the alternation of sand and silt/clay beds.

7) The tidal marsh deposits display similar changes which reflect the decreasing tidal influence up estuary, though these gradual changes are often disturbed by local variability. As a general rule, however, the sediments within marsh rhythmite beds increase in size up estuary. Accompanying this increase in grain-size, rhythmite beds appear to thin up estuary, reflecting decreasing modification of flow by tides and decreasing tidal amplitude. Toward the upstream limit of these deposits beds appear to thicken again, reflecting the dominance of riverine depositional processes.

These gradual changes record the decreasing tidal modification of flow, and a consequent increasing importance of riverine depositional processes. In addition to these up estuary changes in depositional product and character, there exist similar changes upsection. This is seen as a coarsening-upward sequence superimposed on a fining-upward sequence. This latter sequence is the transition from subtidal to intertidal sands, to low and then high marsh deposits (Facies A to C). This sequence represents an increasing importance of tidal control on sedimentation, which becomes most evident near the mean low water mark. Along much of the estuarine reach, these deposits are overlain by Facies E and F, which represent a coarsening-upward sequence cap. This cap reflects the decreasing relative importance of tides as the delta/floodplain surface...
accretes up to and beyond the mean high tide water mark. A conceptual model of upsection changes of the relative influence of tidal and fluvial depositional processes is shown in Figure 6.1. In this model, tidal influence rapidly increases as the surface is raised above mean low tide level, then gradually decreases as the surface continues to accrete. Toward the top of estuarine sections, fluvial processes dominate, with the possible exception of the deposits of facies G, which show far greater tidal influence than do the deposits of Facies F. In addition to showing the relative influence of tides and river flow, Figure 6.1 also shows the Facies associated with these relative changes, and the relative components of sand and fines present within deposits at different heights in section. This figure essentially mirrors the form of section diagrams drawn for estuarine facies (for example, those shown in Figure 3.3, p. 82).

The landward limit of detectable tidal deposits marks an important boundary termed the bayline. Within Squamish estuary this boundary is marked by the upstream limit of Facies B deposits. These are observed in section up to 3 300 m upstream from the river mouth, though it is considered more accurate to report the bayline location as a straight line distance from the delta head. By this definition, the bayline is located 3 100 m up estuary. Sedimentologists should be cautious, however, when determining bayline location, as this marker may not necessarily record the present upstream limit of tidal deposition. There should exist a lag between shifts in the upstream-most location of tidal deposition and erosion of tidal deposits in regions no longer subject to tidal sedimentary influence. Within Squamish estuary this is not of concern, however, as the upstream-most tidal deposits are known to have been produced by recent flows. The location of this bayline in relation to the estuarine reach of river (around 5 500 m) reflects the fluvial dominance of this system.
Figure 6.1. Conceptual model of changes of the relative influence of tidal and fluvial depositional processes with increasing elevation of the delta/floodplain surface. This model is based on sedimentologic and stratigraphic analysis of estuarine sequences. Also shown are the relative proportions of sand and mud within deposits at different heights in section.
Rates of Estuarine Channel Bank Erosion

Rates of estuarine channel bank erosion have been assessed from 8 aerial photographs over a 33 year period from 1957 to 1990. This period of photographic record allows comparison of erosion rates prior to and after construction of a river training dyke which served to isolate flow to a single channel to the west of Squamish delta.

Within the estuary, bank erosion has been assessed within 4 zones which reflect different morphologic zones of Squamish River. The results of this study are listed below.

1) Within 3 of the 4 zones under investigation, rates of bank erosion remain fairly constant over the entire period of study; dyke construction does not appear to have led to an increase in the rate of bank erosion.

2) Squamish west delta has been decreasing in size over the period of study. Between 1957 and 1990, over 150,000 m$^2$ of west delta deposits have been lost by direct erosion due to channel migration, and by dredging which aimed to straighten the estuarine channel. There is no indication that this rate of erosion has increased since 1972, but results suggest that the west delta will keep eroding and reducing in size as this reach of Squamish River continues to migrate westward.

3) Within zone 4 (that closest to the dyke), rates of channel bank erosion have increased dramatically since dyke construction. Average yearly losses over the 12 year period from 1957 to 69 are 2,151 m$^2$/yr, whereas over the 18 year post-dyke period (1972-90) erosion rates average 7,592 m$^2$/yr. This zone of rapid erosion includes the areas of
intense post-dyke erosion identified by Bell (1975), Zrymiak and Durette (1979), and Levings (1980). Bank erosion within this zone is associated with the loss of material along the outer bank of a large meander which is migrating both across and down valley.

The erosion within this zone is of great importance as the river training dyke parallels the channel bank at this location. At one location (which recorded erosion over the 1990 to 1992 period) within this zone the channel bank lies within 5 m of the toe of the dyke. It is felt that, without remedial engineering, Squamish River will continue to migrate eastward at this location, and will eventually breach the dyke. If the recorded short-term rates of erosion continue, dyke stability may soon be threatened. If the channel were permitted to continue migrating freely, Squamish River main flow would eventually switch to the central channel. If this distributary channel switching is prevented, the west delta will continue decreasing in size as it has over the past 33 years.

Observations of Erosion

On one occasion, rapid erosion of a portion of west delta sediments enabled a number of observations to be made regarding the erosional process within the estuary. This erosion was initiated by the removal of the protective toe along the outer meander bank. As such, erosion (or at least the initiation of that erosion) appears primarily controlled by basal endpoint control (Thorne, 1982). Once initiated, further erosion is determined by a number of factors including stratigraphy, tidal stage, localised velocity fluctuations associated with boil production, and wind-generated waves.

Erosion occurs through the removal of the basal sands of Facies A, which leads to undercutting of the finer tidal marsh deposits. As these fine deposits are cohesive, bank
retreat occurs as large blocks of sediment (up to 5 m$^3$ in size) collapse. This erosional process continues until these slump blocks litter the near-bank zone, reforming a protective toe which covers the basal sands.

**Tidal Marsh Accretion**

Rates of west delta marsh accretion have been determined from a number of stakes on three occasions between May 1992 and November 1993. Though many of these stakes became disturbed or removed, a number of observations may be made regarding accretion rates and patterns within the west delta. These are listed below.

1) Accretion rates record considerable spatial variability at this scale of investigation (stakes were 50 and 100 m apart). Locally high rates of accretion are recorded at stakes close to the river, reflecting the importance of the fluvial source of sediments. As well as the highest rates of sedimentation, stakes closest to the river record the greatest variability of accretion. This variability decreases away from the river.

2) Accretion rates are highly seasonal, with greatest rates occurring during autumn and winter. The majority of accretion is thought to occur during a limited number of riverine flood events over winter months. This seasonality of accretion has a number of implications for marsh sediment sampling technique and timing.

3) There appears to be a shift in the location of maximum accretion from summer to winter months. Between May and August 1992, 10 of 12 transects recorded greater accretion toward the centre of the marsh (between 60 and 160 m from the river bank) than toward the marsh edge (10 m from the bank). The following measurements from
these same stakes (13 and 18 months after placement), however, revealed something of a reverse of this trend. Of the 12 transects that were able to be measured on these combined occasions, 9 recorded greatest accretion at those stakes closest to the marsh edge. This shift in the location of maximum accretion may reflect seasonal changes in the dominant depositional process. During summer months, accretion occurs gradually during each high tide period, as shown by data from marker beds. Accretion is greater away from the river, as flow velocities decrease and allow greater sedimentation of fines. During winter months, marsh accretion is driven primarily by high energy river flows which transport then deposit considerable amounts of material on the delta surface. As such, accretion rates are highest at channel-proximal locations.

It is noted that contemporary rates of accretion of the west delta cannot be considered in relation to the tidal marsh rhythmite beds of Facies C, as these formed prior to river training dyke construction. Current accretion rates reflect the modified flow regime, and so can only be considered in relation to the deposits of Facies D.

**An assessment of the Influence of Dyke Construction**

The completion of a river training dyke in June 1972 led to a number of changes along the lower 4 000 m of Squamish estuary. The most immediate change was the reduction of channel width that occurred along the lower 3 500 m of the river as the dyke sealed-off flood flow to the central channel. In addition, dyke construction has led to a reduction of effective floodplain width along the entire estuarine reach, as floodwaters are no longer able to inundate the central and eastern deltas. Along the lower 3 500 m of the estuary, effective floodplain widths have been reduced by an average of 60%.
The decreasing width of the effective floodplain has resulted in an increase of floodwater depth, and an inferred increase of flow velocity through certain reaches. Analysis of erosion data show some evidence to suggest that the increased depth and velocity of flood flows after 1972 have caused an increase in the rate of erosion within certain parts of the estuary. In addition to these increased rates of erosion, dyke construction has led to the formation of an anomalous coarse sedimentary sequence within the west delta tidal marsh. Along the lower 800 m of the estuary, the combined effects of channel and floodplain width reduction has been an increase in the depth, duration, and velocity of floodwaters inundating the delta surface. These changes have led to the deposition of thick, coarse-grained deposits of Facies D, which essentially represent levee deposits. At their maximum, these deposits are 134 cm thick, and have been identified as the product of post 1972 flows by comparison with contemporary accretion rates, and dating of deposits.

**Description and Origin of Large Stones on the West Delta Surface**

Over 50 large stones (ranging in size up to around 60 cm in diameter) were found on the surface of the delta in 1993. These stones were all rounded and appeared placed or dropped on the delta surface; no disturbance (such as roll or drag marks) of the surrounding soft fine sediments was observed. These gravels and cobbles often were clustered in discrete groups, and were not associated with another material such as coarse sand or gravel. They were mostly found within 10 m of the river bank, to a point 600 m upstream from the southern limit of the west delta.

These stones are not deposited by rockfall, riverine flood, or river ice, but instead are deposited from large organic debris. They are transported into the interdistributary bay.
environment within the root systems of uprooted trees and root wads which are able to pass onto and over the delta during combined flood and high tide periods. Occasionally, these trees and root wads become stranded on the delta (for periods of days to years) as stage falls; this particularly occurs toward the river bank, where trees get caught then stranded on the levee. If stones are present within the root systems of these trees, they may become removed and fall to the delta surface, where they have the appearance of dropstones. Stone removal is believed to occur as wind-generated waves strike the root systems during periods of delta floodwater inundation.

These stones will become buried and preserved as large clasts within a matrix of silts, clays, and fine sands. As trees eventually float-off the delta surface, there will be no sedimentologic evidence of the transport and deposition mechanisms of these clasts, leading to the danger of misinterpretation of the mode of origin of these deposits.
APPENDIX A

Sediment sample information for the rhytmite bedding at site 01. Data include sample number, depth in section, bed or sample thickness, and sand, silt, and clay content.
<table>
<thead>
<tr>
<th>Sample</th>
<th>Depth from cm</th>
<th>Depth to cm</th>
<th>Rhythmite thickness cm</th>
<th>% clay in each rhythmite</th>
<th>% silt in each rhythmite</th>
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APPENDIX B

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APPENDIX C

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