HOLOCENE RELATIVE SEA-LEVEL CHANGE:
VARGAS ISLAND, BRITISH COLUMBIA

by

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Abstract

A new relative sea-level curve is presented which extends the previous sea-level curve for the west-coast of Vancouver Island from 4,000 yr BP to about 7,900 yr BP. The superceded sea-level curve, which was assembled mostly as a by product of archaeological research, showed late-Holocene emergence, but the early and mid-Holocene sea-level pattern was unknown. Thus, two alternative post-glacial relative sea-level scenarios existed: i) that of monotonic emergence from the post-glacial marine limit; or ii) that of rapid post-glacial emergence to some early-Holocene lowstand followed by Holocene submergence and, finally, the previously documented late-Holocene emergence.

The significance of the Vargas Island sea-level curve, developed from precise topographic surveys of modern and emergent marine landforms at 16 sites, and detailed stratigraphic and paleo-ecological analyses at five of those, is that it provides solid evidence for the latter scenario, with mean sea-level following Holocene submergence attaining a height of 2.4 - 3.4 m higher than today about 5,100 yr BP. Following a stillstand, sea-levels did not start to fall until after 4,000 yr BP. Holocene submergence is shown to be a regional pattern and it is formally proposed that it be given the name the *Clayoquot Transgression* to distinguish it from the early post-glacial marine inundation. The 1,100 year stillstand, referred to as the *Ahous Bay Stillstand,* is well constrained by eight new radiocarbon dates.

As well as extending the relative sea-level curve, this study, by comparison with appropriate models from the literature, attempts to resolve the geodynamic mechanisms producing the observed sea-level pattern. It is concluded that the final 1,000 years of the Clayoquot Transgression resulted from a prolonged isostatic response, possibly forebulge collapse, while late-Holocene emergence is attributed to tectonic uplift. Of particular concern is the assessment of seismic hazard along this portion of the Cascadia subduction zone. No unequivocal, co-seismically produced landforms were observed on Vargas Island. The dominant Holocene landforms (beach ridges, wave-cut and constructional terraces, and marine scarps) represent two discrete strandline elevations. The higher, the 6 m asl strandline, formed during the Ahous Bay Stillstand, 5,100 to 4,000 yr BP, while the lower, the 4 m asl strandline, represents a stillstand ending about 2,200 yr BP suggesting unsteady late-Holocene emergence. Initial emergence and eventual abandonment of the 6 m strandline might reflect co-seismic uplift, while abandonment of the 4 m strandline may have resulted from co-seismic uplift or glacio-eustatic sea-level oscillations. These two processes cannot be distinguished based on present evidence. Continuous emergence for the last 2,200 years is suggested.

May the wild woods remain wild, full of thrush in June, with wolves free to hunt and deer to flee, and with eagles spirallng down to the sparkling sea in their talons-locked dance of love.

Friele

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CHAPTER 1: Introduction

The Study of Relative Sea-level

The study of relative sea-level change involves the examination of the apparent movement of the land with respect to the sea surface (geoid). Geodynamic mechanisms that contribute to relative sea-level change are many and varied (eustasy, isostasy, and tectonics; see Chapter 3), and their interplay can be difficult to resolve (e.g., Radtke, 1987). Determining the dominant processes responsible for observed relative sea-level changes at a site requires reference to an appropriate model against which sea-level data can be tested. The traditional approach assumed that the observed land/sea changes could be attributed to worldwide and simultaneous changes in the geoid. This approach, which allowed the altimetric correlation of marine terraces globally, is now recognized to be invalid, because i) it has been shown that stable coastlines do not exist (Fairbridge, 1968), and ii) the shape of the geoid is always changing (Mörner, 1987).

Without a fixed point of reference it is difficult to resolve the factors contributing to land/sea changes. One way to decipher the various factors is to develop a theoretical model of the dominant process driving sea-level change at a site, and to compare the model's predictions to the observed pattern of sea-level change. For instance, relative sea-level changes following deglaciation can be predicted by modelling the crust's isostatic response to load removal. Such a model can then be tested against sea-level data from glaciated regions, both to refine the model, and to reveal significant inconsistencies, between the model and the observed pattern, that might be attributed to factors other than glacio-isostasy and glacio-eustasy (e.g., Clark *et al.*, 1978).

Along subduction zones relative sea-level changes also have a characteristic signature (Fitch and Scholz, 1971). The most dramatic crustal movements occur co-seismically, and can result in sudden uplift/subsidence on the order of metres. For example, Plafker (1965) documented the co-seismic vertical movements accompanying the 1964 Alaskan earthquake. In that event a broad crustal warp about 800 km long and 300 km wide, caused uplift 70 - 150 km from the margin and subsidence 160 - 270 km from the margin. This broad warp has been shown to be a consistent deformation pattern associated with megathrust earthquakes in subduction zones (West and McCrumb, 1988). Plafker and Rubin (1978) were able to calculate the earthquake recurrence frequency of the active Alaskan coastline based on a flight of emergent marine terraces on Middleton Island, where the most recent marine terrace was produced by the 1964 earthquake. Such geological estimates of the return periods of megathrust earthquakes in subduction zones indicate recurrence times ranging from 300 to 2000 years and averaging 980 years (West and McCrumb, 1988).

Recent Sea-level Studies in the Pacific Northwest

Interest in relative sea-level change in the Pacific Northwest has recently increased given the concern over the potential for megathrust earthquakes along the Cascadia subduction zone (Atwater, 1987; Rogers, 1988). Depositional sequences very similar to those produced during megathrust earthquakes in Alaska (Plafker, 1965 & 1990) and in Chile in 1960 (Wright and Mella, 1963) have been documented in estuaries along the outer-coast of Washington and Oregon (Atwater, 1987 & 1988; Peterson and Darienzo, 1988 &1991), although no such earthquakes have occurred historically. These estuarine deposits seem to record co-seismic subsidence in the landward limb of the crustal warp; regions that might record co-seismic uplift would be trenchward of the Washington and Oregon coastlines.

The outer-coast of Vancouver Island (Figure 1.1), a region between 75 km and 100 km from the plate margin, might record co-seismic uplift (Dragert and Rogers, 1988), but sea-level changes on the outer-coast have not been examined in a neotectonic context. A continuous record of relative sea-level changes for the last 4,000 years along the outer-coast of Vancouver Island is provided by Howes (1981c) and Clague *et al.* (1982). The assembled record, which comes mostly from dated archaeological sites and the occasional littoral marker, indicates long term emergence on the order of 1 mm/yr to 2 mm/yr. Contemporary emergence on the central west coast of Vancouver Island, at Tofino, is about 1.25 mm/yr (Wigen and Stephenson, 1980). This late-Holocene rate of emergence on the west coast of Vancouver Island is low compared to trenchward areas in Alaska (Plafker, 1965, 1990; Plafker and Rubin, 1978), but is within the range reported from other subduction zones around the world (Lajoie, 1987).

Other effects that might be invoked to explain documented late-Holocene sea-level patterns along the Pacific Northwest coast (i.e., emergence along the west-coast of Vancouver Island and submerged estuarine sequences in Washington and Oregon) are forebulge effects (Clark *et al.*, 1978; Quinlan and Beaumont, 1981), and glacio-eustatically induced, rapid sea-level oscillations (Mörner, 1976; Fairbridge and Hillaire-Marcel, 1977; Cronin, 1983) superimposed on tectonically forced landward tilt. Tectonic and alternative models of Holocene relative sea-level change are discussed more fully in Chapter 3.

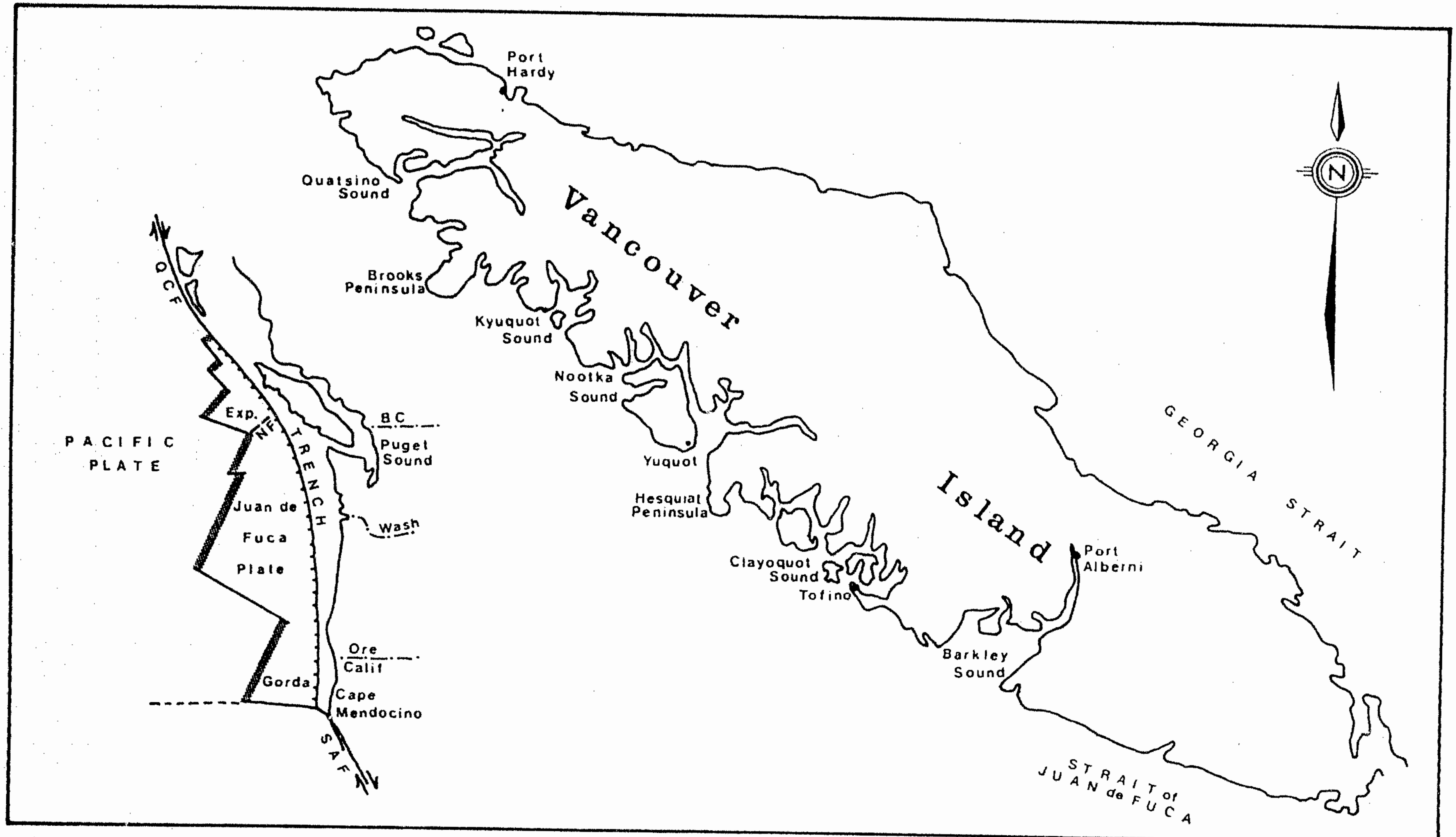


Figure 1.1: Vancouver Island, with place names and tectonic framework. QCF - Queen Charlotte Fault; SAF - San Andreas Fault; NF - Nootka Fault; Exp. - Explorer Plate; Gorda - Gorda Plate.

The Research Framework

Research Objectives

The specific goals of this research were: i) to document the Holocene sea-level record at a site on the west coast of Vancouver Island and extend the existing relative sea-level curve of that region; and ii) to discriminate between the factors driving relative sea-level change by comparison of the pattern revealed in the construction of the sea-level curve with the likely pattern produced by a particular factor, such as the migrating forebulge or co-seismic uplift.

Choice of the Research Area

The coastal plain of the west-coast Vancouver Island is ideal for studying relative sea-level change because of the many islands, inlets, lagoons, and low elevation lakes and bogs along the coastline. Lakes and bogs are particularly important features because they preserve sediments and fossil assemblages that can record sea-level changes (Hafsten, 1983). Low relief is also desirable in order to record the signal of sea-level change on the west coast of Vancouver Island, since in extreme ice-marginal areas elsewhere the amplitude of post-glacial and Holocene sea-level change is known to be relatively small (< 30 m asl) (e.g. south Norway; Hafsten, 1979 & 1983).

Of the sites on the west-coast of Vancouver Island that present themselves for study, Vargas Island, in Clayoquot Sound (Figure 1.2), is perhaps the most attractive. The island displays an extensive sequence of emergent parallel beach ridges that are a prominent, and unique landform on the west-coast of Vancouver Island (Figure 1.3). Although Howes (1981a) had noted the ridges as an obvious indicator of late-Holocene emergence, and Dragert and Rogers (1988) suggested that they may have resulted from co-seismic upheavals, their genesis and relationship to past sea-levels was not well understood. The only superficially similar landform on the west-coast of Vancouver Island is a set of strandlines adjacent to the lagoon on the north side of the Brooks Peninsula (Howes, 1983b). The Vargas Island beach ridges were therefore the stimulus for this study of Holocene relative sea-level change.

Outline of the Thesis

This thesis consists of seven chapters. Chapter 2 describes the regional setting of the study area, while Chapter 3 provides a theoretical discussion of sea-level change focussing on the factors driving relative sea-level change along the Pacific Northwest

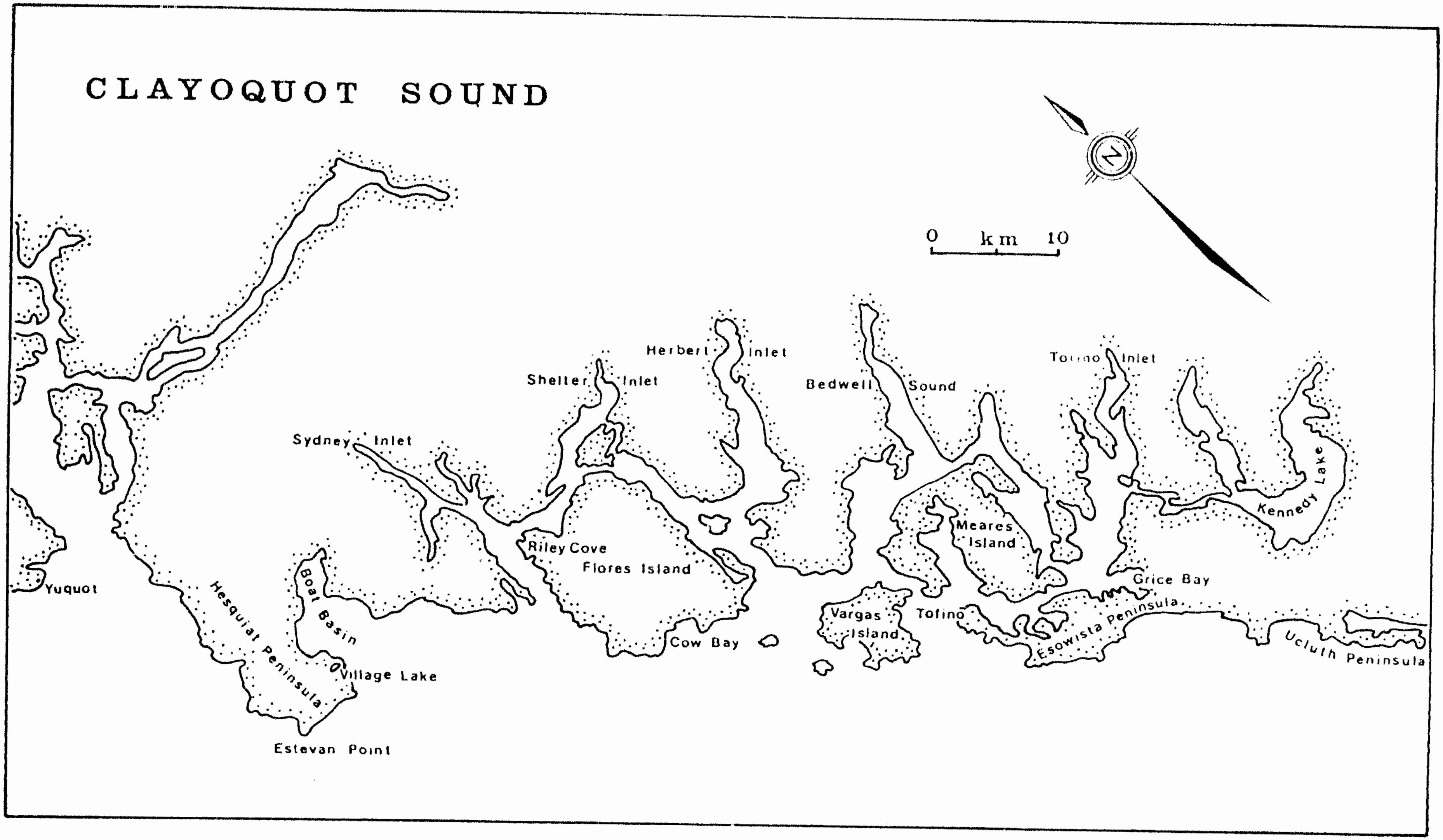


Figure 1.2: Clayoquot Sound, with place names used in this thesis.

Coast. In Chapter 4, I present a critical account of the methods in sea-level curve construction and outline the specific techniques used in this study. Chapter 5 presents the results of the research on a site-by-site basis. In the initial section of Chapter 6, I discuss these results and outline a sea-level curve for Vargas Island. The curve's regional applicability is then discussed and the probable driving forces behind the observed sea-level changes are examined. Finally, Chapter 7 presents the conclusions of the research and suggests targets for further research.

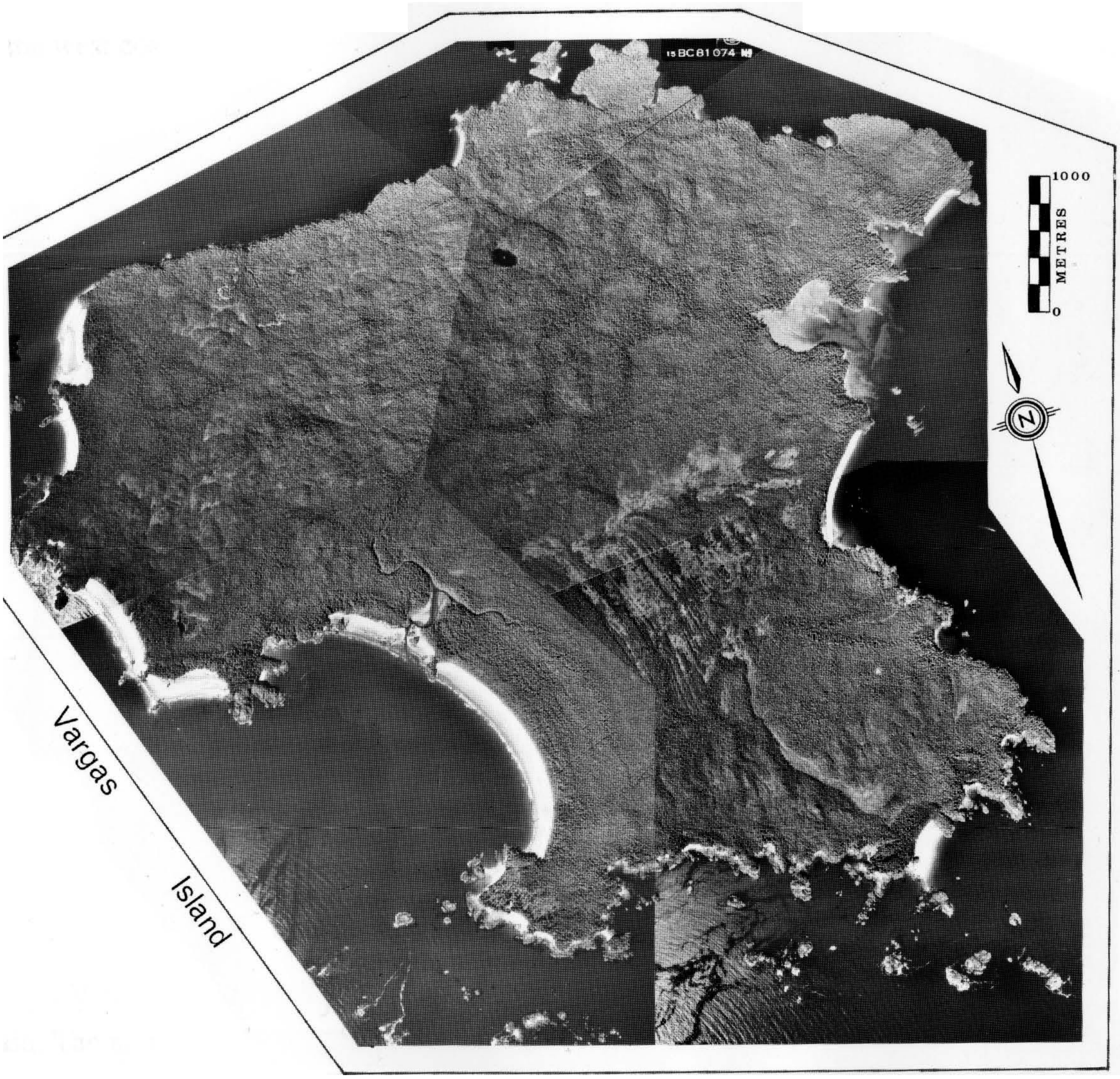


Figure 1.3: Airphoto mosaic of Vargas Island. Note the ridges in the central portion of the island.

CHAPTER 2: Regional Setting

Introduction

This chapter contains descriptive information which is necessary to stage the following chapters. The local physiography, and the oceanographic and climatic variables which act on that landscape are presented as, together, they influence the development and preservation of the sea-level change signal. The Quaternary history and tectonic framework are fundamental to the understanding of the dominant geodynamic mechanisms operating on the west coast of Vancouver Island.

Physiography

Regional Physiography

Southwestern Vancouver Island is comprised of the mountainous Vancouver Island Ranges, attaining heights of just over 2,100 m, and the very narrow, low lying Estevan Coastal Plain, a Tertiary erosion surface, rarely exceeding 50 m in elevation (Holland, 1976). The Estevan Coastal Plain extends from the Brooks Peninsula, southeastward for almost 270 km, to the entrance of the Strait of Juan de Fuca (Figure 1.1). It averages 1 - 5 km in width, increasing to a width of 9 - 13 km on the Hesquiat Peninsula.

The Vancouver Island Ranges are dissected by deep, glacially carved valleys, which down-valley become large transverse fjord systems. From the north to south these are: Quatsino Sound, Kyuquot Sound, Nootka Sound, Clayoquot Sound, and Barkley Sound. On the outer-coast, where the fjord systems dissect the coastal plain, are many low-lying islands. Inland, toward the fjord heads, fjord sidewalls are steep, and the shoreline is unbroken by bays or beaches. Between the sounds the coastline is rocky and fully exposed to the Pacific.

Physiography of Vargas Island

Vargas Island is about 30 km^2 in area and lies entirely within the Estevan Coastal Plain. The highest point on the island is an isolated hill about 160 m in elevation, but more than half of the island area is less than 60 m in elevation, with about 30% of Vargas Island less than 10 m above mean sea-level. On its exposed sides, Vargas is adorned with long sandy beaches, while on the inside, along the channels, the shoreline is more rocky.

In the creeks on Vargas Island and the islands in the vicinity there is no evidence of buried peat sequences exposed along their banks, as reported by Atwater (1987) for the outer-coast of Washington. The paleo-littoral features are generally found behind the modern beaches. The most notable of the emergent features on Vargas Island is the central bog and ridge system which occupies a broad band, approximately 1.5 km wide and 3 km long, across the southern half of the island (Figure 2.1). The ridges are parallel to the present shoreline configuration, and appear to have resulted from the progradation of the modern Ahous and Kelsemaht Beaches (cf. Bird, 1976). This landscape unit is informally referred to as the Ridge System.

The Ridge System is extremely level. During rainy periods the Ridge System becomes a broad shallow pond divided by long parallel islands. Filling and draining of this pond occurs rapidly, with flashy response in the creeks, and after a week of no rain the bogs can be relatively dry, winter or summer. When the creeks are in full flood, the bays they drain into may be red with iron, especially if the bays, such as Hovelaque's Bay, are poorly flushed by successive tides. This high concentration of iron leads to the formation of ferrous cemented sands at or just above the water table.

A tombolo-like feature in the Ridge System, just inland from Kelsemaht Beach, connects the uplands of the north half of the Vargas Island to the upland area behind Buckle Bay, and likely marks the inland limits of the paleo-Ahous and Kelsemaht Beaches (Figure 2.1). A long, narrow lowland extending south from the Ridge System to Medallion Beach, informally named South Bog, separates the upland behind Buckle Bay and Hovelaque's Bay from the upland extending to Ahous Point. It seems that when the sea washed against the paleo-tombolo and occupied the South Bog lowland, Vargas Island would have been an archipelago of smaller islands. Around the upland margins of the Ridge System and along the narrow part of South Bog are over-steepened slopes, exposed diamict, and rocky outcrops all of which suggests that these represent a single paleo-strandline.

On the northwest side of Vargas Island the modern beaches have well developed foredunes. The northwesterly wind has drifted sand high against rock outcrops, blocking drainage, and giving rise to the dune lake behind, known as Monk's Lake. On the southern of these beaches, the Dune Beach, the back-beach dune attains a height of 15 m asl at its crest, before dropping steeply into the forest. In Ahous Bay, immediately to the south, foredune development is minimal. This lack of developed foredunes on Ahous Bay probably results from a low sediment supply; i.e., sand is likely trapped by the rocky point south of Monk's Lake.

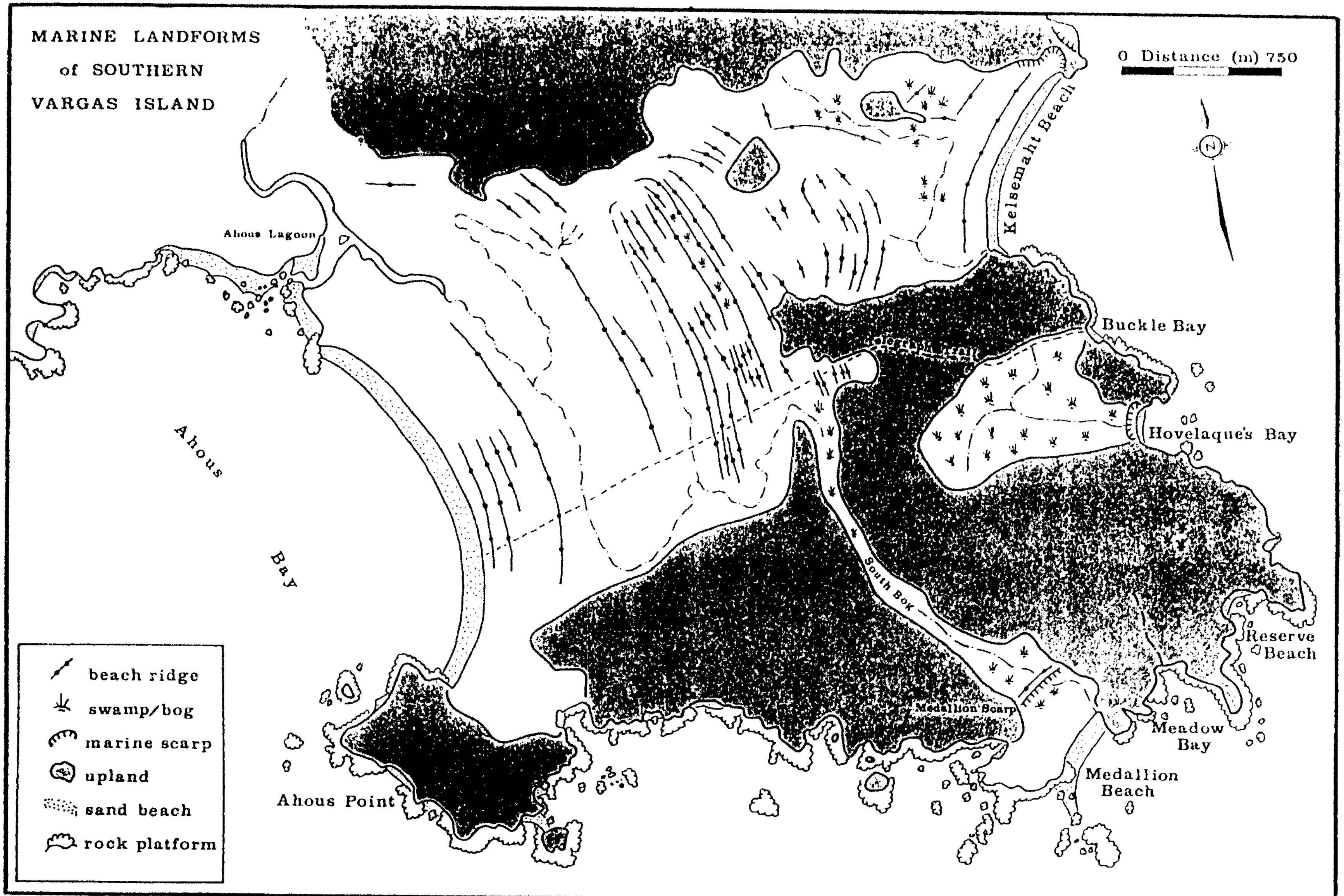


Figure 2.1: Marine Landforms of Southern Vargas Island. Shaded areas represent upland topography higher than about seven metres above mean sea-level. Note that the ridges conform to the modern shoreline configuration.

Climate and Vegetation

Clayoquot Sound has a mesothermal equable, rainy to humid climate. The mean annual temperature is 9.1^{o} C, with the mean July temperature of 14.1^{o} C and mean January temperature of 4.5^{o} C (B.C.D.A., 1975). The mean precipitation is 3,028 mm, of which 38 mm fall as snow.

The study area is in the Hypermaritime subzone *(CWHd)* of the Coastal Western Hemlock Biogeoclimatic zone (Nuszdorfer, 1985). *Thuja plicata* Donn. (Western red cedar), *Tsuga heterophylla* (Raff.) Sarg. (Western hemlock), and *Picea sitchensis* (Bong.) Carr. (Sitka spruce) are the common trees. Many large old individuals of Sitka spruce grow along the shore. *Gaultheria shallon* Pursh (salal) and *Vaccinium ovatum* Pursh (evergreen huckleberry) are dominant understory shrubs in the forest. *Pinus contorta* Dougl. (Lodgepole pine) and *Chamaecyparis nootkatensis* (d.Don.) Spach (yellow cedar) occur in boggy or incipient boggy sites.

Winds and Tides

The prevailing winds affecting the south-west coast of Vancouver Island are the result of the Aleutian Low, which builds in intensity from August to January and tapers off by July, and the Pacific High, which dominates the summer months between June and August. At Estevan Point, on the Hesquiat Peninsula, cyclonic conditions dominate the months September through April with prevailing winds from the southeast, while anticyclonic conditions dominate from May through August with prevailing winds from the northwest. Southeast winds occur about 40% of the time for the stormiest months of October through December, while northwest winds occur about 30% of the time from May through August, reaching a peak in the months of June and July (Thomson, 1981).

Tides at Tofino are of the mixed semidiurnal type. The mean tidal range at Tofino is 2.8 m and that of the large tide is 4.1 m. The recorded extreme of higher-high water was 2.7 m asl, exceeding the mean large tide by 0.7 m (C.T.C.T., 1990).

Quaternary Geology and Sea-level Change

Ice of the Fraser Glaciation (Armstrong *et al.*, 1965) began to build up in the mountainous regions of British Columbia as early as 29,000 yr BP in the north and 25,000 yr BP in the south as indicated by the diachronous growth of the Quadra Sand in Georgia Strait (Clague, 1977), and correlative sediments on Vancouver Island (Howes, 1981b, 1983a). However, this ice remained largely confined to mountainous valleys until after 20,000 yr BP on northern Vancouver Island and 18,000 yr BP in the lower mainland and southern Vancouver Island (Clague *et al.*, 1980). The Puget lowland was free of ice until after 15,000 yr BP (Mathews *et al.*, 1970). The extreme west-coast of Vancouver Island at Tofino remained ice free until after 16,700 yr BP (GSC-2768; Clague *et al.*, 1980).

The ice advance over the coastal plain on the west coast of Vancouver Island appears short lived, between about 2,000 to 4,000 years. Deglaciation in northern Vancouver Island may have commenced prior to 13,630 ± 310 yr BP (WAT-721), a date obtained from basal peat in a bog near Port Hardy (Howes, 1983a). A date of 13,000 ±100 yr BP (GSC-2976) from wood (*Pinus contorta*) in marine silt on the Hesquiat peninsula (Howes, 1981c) and a date of 12,250 ± 790 yr BP (WAT-924) from lake gyttja on the Brooks Peninsula (Howes, 1983b) indicate ice-free conditions for the west-coast by about 13,000 yr BP. The oldest post-glacial date for southern Vancouver Island indicates that the interior uplands to 400 m were free of ice by 13,100 ± 130 yr BP (GSC-2223; Alley and Chatwin, 1979).

Early post-glacial sea-levels were higher than present in all of southern coastal British Columbia. The elevation of the marine limit varied regionally in proportion to ice thickness, thus at Vancouver post glacial sea levels were as high as 200 m asl, whereas on the west-coast of Vancouver Island where ice was thinner the marine limit stood at about 30 m asl (Clague *et al.*, 1982). Following deglaciation the land emerged rapidly and, in the inner and mid-coast regions of the mainland and eastern Vancouver Island, sea levels fell to a lowstand of about -12 m asl by about 9,000 years ago. The lowstand was followed by relatively rapid submergence and by the mid-Holocene sea-level was close to what it is today (Clague *et al.*, 1982; Williams and Roberts, 1989).

As stated earlier, the sea-level curve for the outer-coast of Vancouver Island is known only for the last 4,000 years. The occurrence of a number of late-Pleistocene deltas at 32-34 m asl in Nootka Sound and the date of 13,000 ±100 yr BP (GSC-2976) on wood in marine muds at Hesquiat Harbour indicate a post glacial marine limit of 32-34 m asl, ca 13,000 yr BP in the vicinity of Nootka Sound and Clayoquot Sound. Although the shape of the sea-level curve from the 13,000 yr BP to 4,000 yr BP is unknown, raised marine landforms attributable to this period have been identified (Howes, 1981c). Dates from Yuquot suggest that the high water level 4,000 years ago was 2.5 to 3.5 m higher than present. These dates, the oldest dates constraining Holocene sea-levels previously recovered for the west-coast of Vancouver Island, suggest emergence of 1-2 mm/yr. Contemporary emergence is 1.25 mm/yr measured from the tidal station at Tofino (Wigen and Stephenson, 1980).

Howes (1981c) observed that marine features above about 8 m (above the high tide limit) are continuously distributed and probably formed during the last deglacial sequence, whereas those below about 8 m are mostly Holocene in age, cluster at certain elevations, and might be ascribed to co-seismic uplift.

Tectonic Framework and History

The west-coast of Vancouver Island lies on an active tectonic margin known as the Cascadia Subduction Zone. Cascadia is composed of several subducting plate segments collectively referred to as the Juan de Fuca plate system. The Juan de Fuca plate system extends from northern Vancouver Island south to Cape Mendocino in northern California (Figure 1.1). The oceanic lithosphere of Cascadia is formed offshore along a stepped spreading ridge system, and is consumed by subduction at the continental margin, 75 - 100 km offshore, beneath the westward advancing America plate (Riddihough, 1977).

The Juan de Fuca plate system is composed of three separate plates: the Explorer, the Juan de Fuca, and the Gorda plates. The Explorer plate extends from Queen Charlotte Sound south to Nootka Island, where it is separated from the Juan de Fuca plate by a zone of seismicity known as the Nootka fault (Hyndman *et al.*, 1979). The Juan de Fuca plate extends from Nootka Sound south to northern California, where it abuts the Gorda plate. The Gorda plate extends south as far as Cape Mendocino. To the north and south of Cascadia are bounding transform faults; in the north the Queen Charlotte fault, and in the south, the San Andreas fault.

The Juan de Fuca plate system is a remnant of the Farallon plate, whose spreading ridge was consumed beneath the America plate about 30 Ma ago, an event which established the contemporary tectonic framework of Cascadia (Atwater, 1970). Riddihough (1977, 1984) has described the history of the Juan de Fuca plate system. Early in its history, until about 10 million years ago, the Juan de Fuca plate system acted as a single plate with subduction of 50 -60 mm/yr. In the last 10 Ma subduction has slowed dramatically, and breakup of the plate about 3 - 4 Ma ago was caused by the resistance of young, more buoyant lithosphere to subduction. First the Explorer plate broke off about 4 Ma ago, then the area of resistance shifted to the south causing the Gorda plate to break off about 2.5 Ma. Both these fragments are acting independently from the Juan de Fuca plate and seem to be rotating in a clockwise direction. Convergence of the Explorer and Gorda plates with the America plate is about 22 mm/yr, and is entirely due to overriding of the westward advancing America plate (Rogers, 1985). Present convergence of the Juan de

Fuca and America plates off southern Vancouver Island is 40 - 50 mm/yr (Hyndman *et al.*, 1990).

Bedrock and Surficial Geology

The rocks of the North American continental margin have been divided into discrete units called terranes which originated as island-arc systems or flysch deposits that were transported via plate motions from elsewhere to this latitude, and by subduction were emplaced, or docked, on the edge of the continent (Monger *et al.*, 1972; Muller, 1977a). The terranes on the southwest coast of Vancouver Island are, from the west, the Crescent Terrane, the Pacific Rim Complex, and the large Wrangellia Terrane, a volcanic-arc complex which forms the Vancouver Island Mountains. Trenchward of the Crescent Terrane is the modern accretionary wedge, formed by the scraping off of incoming sediments from the subducting slab. This geological sequence was completed with the subduction of the Farallon Ridge at the end of the Eocene some 40 million years ago, marking the establishment of the modern tectonic framework, as previously described.

The Pacific Rim Complex is exposed in outcrops along Ucluth and Esowista Peninsulas, and on Vargas Island (Muller, 1973 & 1977b). It consists of a Mesozoic sedimentary sequence, with two discrete rock units recognized: the Ucluth Formation lying above a sequence of sedimentary melange rocks. The sedimentary melange sequence, which dominates the Vargas Island shoreline, is composed of three units. The lower (stratigraphic position) melange unit is composed of interbedded mudstone, chert, sandstone, and green tuff, and contains exotic blocks derived from Ucluth Formation. Fossils from this unit give an early Cretaceous age. The middle and upper melange units are chaotically bedded assemblages of mudstone, sandstone, and rare ribbon chert (Brandon, 1989).

Surficial deposits of the Estevan Coastal Plain are generally drift materials of the Fraser Glaciation, and the subsequent post-glacial marine inundation (Howes, 1981a). They include fluvial and fluvioglacial materials, till, and marine deposits along the coastal fringe. Marine deposits include glaciomarine silts and clays, sand and gravel beaches, and lag deposits. Where the coastal plain is underlain by harder rocks, and the surface is irregular and hummocky, then a mantle of bedrock derived colluvium may occur.

CHAPTER 3: Literature Review

Introduction

Relative sea-level change results from the change of mean sea-level and/or crustal movements. This chapter discusses the interacting factors controlling relative sea-level change on British Columbia's coast. The dominant forces are glacio-eustasy, glacio-isostasy, and neotectonics. The chapter is biased in its discussion of tectonics, earthquake potentials, and evidence for seismic events in the geological record, because the issue of earthquake hazards in Cascadia is of concern not only to the scientific community, but also to residents of the area.

Clark *et al.* (1978) predicted global post-glacial relative sea-levels by integrating glacio-eustatic, hydro-isostatic, and glacio-isostatic changes of sea-level in a self-consistent model. Of the six distinct response zones which they identified worldwide, three occur in Canada. Glaciated areas which emerged monotonically constitute Zone I, and ice-marginal areas which submerged monotonically constitute Zone II. Sea-level change on the British Columbia coast behaves as the transition between Zone I and Zone II, with post-glacial emergence followed by submergence. The transitional Zone I/II response owes its existence to the migration of the forebulge in the wake of ice retreat (Clark *et al.* ,1978; Quinlan and Beaumont, 1981). The model of Clark *et al.* (1978) does not take into account local tectonic forces which might confound the late-Holocene signal in areas such as the Pacific Northwest (see Riddihough, 1982; Adams, 1984).

Factors Controlling Relative Sea-level Change

Eustatic Sea-level Change

Eustatic sea-level change is a term that has undergone considerable evolution of meaning. Eustasy was originally defined, as "worldwide and simultaneous" changes of sea-level, during a time when it was believed that the crust was more or less rigid and the geoid possessed an unvarying shape. It followed from this definition that a vertical change in sea-level in one part of the globe could be recorded at the same geodetic elevation elsewhere and that any local variation could be attributed to local crustal movements or basin infilling. With the recognition of hydro-isostasy and changes in geoidal relief, this strict definition of eustasy was no longer tenable (Mörner, 1987; and references therein).

Thus, broader definitions of eustasy evolved. For instance, Mörner (1987) proposed that eustasy should imply simply "any ocean-level changes", a definition under which all processes of relative sea-level change are subsumed. He has divided these processes into three categories: i) tectono-eustasy, ii) glacial-eustasy, and iii) geoidal-eustasy. Tectono-eustasy subsumes plate-tectonic crustal movements, sediment in-fill processes, and various isostatic processes. Geoidal eustasy subsumes processes that cause a deflection of the plumb-line, such as gravitational waves, changes of the earth's tilt, and changes of the earth's rate of rotation. Possibly closer to the original meaning of eustasy is the definition used by Peltier (1987) in which eustatic changes in sea-level are those due to "a change in the volume of the water in the global ocean or in some local part of it" (p 57). These eustatic changes are the result of changes in ocean volume predominantly in response to climate change. Peltier sees eustatic and isostatic changes (discussed later) as being completely distinct.

Used loosely, eustasy refers to ocean-wide changes of sea-level, and all authors agree that ... "nowadays, each region must define its own relative sea-level from which the crustal components are - if possible - later subtracted so that the regional eustatic changes are established" (Mörner, 1987; p. 341). However, there still exists a "Holocene eustatic sea-level problem" (Mörner, 1971). From the glacial sea-level lowstand 18,000 years ago, sea-levels rose dramatically in response to ice melt, and by 6,000 years ago sea-levels were close to present. This is known as the Flandrian Transgression (Fairbridge, 1983). The problem is whether Holocene eustatic sea-levels rose steadily as shown by Shepard (1963), or whether sea-level oscillations were superimposed on the rising trend, with periods when levels may have been higher than today as indicated by Fairbridge (1961). "As the Holocene climate was oscillating and sea-level oscillations are well recorded in all areas studied in sufficient details, I think a low amplitude oscillating eustatic curve sooner or later will be generally accepted" (Mörner, 1971; p 699). Reference to Chappell (1987) and Shennan (1987) indicates that this debate is ongoing.

Isostatic Sea-level Change

Isostatic changes are defined as changes in the equilibrium condition of the geoid (sea-level) in response to changes of load on the earth's surface. Changes of load are due to the growth or decay of ice sheets and to changes in water depth. These forces are coupled so that any loss of ice volume on land will be accompanied by the redistribution of load in the oceans (Walcott, 1972, Clark *et al.*, 1978). Thus, in glaciated areas crustal depression is proportional to ice thickness, but in ice-marginal zones, or beyond, isostatic response is a function of coupled glacio- and hydro-isostatic forces.

Isostatic adjustments are due to both lithospheric flexure and the lateral migration of viscous mantle material. Stresses caused by the ice load cause the lateral extrusion of mantle material away from the load and elastic upward bending of the crust in the periphery (Walcott, 1970 & 1972; Peltier, 1981). This peripheral bulge has been termed the forebulge. The magnitude and timing of glacio-isostatic relaxation at a site is a function of ice thickness and position relative to the center of ice accumulation. Areas experiencing depression will rebound (Zone I), and areas on the forebulge beyond the margin will subside as the forebulge collapses (Zone II). A more complicated signal may develop as the forebulge migrates in the wake of the retreating ice mass (Zone I/II transition; Clark *et al.*, 1978).

Rebound of areas near the ice accumulation center seems to be delayed by the return of laterally extruded mantle material. For instance, in the Hudson's Bay lowland a significant amount of uplift is still expected to occur, as indicated by negative gravity anomalies (Walcott, 1970; Wu and Peltier, 1983). In contrast, areas near the ice margin, such as along the coast of southern British Columbia (Clague, 1983), experienced rapid post-glacial uplift. Workers studying post-glacial sea-level change in the Pacific Northwest (Mathews *et al.*, 1970; Clague, 1975 and 1985; Fulton and Walcott, 1975; Andrews and Retherford, 1978; Clague *et al.*, 1982; Thorsen, 1989; Williams and Roberts, 1989) suggest that glacio-isostatic rebound has been negligible since the mid-Holocene. "On the west-coast uplift was largely completed by 8 Ka and subsequent submergence is generally ascribed to eustatic sea-level rise" (Andrews, 1989; p 557).

Contemporary Tectonics and Sea-level Change in Cascadia

Tectonic vertical movements of the Cascadia subduction zone are related to the structural behavior of the crust during deformation. Contemporary movements on the Cascadia subduction zone are fundamental to a discussion of relative sea-level change because the subduction of the Pacific plate beneath the America plate causes local changes that are superimposed on the glacio-isostatic and glacio-eustatic patterns discussed previously (Adams, 1984; Riddihough, 1982).

In the Pacific Northwest there is the added potential for megathrust earthquakes (Rogers, 1988). Megathrust earthquakes are shallow earthquakes that occur at the interface between the subducting slab and the overlying crust. This zone is known as the interface thrust zone (ITZ)(Spence, 1987). For megathrust earthquakes to occur, the ITZ must lock and build up strain, with catastrophic release typically at magnitudes exceeding 8.0 M_W.

If megathrust earthquakes have not occurred along an active subduction zone for an extended period of time the subduction zone may be aseismic, or comprise a seismic gap. The lack of megathrust earthquakes in Cascadia in historic time ... "is one of the most remarkable to be found anywhere in the circum-Pacific belt. ... The best examples of seismically quiescent plate boundaries are ones that have experienced great earthquakes, but that could be considered as otherwise locked" (Heaton and Kanamori, 1984; p 939).

Fitch and Scholz (1971) developed a model for pre-seismic, co-seismic, and post-seismic deformations of the upper plate, based on 100 years of repeated surveys before and after megathrust earthquakes in Japan. During the period of strain accumulation the trenchward portion of the overriding plate is dragged down, causing lithospheric flexure and the formation of a crustal upwarp inland. Still further landward, a downwarp may lead to slight subsidence. Accumulated strain is released elastically during an earthquake, thus co-seismic movements of the overriding plate are opposite in sense to pre-seismic ones. Simple landward tilt occurs in aseismic subduction zones.

The south Puget Sound earthquake of 1949 (7.1 M_S) occurred at depth, and is considered to be the result of the sinking of the slab beneath western Washington (Weaver and Baker, 1988), implying active subduction in that region. Further evidence of subduction along the Juan de Fuca plate is indicated by landward tilting and crustal shortening measured from geodetic networks (Ando and Balazs, 1979; Savage *et al.*, 1981; Reilinger and Adams, 1982; Dragert, 1986; Holdahl *et al.*, 1989; Savage and Lisowski, 1991).

Ando and Balazs (1979), appealing to the model of Fitch and Scholz (1971), suggested that the simple landward tilting measured from geodetic surveys in Washington was evidence that subduction was occurring aseismically. In the absence of megathrust earthquakes, some researchers conclude that a strong case can be made for either mode of subduction and neither is unequivocal (Savage *et al.*, 1981; Heaton and Kanamori, 1984; Holdahl *et al.*, 1989). Debate as to whether subduction of the Juan de Fuca plate is aseismic or seismic, continues to stimulate geophysical research (see Heaton, 1990; Savage and Lisowski, 1991). Slow subduction, 3.5 cm/yr as recently as 1-2 Ma ago, may contribute to a long recurrence interval between megathrust events (Heaton and Kanamori, 1984), if they occur.

Analysis of the contemporary subduction process along the Cascadia margin is difficult for two reasons. First, there is a paucity of Benioff seismicity in the downgoing slab, making it difficult to study the subduction process by the analysis of earthquake focal mechanisms (Rogers, 1988). Second, the Cascadia subduction zone is notable for the youth of its subducting lithosphere, and may not conform to the model developed from the subduction of old lithosphere (Isacks *et al.*, 1968). As the Juan de Fuca ridge system is close to the continental margin, the age of the subducted lithosphere is less than 10 Ma old. In all other subduction zones where young lithosphere is consumed the age of the material is between 10-20 Ma in age (Rogers, 1988). This identifies "the Cascadia arc-subduction complex as an extreme end member in the spectrum of convergent plate margins" (Duncan and Kulm, 1989).

The problem of the mode of subduction and the megathrust potential in Cascadia is further confounded by the possibility of segmentation of the Juan de Fuca plate. In plan view the azimuth of the subduction zone, defined by the convergent margin, trends north-south along the Oregon coast to Puget Sound, where it changes to approximately north 30° west (Figure 1.1). Significantly, the region beneath Puget Sound is the most seismically active in Cascadia (Weaver and Baker, 1988). Some workers suggest that this change in azimuth of the margin affects the geometry of the slab causing it to undulate, or arch (Rogers, 1983; Crosson and Owens, 1987; Weaver and Baker, 1988), and it may be torn. Segmentation of the Juan de Fuca plate beneath Puget Sound introduces the possibility that the areas north and south of this zone may behave independently, and may have experienced different seismic histories.

Another break in the Juan de Fuca plate may occur in northern Oregon (Peterson, 1991). In most of Oregon there is a lack of Benioff seismicity. The volcanic front in southern Oregon abruptly shifts east, and the andesitic volcanic chain is very narrow. The presence of the andesitic line implies that the slab is at a depth of about 100 km (the depth necessary to induce melting), and its narrowness implies the slab is steeply dipping (Guffanti and Weaver, 1988). Since Benioff seismicity is associated with a sinking slab (Spence, 1987), the absence of Benioff seismicity is used to infer that the steeply dipping section may be broken from the main plate. As well, the region oceanward of the andesitic line exhibits a zone of high heat flux (Michaelson and Weaver, 1986), a circumstance that would not be expected if the slab were intact in that area.

The potential for a megathrust earthquake in Cascadia has been discussed by Heaton and Kanamori (1984), Heaton and Hartzell (1987), and Rogers (1988). Their ideas follow from the recognition of the unique characteristics of youthful subduction zones (Ruff and Kanamori, 1980).

Heaton and Kanamori (1984) looked at the characteristics of youthful subduction zones that have experienced large subduction earthquakes. As the Cascadia subduction zone possesses all those characteristics (shallow trench, lack of back-arc basin, shallow Benioff seismicity, and seismic quiescence on the ITZ) they conclude that the potential exists in Cascadia for a large subduction earthquake. Conversely, they argue that the lack of these features makes a case against the traditional, Marianas type, aseismic subduction.

Yet, other mechanisms, besides the voluntary subduction and rollback of old, dense lithosphere, may account for aseismicity on the ITZ. Kanamori and Astiz (1985) suggest that as the subducting plate approaches zero in age, the high temperature of the plate may cause partial-melt lubrication and, thus, aseismic slip. It is not known at what age this mechanism would become active (Rogers, 1988). Further, the high volume of sediment off the continental margin might give rise to aseismicity by lubricating the ITZ (Duncan and Kulm, 1989). Note that youth and subsequent high buoyancy are usually related to strong coupling and seismicity (Heaton and Kanamori, 1984).

Rogers (1988) notes that of the 7 subduction zones where lithosphere younger than 20 Ma is being subducted, 5 have experienced large subduction earthquakes. Cascadia and the southern Chile subduction zone have not. He feels that the plate most similar to the Juan de Fuca is the Rivera plate subducting beneath Jalisco, Mexico (Eissler and McNally, 1984). This system, like Cascadia, is isolated and not attached to an older plate system. The lithosphere is about 10 Ma in age and its convergence rate is 25 mm/yr. These parameters are very similar to those of the Cascadia subduction zone. The Jalisco earthquake of 1932 ruptured along a 300 km interface with a magnitude of 8.2 M_W. The zone is now seismically quiescent.

Heaton and Hartzell (1986) and Rogers (1988) consider the potential magnitude and recurrence time of a large subduction earthquake in Cascadia. Their consensus is that a magnitude ~8.0 M_W or greater event will occur, with a recurrence time quite easily exceeding 500 years. The magnitude of such an earthquake is proportional to the area of the coupled zone, and therefore the smaller magnitudes (~8.0 M_W) would be expected if the plate ruptured in segments (Rogers, 1988). Adams (1984, 1990), considering the return frequency of turbidite sequences on the continental shelf near the the mouth of the Columbia River, suggests that the return frequency of large earthquakes is no less than 400 years.

Geological Impacts of Megathrust Earthquakes

A few examples of co-seismic deformation in areas that have experienced megathrust earthquakes are presented to develop expectations for the geologic record in Cascadia. Co-seismic deformations are both horizontal and vertical, and can involve simply broad warping of the crust or surface faulting. In the two most familiar major subduction earthquakes, the 1960 Chilean earthquake (M_S ~ 8.5) and the 1964 Alaska earthquake (M_S ~ 8.4), vertical deformations were very dramatic, and both produced destructive seismic sea waves, or tsunamis (Plafker, 1972).

The Alaskan earthquake of 1964 was accompanied by a broad band of deformation that extended for about 800 km parallel to the trench, and was about 300 km wide (Plafker, 1972). Significant surface faulting was found at two localities on Montague Island in the zone of highest uplift. The longer fault was traceable for 16 km, while the shorter one was traceable for 5 km. Maximum vertical displacement on the faults was about 2.5 m on the long one and 5 m on the short one. Maximum observed uplift was 10 m on land and 15 m offshore, while maximum subsidence observed was about 10 m. A suite of marine terraces rising to 40 m asl on Middleton Island was attributed to megathrust earthquakes with a return period on the order of 300-700 years (Plafker and Rubin, 1978). Large deformation was also associated with the Yakutat Bay earthquake of 1899 (Tarr and Martin, 1912).

The Chilean earthquake of 1960 was similarly accompanied by a broad band of deformation parallel to the margin that was about 200 km wide and 1000 km long (Plafker, 1972). Deformation occurred as a large warp with uplift on the seaward half of the affected zone and subsidence on the landward half. No surface faulting was observed. Maximum observed uplift was 5.7 m, while maximum observed subsidence was 2.7 m.

Imamura (1930) reported 26 earthquakes in Japan for the period 684 to 1927 AD from which topographical changes were observed. Some of these changes involved surface rupture, but not all. For 12 of these events, going back to 1891, precise levelings were made before and after the events. The largest vertical changes occurred during the 1923, Kwanto earthquake which ruptured over a length of about 100 km. Maximum uplift recorded was 2 m, while maximum subsidence was 1.6 m. In the great Kwanto quake surface faulting occurred.

West and McCrumb (1988) compiled surface deformation data from a number of megathrust earthquakes. As a general rule of thumb, uplift occurs in the zone 20-100 km from the trench and subsidence occurs 160-300 km from the trench. Owing to the uniqueness of each subduction zone, in the interval 100-160 km from the trench, either uplift or subsidence may occur.

These data indicate that the patterns of deformation observed in Chile and Alaska might be replicated during a megathrust earthquake in Cascadia. Yet, we have no direct geophysical methods of delineating where the bands of subsidence and uplift might be situated. As a predictive tool, Dragert and Rogers (1988) overlaid the pattern of deformation of the 1960 Chilean earthquake on the Cascadia subduction zone. This model predicts that the west-coast of Vancouver Island would experience co-seismic uplift, while the outer coast of Washington would experience subsidence.

Thus, in the geological record, in areas of low lying coastal topography, we can expect to find submerged intertidal or terrestrial deposits in areas of subsidence, and exposed intertidal or subtidal platforms and deposits in areas of uplift. Submerged intertidal marshes and floodplain soils were recorded after the 1960 Chilean earthquake (Wright and Mella, 1963) and the 1964 Alaskan earthquake (Plafker, 1965). Holocene marine terraces are commonly attributed to co-seismic uplift (Lajoie, 1987; Plafker and Rubin, 1978), and have been used to estimate recurrence times of major earthquakes (Matsuda *et al.*, 1978; Plafker and Rubin, 1978).

Yet, Plafker (1990) also notes that some coastal areas subject to co-seismic uplift experience significant inter-seismic subsidence such that in the long term no net uplift is observed. "For example, at Montague Island inter-seismic tectonic subsidence plus eustatic sea-level rise is almost equal to large increments of co-seismic uplift so that strandline elevations show less than a metre of net change over a period of about 4,500 years (Plafker, 1990; p 253)."

Washington-Oregon

Adams (1984, 1990) suggested that turbidite sequences observed off the mouth of the Columbia River were triggered by earthquakes, possibly megathrust earthquakes, and related the chronology of these deposits to sequences of buried peats and estuarine muds in five estuaries along the outer coast of the Olympic Peninsula (Atwater, 1987). These estuarine sequences contain between one to six buried peat layers. From a sharp contact at the top of a peat, grey mud grades upward into mottled mud, and then into the succeeding peat. Capping some of the peats are beds of sand, which become thinner and finer grained landward, indicating a bayward source for the sand.

Peterson and Darienzo (1988) listed six bays in Oregon where similar peat sequences are found. Netarts Bay, in northern Oregon, has the longest and most unambiguous record. The Netarts bay record goes back 3,000 years, and at least 7 distinct marsh horizons are apparent. Peterson and Darienzo also describe sand-rich capping layers at the sharp contact between the peats and the overlying mud. These sand-rich layers are traceable across the marsh system and are massive in structure.

Atwater (1987) and Peterson and Darienzo (1991) argued that the sharp contact between the peats and the overlying sediment indicates rapid subsidence of these marsh areas, and any process of eustatic submergence would be too steady to account for the abrupt transition. Further, Atwater has found, in section, zones where the sand layer is disrupted by fossil grass (*Deschampsia cespitosa*) clumps in growth position, indicating that the sand was draped over the vegetation during deposition (Atwater and Yamaguchi, 1991). Similar post-seismic burial of plants was described in Portage Inlet, Alaska, following the 1964 earthquake (Bartsch-Winkler *et al.*, 1982). These workers feel that the only possible depositional mechanism for such an abrupt stratigraphic change is co-seismic subsidence, followed by seiche effects or tsunami.

All possible alternative hypotheses that could have led to the episodic and cyclic coastal submergence documented by Atwater were examined by Donald West (Golder

Associates, 1988). He concluded that of the plausible non-tectonic alternatives (sediment compaction, etc.), the late-Holocene glacio-eustatic fluctuations proposed by Fairbridge (1961) and Fairbridge and Hillaire-Marcel (1978) are a viable alternative to the co-seismic model. Fairbridge and Hillaire-Marcel stated that ... " it is clear that these sudden drops or rises of sea-level of the order of 2-5 m are world-wide in effect and may occur within timespans of 100-200 years. These negative sea level oscillations correlate with documented climatic cooling and are assumed to be glacio-eustatic." (p 413).

West predicted that if these eustatic fluctuations are responsible for the observed pattern, then i) there should be a similar number of eustatic rise events and buried soils for a given period of time, ii) their recurrence times should be similar, and iii) the death of the marsh should coincide with the initiation of a period of eustatic rise. West plotted the available radiocarbon dates of the marsh horizons studied by Atwater against the Fairbridge and Hillaire-Marcel eustatic curve (Fairbridge and Hillaire-Marcel, 1978; their fig. 2). He found that there were 8 observable marsh surfaces versus 10 documented eustatic rise events, that 7 of the 8 marsh horizons were correlated with glacio-eustatic fluctuations, and, by extension, there were similar recurrence times.

Finally, it is interesting to note that Plafker (1990) has no trouble attributing flights of late-Holocene coastal terraces to a co-seismic origin, but he states, regarding 7 submerged organic layers as much as 4,500 years old and to a depth of 12 m in the Cook Inlet region, that ... "it is not clear how many of these organic layers are earthquake related" (p 255), implying that some may have a eustatic origin.

British Columbia

The search for geological evidence for earthquakes in British Columbia is part of the mandate of the Geological Survey of Canada and the Provincial Survey (Clague, 1989; Clague *et al.*, 1989; Bobrowsky and Clague, 1990; Clague and Bobrowsky, 1990). This work has concentrated on the highly populated region of southern Georgia Strait.

Clague (1989) and Clague and Bobrowsky (1990) report stratigraphic evidence from 14 localities on the southern tip of Vancouver Island and adjacent Fraser delta. All the sites give evidence for late-Holocene relative sea-level rise. At the westernmost site, Muir Creek, there may be evidence for one or more transgressions. The latest transgression, documented at a number of sites, is represented by data from Island View Beach. At Island View Beach, on the Saanich peninsula, in-situ fossil stumps in the intertidal zone have been dated at 2,040 yr BP. A peat horizon buried by intertidal muds has been dated at ca 2,500 yr BP. Clague and Bobrowsky (1990) comment: ... "sea-level, however, has not fluctuated more than 1 m from its present position during the last 2,000 years, and there is no evidence during this period for regional co-seismic subsidence or uplift which might be expected during great subduction earthquakes on the Cascadia subduction zone" (p 245).

Clague *et al.* (1989) found evidence, in four of five surveyed Vancouver Island lakes, of subaqueous landsliding possibly induced by the Vancouver Island earthquake of 1946. Bobrowsky and Clague (1990) took a series of vibracore samples from Saanich Inlet. The cores record varved sediments periodically interrupted by allochthonous turbidite beds. The return frequency of these turbidite beds is on the order of 100 years. They are attributed to earthquakes, but not necessarily megathrust earthquakes.

Tsunamis: Their Occurrence and Impact on the Outer-coast

Tsunamis are ocean waves generated by sudden sea-floor displacement during major earthquakes. They are mentioned here because they are commonly associated with megathrust earthquakes, they may have had an influence on littoral features, and are possibly evident in the geological record.

The following historical example from Japan (Imamura, 1930) shows they can be very common events in susceptible areas:

> "As generally happens, a sudden retreat of the sea pointed to a pre-seismic upheaval that was underway, and which an earthquake soon followed. About 150 metres off the shore of Hamada-ura is the islet of Turu-sima, to the foot of which it is said the water withdrew and left a dry connexion with the mainland. Fisherman were soon busy gathering live fish that were left stranded on the newly formed land and collecting earshells from the foot of the islet. Hardly ten or twenty minutes had passed when one of the men, an old seasoned fisherman, shouted excitedly to his comrades urging them to make at once for the hills to escape the tsunami, which experience had taught him to expect from phenomena such as he had just witnessed. Before long a tremendous shock was felt, and ten minutes after it the earthquake wave came rolling on, which in this instance was not at all destructive" (p 38).

Compare the above with a Pacific Northwest oral tradition. In 1868 Sproat published a Clayoquot story of a great ebb and flow of the sea (Sproat, 1868):

> "The tide ebbed away from the shores of [Barkley] Sound and left it dry, and the sea itself retreated a long distance. This continued for four days, and the Sheshaht made light of the occurrence. There was one, however, Wispohahp, who, with his two brothers, did not do so. After a mature consideration of the circumstance, he thought it likely that this ebb would be succeeded by a floodtide of corresponding height and power. Accordingly, he and his brothers spent three days in the forest collecting material for a rope of cedar inner bark, which, when made, was so large as to fill four boxes. There was a rock near the village, from the base of which sprang a group of bushes... Round these bushes Wispohahp fastened one end of his rope, attaching the other to his canoe... After four days the tide began to flow, and crept up slowly to a point about halfway between the point of its furthest ebb and the houses. At this point its pace was suddenly quickened, and it rushed up at fearful speed... They were all soon caught by the rising water; and while Wispohahp rode safely at anchor, the Sheshaht, unable to resist its force, drifted in their canoes to distant parts. Finally the water covered the whole country, except Quossakt, a high mountain near Toquaht... At the end of four days the floodtide began to abate" (p124).

This same story was recorded by Swan in 1868 from the Indians of Cape Flattery (quoted in Heaton and Snavely, 1985). Before the event Cape Flattery was an island separated from the peninsula by a narrow channel then:

"The water suddenly retreated leaving Neeah Bay perfectly dry. It was four days reaching its lowest ebb, and then rose again without any waves or breakers, till it had submerged the Cape, and in fact the whole country, excepting the tops of the mountains at Clyoquot. The water on its rise became very warm, and as it came up to the houses, those who had canoes put their effects into them, and floated off into the current, which set very strongly to the north. ... and when the waters assumed their accustomed level [four days later], a portion of the tribe found themselves beyond Nootka, where their descendants now reside" (p. 1456).

Heaton and Snavely (1985) question the details of height and timing in this account, which may have become exaggerated in the oral tradition, but, regardless of the native concept of height and time, the event behaved just like a tsunami. For comparison, the 1964 tsunami, as described by a resident of Tofino, was preceded by a large ebb, greater than any he had seen before. It was so striking, he immediately attributed it to a tsunami. Shortly thereafter the water rose so high (on Chesterman Beach) that it threatened the foundations of his house (Neil Buckle, 1990; pers. comm.).

A catalogue of Chilean earthquakes (Lomnitz, 1956) larger than 7.5 magnitude for the period 1535-1955 AD contains 40 major earthquakes from which 24 tsunamis were observed and recorded in the region. Only two of these tsunamis were described as destructive. The data suggest that not all major earthquakes generate tsunamis, and that the impact of these tsunamis can vary greatly.

Wigen (1979) assembled a frequency magnitude relation for all historical tsunamis recorded at Tofino and Port Alberni. The historical record goes back to 1906 and includes 33 events. The tsunami generated by the 1964 Alaskan earthquake has a 200 year return frequency and the 1960 Chilean earthquake induced tsunami has a 100 year return. It is apparent that significantly large tsunamis occur with regular frequency in our area.

White (1966) surveyed the effects in Canada of the 1964, Alaskan earthquake. The most severe damage was caused by the ensuing tsunami. The areas most greatly affected were the sounds and inlets on the west coast of Vancouver Island. At Tofino the wave had an amplitude of 2.40 m, whereas, due to amplification in the Alberni Canal, at Port Alberni the wave reached at least 8 m (Wigen, 1979). The true height of the wave at Port Alberni could not be measured because it exceeded the range of the tidal gauge.

Tsunami damage on the west coast of Vancouver Island from the 1964 Alaska earthquake was experienced in San Josef Bay (north of Quatsino Sound), Quatsino Sound, Kyuquot Sound, Nootka Sound, Hot Springs Cove, Barkley Sound, and Alberni Inlet. In many places buildings and logging booms were damaged. In Tahsis Narrows, at the head of Nootka Sound, bottom sediments were disturbed, driving bottom fish to the surface. In San Josef Bay trees were swept from the banks of the San Josef River and significant shifting of bar sands was observed. On one small creek sand was stripped from a clam bed leaving it denuded.

Results of modelling a hypothetical megathrust earthquake on the Explorer plate were published by Ng *et al.*(1990). The tsunami wave reached the west-coast of Vancouver Island within 30 minutes and Victoria in two hours. The amplitudes were not much different than for the 1964 tsunami, but it was shown that the event could alter tidal amplitudes over a full tidal cycle.

From this evidence it is clear that tsunamis have had an impact in the region historically, and possibly prehistorically, but it is unclear whether they have left a lasting signal in the geomorphic record of the littoral zone along the outer coast of British Columbia.

CHAPTER 4: Research Methodology

General Methodology

Construction of the Sea-level Curve

The work of sea-level curve construction requires the identification of former water levels, measuring the elevation of those paleo-levels, and determining when they were occupied by the sea. The methods used in this study consisted of a five step procedure: i) the identification of paleo-littoral features by airphoto analysis and ground reconnaissance; ii) topographic surveys of modern and paleo-littoral landforms; iii) stratigraphic description of paleo-littoral deposits by means of exposures, cores, and pits; iv) radiocarbon dating of wood, shell, and peat of known stratigraphic association; and v) laboratory analysis of sediments and fossil assemblages.

Error and uncertainty are intrinsic to this process. A good discussion of shoreline data and the uncertainty associated with building a sea-level curve is given by Andrews (1989; & references therein). In brief, the factors leading to uncertainty are i) the precision of the survey, ii) the relation of the watermark to datum, with the potential for change in the relation through time, and iii) ... "the clarity of the association between the material being dated and the paleo-sea-level" (ibid). For instance, regarding age, logline wood may be much older than the age of the shoreline, or, regarding elevation, the in-situ shells of molluscs, which might date a landform more accurately, cannot necessarily be tied to a specific elevation. Finally, iv) samples for radiocarbon dating may be contaminated, and radiocarbon ages can vary with the laboratory method and thus the laboratory used. For useful results all assumptions, estimates, and associations must be made explicit - advice which is adhered to in this research.

Littoral Features and the Nomenclature Used

Fairbridge (1968) describes the *littoral zone* as extending from the lower limit of wave influence to the front of the *littoral* sand dunes, beach ridges, or cliff line. On a typical temperate beach there is the beach face, or swash zone, then a band of logs, and finally the forest edge. In this thesis, the seaward limit of log distribution is referred to as the ***logline***. The ***vegetation edge*** refers to the seaward extent of plant community establishment. On periodically eroded or retreating shorelines the vegetation edge will be marked by a vertical *scarp*, or possibly a small cliff. On beaches with foredunes this scarp is referred to as the *foredune cut* (Bird, 1976). The swale behind the first foredune is referred to as the *foredune swale* (see Figure 4.1).

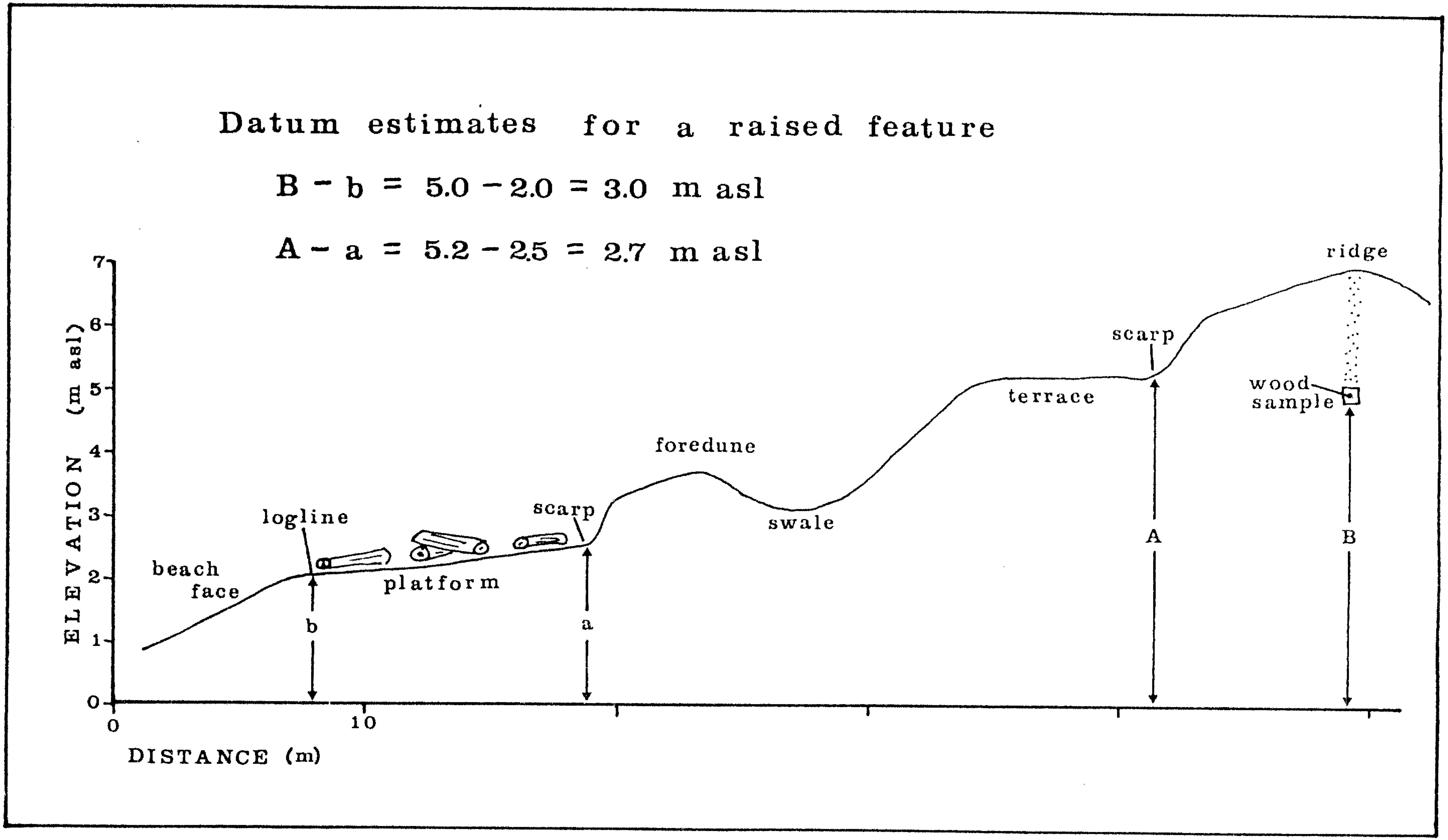


Figure 4.1: Hypothetical beach and back-beach profiles.

The most prominent landform on Vargas Island is a sequence of parallel sand ridges (Figure 2.1). Morphologically, these ridges could be either *cheniers* or *beach ridges*. Cheniers have been formally defined as beach ridges resting on silty or clayey deposits. The progradation of a mudflat in a deltaic environment, periodically interrupted by chenier formation, leads to the development of a *chenier plain* (Augustinus, 1989). The Vargas Island ridges are not developed on a progradational mudflat, they represent the progradation of back-beach foredunes onto a wave-cut platform and are termed beach ridges (Bird, 1976).

Paleo-littoral features commonly take the form of terraces with the landward side often marked by a scarp. The base of this scarp represents the extreme high tide shoreline of the raised feature, and is called the *strandline* (Fairbridge, 1968). Marine terraces can be either constructional or erosional. Erosional terraces are referred to as *wave-cut terraces*, a term equivalent to the Scandinavian *strandflat*, although it seems that this term has been applied to more extensive physiographic features such as the *Milbanke Strandflat* (Holland, 1976). For example, strandflat, as used by Forman (1990), refers to the whole of the coastline below the late-Weichselian marine limit, and so includes a sequence of strandlines below it. Wave-cut terraces are the raised versions of contemporary *wave-cut platforms* (Fairbridge, 1968). Where a broad terrace did not form, or was subsequently eroded away, the previous shoreline position may be indicated by only a narrow *step* with a small scarp.

Field Methods

Identification of Paleo-Littoral Landforms

1:16,000 scale airphotos (series BC84084; 68-73; 79-82; &130) were used to identify the main landforms and prepare a working base map of the southern half of Vargas Island. 1:20,000 scale airphotos (series BC81074, 252-254; & series BC81075, 21-24) gave complete coverage of the island and were used for orientation while ground truthing. Ground reconnaissance was conducted in September and April of 1990, and throughout the field season from May to August, 1990. Reconnaissance involved exploring all creeks and back-beach environments for interesting stratigraphic exposures and marine landforms. The extensive low elevation bog of the southern half of Vargas Island (Figure 2.1) was fully explored to investigate the character and perimeter of the beach ridge system and some monadnocks, and to properly map the drainage of this extremely level and sluggishly drained surface. Other areas explored were the creeks and back-beach areas on nearby islands.

Topographic Surveys

Topographic surveys, tied-in to most recent high tidelines, and running inland from the beach are the main investigative tool used in this research. These surveys establish the profile of modern and raised littoral features, as well as the elevations of important stratigraphic details in these landforms. Two instruments were used, the Sokisha BT-20 transit and the Wild NK-1 level. See Appendix 1, Table A1.1 for a list of the instrument used, and the nature of, a particular survey.

The surveying was conducted by traversing, using the tachymeter for distance measurements, and aligning the scope to magnetic north. On the Sokisha this could be done using the built-in compass, but when aligning the NK-1 a Silva Ranger was held against the instrument. Holding the Silva so close to the NK-1 caused a constant deflection of the compass needle 7° to the west. This deflection was corrected when plotting. These methods yield imprecise bearing measurements, but horizontal control was not an important concern.

Great care was taken to maintain vertical control. On the two longest surveys, the Ridge Traverse and the South Bog/Medallion Traverse, both just over 2 km in length, benchmarks were established roughly every 70 m on the leg out from the beach and were closed on the return leg. Thus, a traverse consists of a series of closed loops, allowing frequent error checks along its length. Large closure errors or blunders subsequently noted while reducing the field data could then be easily resurveyed. Field data was reduced by the rise and fall method (Clark, 1972).

Due to the thick bush and spongy ground, and the use of the transit as a level, a 0.05 m closure error was tolerated. Closure errors larger than 0.05 m were resurveyed. The benchmarks consist of 0.05 m by 0.05 m cedar stakes about 0.70 m in length driven into the substrate with a sledge hammer. The stadia rod was rested on a nail driven horizontally into the stake at ground level. These stakes are numbered and have been left in position for future reference. Appendix 1, Tables A1.2 a & b list the elevation and distance inland of benchmarks on the two main traverses. All shorter traverses of raised marine landforms were closed. Some short profiles of modern back-beach topography were left open due to time considerations, yet the error is believed to be negligible as the NK-1 was used, and these profiles had few turning points.

To provide accurate representation of features all slope breaks and ridge crests were surveyed. Data from the beach-face included position of the tideline, the logline, and the vegetation edge. When measuring the elevation of a logline, vegetation edge, or foredune swale the stadia was held on sand at that position. Scarps are measured at the base, or strandline. When the vegetation edge and a scarp coincide, the elevation refers to the scarp.

The elevation of raised feature refers to the strandline, as that point is constrained by a modern equivalent. There are, however, problems with measuring the exact elevation of the strandline due to collapse of the scarp, and the development of a thick forest floor, or peat infil.

The Datum Estimate

The datum used in this research is mean sea-level, and all elevations expressed are relative to metres above mean sea-level (m asl) at Tofino. Most sea-level indicators (high marsh, loglines, tidelines) relate to the high tide level, which is a function of tidal range and varies regionally. Thus, these indicators must be corrected for the local tidal amplitude so that regional correlations of mean sea-level can be made (Clague *et al.*, 1982).

The tidal range on Vargas Island was assumed to be identical to that of Tofino, the nearest reference port. Comparison of the two outer-coast reference ports in Clayoquot Sound, Riley Cove on the north side of Flores Island, and Tofino, show this to be an acceptable assumption, as the critical measure, the difference between the higher-high large tide and mean water, is similar in both places (Table 3.1). In contrast, at Kennedy Cove, up Tofino Inlet at the mouth of the Kennedy River, this difference is 0.30 m larger (For site locations see figure 1.2).

Table 4.1: Tidal data for Tofino and secondary ports in Clayoquot Sound

	Elevation (m asl)		
	Tofino	Riley Cove	Kennedy Cove
Higher-high large tide (HT)	4.11	4.15	4.79
Lower-low large tide	-0.03	-0.18	0.0
Mean water level (MW)	2.13	2.13	2.50
HT - MW	1.98	2.02	2.29

Source: Canadian Tide and Current Tables 6, 1990.

Change in the tidal amplitude through the Holocene must also be considered. For example, in the Bay of Fundy a relatively small change of sea-level in the Holocene was reflected in a much larger change in the tidal range, induced by significant changes in natural period of the bay (Grant, 1970). In this study, it is assumed that Clayoquot Sound, with its deep, narrow fjords has had a constant tidal range throughout the Holocene.

In this research the water plane to which topographic surveys were tied was the most recent high tideline, with the assigned elevation being the Canadian Tide and Current

Tables, 1990 prediction for the appropriate tide, using Tofino as the reference station. Two obvious sources of error associated with this approach had to be evaluated. These were i) the difference between the predicted and observed tidal elevations, and ii) the bias induced by swash in the formation of the tideline.

To estimate the error between the predicted and observed tidal elevations, a tide gauge was established in Ahous Lagoon, in a small enclave out of the tidal stream. The lagoon has a very narrow entrance and is guarded by rocks, and in low swell conditions the lagoon makes an effective stilling pond. The gauge consisted of a graduated two-by-four driven vertically into the sand and stabilized by a log projecting out over the pond. This jury-rig was washed away in a November storm. The gauge was read at the time of the predicted high tide on 14 separate occasions spaced through the research period (Appendix 1; Table A1.3). The error associated with reading the gauge was ± 0.03 m, while the standard deviation of the residuals between predicted and observed readings was ± 0.07 m. Thus surveys tied to this gauge have an inherent error of ± 0.10 m.

To estimate the accuracy of tideline estimates, four successive tidelines were tied-in to the lagoon tide gauge, providing a comparison of predicted high water with the resultant tideline (Appendix 1; Table A1.4). Although this is a small sample, the data do indicate a positive bias of 0.02 m up to 0.18 m. On large swell days in the summer, 0.50 m of bias might be expected. Surveys in this research were tied to the tideline on low swell days, and I feel confident of an estimate of at the most 0.25 m positive bias.

Thus, for this study, the precision of estimates of mean water is ±0.10 m, and the positive bias of high tidelines is estimated at 0.0 m to 0.25 m. So the estimate of mean water to which a particular survey profile is tied falls in the bounds -0.10 m asl ≤ mean water ≤ 0.35 m asl.

The Use of Modern Shoreline Analogs

In this study, elevation data from modern shorelines is used to constrain the elevation of former water levels represented by paleo-littoral features. For instance, on a beach where a modern marine scarp and a sequence of strandlines exist, the elevation of the modern scarp is subtracted from that of the strandline to estimate the paleo-sea-level (Figure 4.1). As well as scarps, profiles of modern beach gravels and foredune topography are used as analogs. For the beach ridge system, three modern beaches with developed foredunes were surveyed to provide some comparison, and gain insight into beach formation processes. It is important when using these estimates that they are applied to features with the same general exposure and the same sedimentary characteristics. Of course, analogs are no substitute for biological indicators of tidal elevation, but in the absence of such data analogs are a useful tool. In this research, as many independent estimates of mean water as possible are applied to any one feature, and the consistency of the results is assessed. The elevations used to represent a feature will be the ones with the greatest precision, providing there is consistency among estimates.

Stratigraphic Description

Most stratigraphic descriptions of surficial materials on Vargas Island relied on coring and dug pits. The top of all natural exposures, pits, and cores were tied to the nearest benchmark with the NK-1 level. Depth to contacts was measured with a tape.

Sediments were logged and photographed, and were sampled for laboratory analyses. Sedimentary units were described on the basis of texture, color, oxidation, and the presence of organics. Fine clastic sediments were classified as muds because textural analyses of fine sediments was not conducted. Muddy diamicts were referred to as sandy muds if sand grains were the minor constituent and muddy sands if sands predominated. Gravels were not analyzed, and were simply referred to by the dominant clast size and an appropriate modifier. If the material was peat, then it was described as sandy, muddy, woody, fibrous, or gyttja. Stratigraphic units are described from the stratigraphically oldest to the youngest, or from the bottom up. Figure 4.2 is a master legend for symbols used in Chapter 5.

Sediment Sampling

A total of 17 sampling transects oriented across foredune, swash, and beach face facies were established on Ahous Bay and the two northwestern beaches. Sand samples of approximately 300 grams were collected at 7.5 m intervals along each transect. Samples from each facies were collected in each transect. A total of 200 sand samples were collected. Also, sand samples were collected from 17 individual ridges along the Ridge Traverse. Sediment samples were also collected from every unit defined in all pits and exposures.

Peat sampling was conducted with the Hiller Corer. Reconnaissance coring with the Hiller corer at intervals along South Bog revealed the basin profile and stratigraphy of the bog. The number of cores needed to define the continuity of internal stratigraphy was judgementally decided. Extracted core was bagged in 0.10 m or 0.20 m intervals, and core station and depth below surface was recorded, as well as the woodiness and sand content of the peat. Fifty seven samples from six core sites in the Medallion subbasin were extracted for study. Samples were air dried in the laboratory. I also experimented with lengths of PVC pipe with sharpened, serrated ends, to try to obtain undisturbed core. As

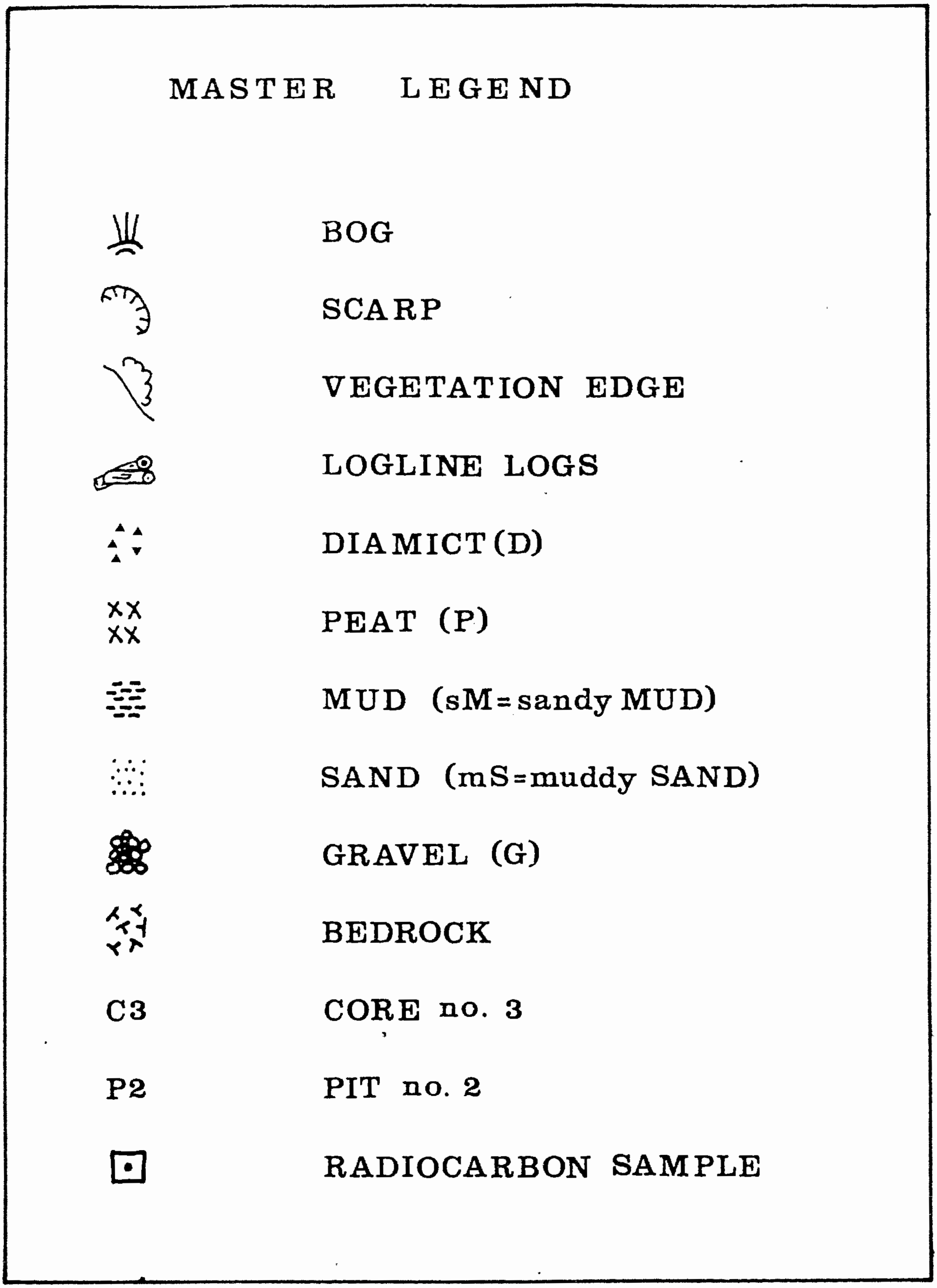


Figure 4.2: Master Legend for symbols used in text

the pipe was gently forced down into the peat, it was twisted to saw rhizomes. Core recovery was unsatisfactory with this method, but two core samples (C-11 and SB-3) were retrieved, sealed, and brought back to the lab undisturbed.

As well as peat samples from core, one peat sample from each of four modern depositional environments were grab-sampled to provide modern analogs for comparison with the fossil assemblages. These were swamp, fen, bog, and intertidal high marsh peat.

Selection of Samples for Radiocarbon Dating

A total of 10 radiocarbon dates were obtained during the 1990 field season; 4 on wood, 3 on shell, and 3 on peat. In addition, three unpublished dates from Rogers *et al.* (1986) are referred to in this study. Materials selected in 1990 for radiocarbon dating were wrapped in foil and transported to the laboratory where they were air dried. The stratigraphic context, and elevation of these samples was noted in detail so that the significance of the date would be clear. Some samples were sent to the radiocarbon laboratory at Simon Fraser University, but the majority of samples were sent to Beta Analytic Inc. in Florida. Shell material being dated was hand selected to clean the pieces and obtain a single-species date. Shell dates have been corrected for an 800 year reservoir effect (Robinson and Thompson, 1978). None of the wood or shell samples were obviously contaminated, and all were of optimum size (Beta Analytic Inc., 1990). Two of the peat samples were from about 1 m depth in South Bog and no obvious contamination was noted, but the peat sample from the Ratcliffe site was immediately below forest litter and may have been contaminated by modern rhizomes or roots of forest plants.

In this study, radiocarbon dates provide minimum or maximum age estimates for the associated littoral features. Logline wood provides a maximum age estimate of a feature as the tree had died before being incorporated into the logline. In-situ shell is more or less coeval with the associated sediments because the organisms were living in the sediment they were recovered from. Midden shell provides a coeval and/or minimum age because the shell fauna was harvested then deposited on the littoral and/or back-beach. Basal peats provide a minimum age because the organic mat became established after the formation of the feature.

Laboratory Methods

Sediment Analysis

Sediment analyses followed guidelines in Folk (1974). Sands from the beach ridge system were heavily oxidized and aggregated, with significant quantities of limonite present. These sands were treated with 10% HCl to dissaggregate samples and remove limonite. Sands from the modern beach, swash zone, and foredune environments were not aggregated and were not chemically treated. All samples were split to between 100-200 grams. Sieving was conducted in the sediment laboratory in Geography at U.B.C. Samples were mechanically shaken for the standard 15 minutes. Texture was analyzed at 1/2 phi intervals. Raw data was entered into the MYSTAT data editor on the Macintosh computer. The mode of each sample was recorded, and skewness was calculated using the built-in program. A total of 120 samples were analyzed.

Shell Fauna Identification

Shells were keyed using Quayle (1970), and Morris (1966). Habitat preferences of shell species were taken from Morris (1966) and Bernard (1970). Taxonomical nomenclature follows Morris (1966). Shell fauna were later sent to the Royal Provincial Museum for cross-checking of identification, and donation to the museum collections.

Analysis of Macrofossils in Peat

Macrofossils are locally derived (see Jacobsen and Bradshaw, 1981), often reflecting the growth of individual plants or possibly very local disturbance events such as windfall. Thus macrofossils are spatially variable and results should generally be interpreted with caution (Birks and Birks, 1980). The standard procedures for macrofossil analysis can be found in Birks and Birks (1980). One hundred milliliter samples are recommended, but sample sizes for this project were limited by the size that could be recovered with the Hiller corer (ca 10-35 ml). Therefore sample sizes were much smaller than ideal. To strengthen the conclusions of these analyses, two cores, C-3 and C-5, were chosen for study.

Air-dried samples were weighed, and soaked in 10% NaOH for a number of days to dissolve the humic acids. Sample volume was estimated by displacement. After soaking, the organics were gently decanted onto screens in an attempt to retain sands in the beakers. The organics were washed gently through the screens under tap water. Three sieve sizes were used in the screening process. The largest was a 2 mm mesh, and the smallest was 0.250 mm mesh. For C-3 a 1mm mesh was used as the intermediate screen, while for C-5 a 0.750 mm mesh was used as the intermediate screen.

Material from individual screens was sorted under a Wild dissecting microscope, and all fungal sclerotia, animal parts, seeds, and recognizable vegetative parts were retained and stored in ethyl alcohol. Seeds were later dried for identification and then placed individually on drops of glycerine for permanent storage.

Identification reference manuals used were Martin and Barkley (1961), Mason (1957), U.S. Dept. of Agric. (1974), and Hitchcock and Cronquist (1973). Use was also made of the reference collection in the paleoecology laboratory at Simon Fraser University.

To prevent double counting of spruce and hemlock needles only those with a base attached were tabulated. Pine needles appeared to preserve poorly and thus two categories, short shoots and needle fragments, were tabulated. The two cedars were not separated in the count. As cedar twigs readily dissaggregate, two categories, twig fragments and leaves, were tabulated. Tabulated spruce and hemlock needle bases, cedar twig fragments, and pine short shoots were used in the needle equivalent sum. Needle counts are presented as percentages of the needle equivalent sum. If no needle equivalent was tabulated, but leaf fragments were present, then the presence of the species is noted with a dot. Seeds are represented as concentrations; number/100 ml was chosen so that infrequent fossils could be represented with whole numbers, although this multiplying-up tends to make the results appear more significant than they are.

Microfossil analysis

Samples from core at SB-3 were sent to Richard Hebda, at the British Columbia Provincial Museum, for microfossil analysis and ecological interpretation.

CHAPTER 5: Results

Introduction

Data from sixteen survey sites are presented in this chapter. These sites include both modern and relict littoral landforms.

Study sites augmented with radiocarbon dates and paleoecological information are termed primary sites. These, the Ridge Traverse, Ahous Lagoon Measured Section, the Medallion/South Bog Traverse, the Buckle Berm site, and the Ratcliffe Terrace, reveal the pattern of sea-level change on Vargas Island. Secondary sites are profiles of raised littoral features that augment the evidence provided by the primary sites. These secondary sites are the Meadow Bay Gravels, the Medallion Strandflat, Miltie's Beach, Kelsemaht Beach, McIntosh Bay, and Dick & Jane's Terrace. The remaining surveys are of modern beaches which provide analog data for estimating previous water levels. For the location of study sites see Figure 5.1.

Elevations of features on the modern beach face are presented and discussed first because these data are used in the interpretation of primary and secondary sites. Data from paleo-marine landforms are presented on a site-by-site basis. Primary sites are discussed in an order which develops the sea-level narrative. Secondary sites are arranged from the higher strandlines (deglacial) to the lower strandlines (Holocene). In Appendix 2, Table A2 ten new radiocarbon dates, and three dates provided by Rogers *et al.* (1986), are presented. These dates are from raised littoral features on Vargas Island and they help constrain relative sea-levels from about 12,000 yr BP onward to present.

The Modern Littoral Zone

Elevations of Features on the Modern Beach Face

Presented in Table 5.1 are elevations of beach log distribution, the vegetation edge/foredune cut, and the foredune swale from modern beaches. These data are referred to throughout the text. Elevations of the vegetation edge, or foredune cut, indicate the extent of the modern littoral zone. On average it extends to about 3 m asl. On the steep cobble beach at Moser Point the littoral extends to 4.1 m asl, while on some sand beaches it may extend only to 2.4 m asl. In extremely sheltered areas, such as the lagoon, the littoral zone would extend only as high as the higher-high tide, or 2 m asl.

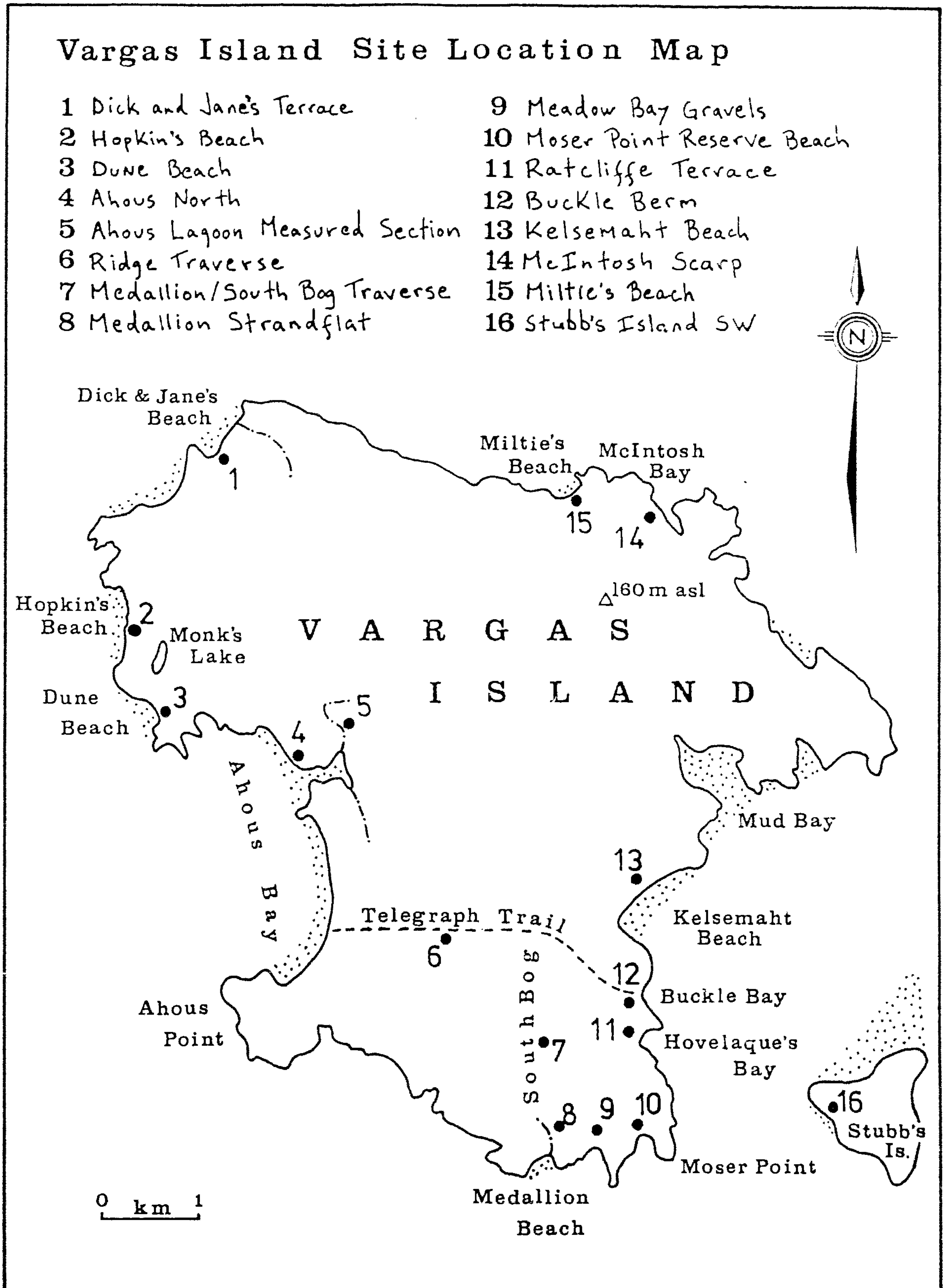


Figure 5.1: Study Site Location Map for sites on Vargas Island.

Table 5.1: Elevations of Features on the Modern Beach Face

Beach	Aspect (°)	Elevations (m asl) Logline	Highest logs	foredune cut veg. edge	foredune swale
Hopkin's Beach	265	2.2	3.7	2.7	3.0
Dune Beach	195	2.3	4.0	2.8	3.7
Ahous North	220	2.2	4.5	3.0	3.7
Ridge Traverse	251	2.0	3.2	2.7	**
Moser Point Beach	144	1.7	4.8	4.1	**
Medallion Beach	130	1.8	3.4	3.1	**
Buckle Bay	073	**	**	2.9	**
Dick & Jane's Terrace	290	2.0	2.6	2.6	**
Hovelaque's Beach	102	1.7	3.3	2.8	**
Kelsemaht Beach	122	2.2	3.1	2.6	**
Miltie's Beach	304	1.9	3.0	2.7	**
McIntosh Bay	042	**	**	2.4	**
Stubbs Island SW	220	2.1	3.2	2.9	**

**: indicates feature altered or absent.

Elevation Distribution of Logline Wood

Wood recovered from relict features is often assumed to represent the logline, and workers generally subtract the elevation of higher-high tide to arrive at a *maximum* estimate of mean sea-level at the time of deposition. The data in table 5.1 show that the seaward edge of the logline is consistent with the higher-high large tide (~2.0 m asl), but that logs can be found significantly higher on the modern beach, the upper limit being governed by the steepness of the beach, and the height of back-beach topography. Thus, if a log was deposited at the seaward edge of the logline then it would provide an accurate estimate of mean sea-level at the time it was deposited, but if the assumption is false, and the wood was deposited above the logline, then an over estimate of mean sea-level will result.

Due to the ability of storm waves to toss materials, the precision of a logline estimate of paleo-datum will be better on sand beaches with low back-beach relief than those from beaches that were steeper, and/or had greater back-beach relief. For example, compare the elevation range of log distribution on a sand beach with low back-beach relief (i.e., Medallion Beach; ca. 1.6 m range), to that of a steep gravel beach (i.e., Moser Point; ca. 3.1 m range). This discussion assumes logs are not deposited below the higher-high tide, as occasionally happens when logs are waterlogged. In general, compared to other methods, logline wood is a poor estimator of previous water levels, and other independent estimates should be sought, or an effort should be made to determine the position of the log on the paleo-beach.

Primary Sites

The Ridge System

Description

In May, 1990, the Ridge System was precisely levelled, from Ahous Bay, 2 km inland along the Telegraph Trail (Figure 5.1). This transect is referred to as the Ridge Traverse (Figure 5.2). A series of shorter transects were also carried inland a short distance from the Kelsemaht beach. On the Ridge Traverse the profile rises gradually from Ahous Beach, for 300 m, to a slope break at 4 m asl. Then, over a distance of about 50 m the profile rises to 5.5 m asl. Inland a further 400 m the 5.5 m asl swale elevation increases to 6.5 m asl. This 6.5 m asl elevation continues right across the island to the paleo-tombolo, and beyond. The profile on the Kelsemaht side is similar but much shorter. The main body of ridges thus forms a terrace about 2.5 km wide with an average swale elevation of 6.5 m asl. Relief of individual ridges varies from 0.5 m to 1.5 m.

Rogers *et al.* (1986) drilled a series of 11 auger holes along the Ridge Traverse. The auger was able to reach 2 m depth in most cases. Generally, about 0.25 m of peat overlay medium sands. There was no obvious stratigraphy in the sands and only in their easternmost hole did they contact bedrock. Structure could not be described from the auger holes. Three samples were taken for radiocarbon dating. Just east of BM-8, 1,320 m inland from Ahous Beach, a log resting on sand at the base of peat, gave a radiocarbon age of 4,860 ± 90 (Beta-18336). Just east of BM-6, 1,480 m inland from the beach, a basal peat gave a radiocarbon age of 510 ± 80 (Beta-17517). Between BM-A and BM-B, 1,975 m inland from the beach, a log in sand gave a radiocarbon age of 4,030 ± 100 (Beta-17516).

The samples Beta-18336 and Beta-17516 are inconsistent with one another, because, logically, the younger date should be farther seaward. This suggests that Beta-18336 was derived from reworked material, or that Beta-17516 is contaminated, or it may be a root.

Do the Ridges Represent Progradational Beach-ridges?

The parallel morphology of the ridges suggests that they are progradational foredunes (cf. Bird, 1976), comparable, in terms of their ridge and swale topography, to modern foredune profiles (e.g., Ahous North, Hopkin's Beach, and the Dune Beach;

Figure 5.2: Topographic Profile of the Ridge Traverse.

Figure 5.3). Analyses of sediment texture, skewness, and bedding structure in modern and paleo-foredune features were undertaken to substantiate this conclusion.

Results of the analysis of 120 textural samples from transects across three beach facies show that, on the basis of skewness and modal class, foredune, logline, and beach-face depositional environments from the same beach cannot be distinguished. If progressive winnowing were to occur it should be present at the Dune Beach, with its 15 m asl back-beach dune, but the Dune Beach samples all had a modal grain size of 0.177 mm to 0.250 mm with no consistent trend in skewness. All samples on Ahous Beach had a modal grain size of 0.125 mm to 0.177 mm, again with no trend in skewness. Table 5.2 presents the skewness data for a typical transect on Ahous Beach (the remaining data are not presented because they are uninformative). Sand samples from the Ridge Traverse were less peaky with the modal grain size of 0.125 mm to 0.177 mm. The results show that texture may characterize a particular beach system, but that individual depositional environments within that system cannot be texturally differentiated. For future reference, sieving at 1/4 phi intervals or finer would be recommended for this type of analysis.

Table 5.2: Skewness values for a typical sampling transect on Ahous Beach

Distance	-7.5	0	7.5	15	30	45	60
Facies	F	FC	LL	LL	BF	BF	BF
Skewness	1.764	0.954	1.361	1.423	1.736	1.492	1.244

Distance (m); F, foredune; FC, foredune cut; LL, logs; BF, beach face.

An examination of a 4 m high cutbank on the south side of Ahous Lagoon, just behind the modern beach, revealed an abrupt change in bedding structure at 2.6 m asl, an elevation equivalent to the that of the modern foredune cut (Table 5.1). At the contact, a rotting piece of logline wood was projecting from the bank. Below the contact the sediments were planar bedded, indicating tidal influence in deposition, whilst above, the bedding appeared scalloped, showing frequent changes in dip and many truncations, indicating periodic tidal/wave incursions and wind as agents of deposition. In comparison, an examination of a 1.5 m exposure in an excavated ridge on the Ridge Traverse (BM-6, 1473 m inland) revealed this same scallopped bedding structure.

Thus, on the basis of morphology and bedding structure it is concluded that the ridges developed as foredunes on a prograding beach.

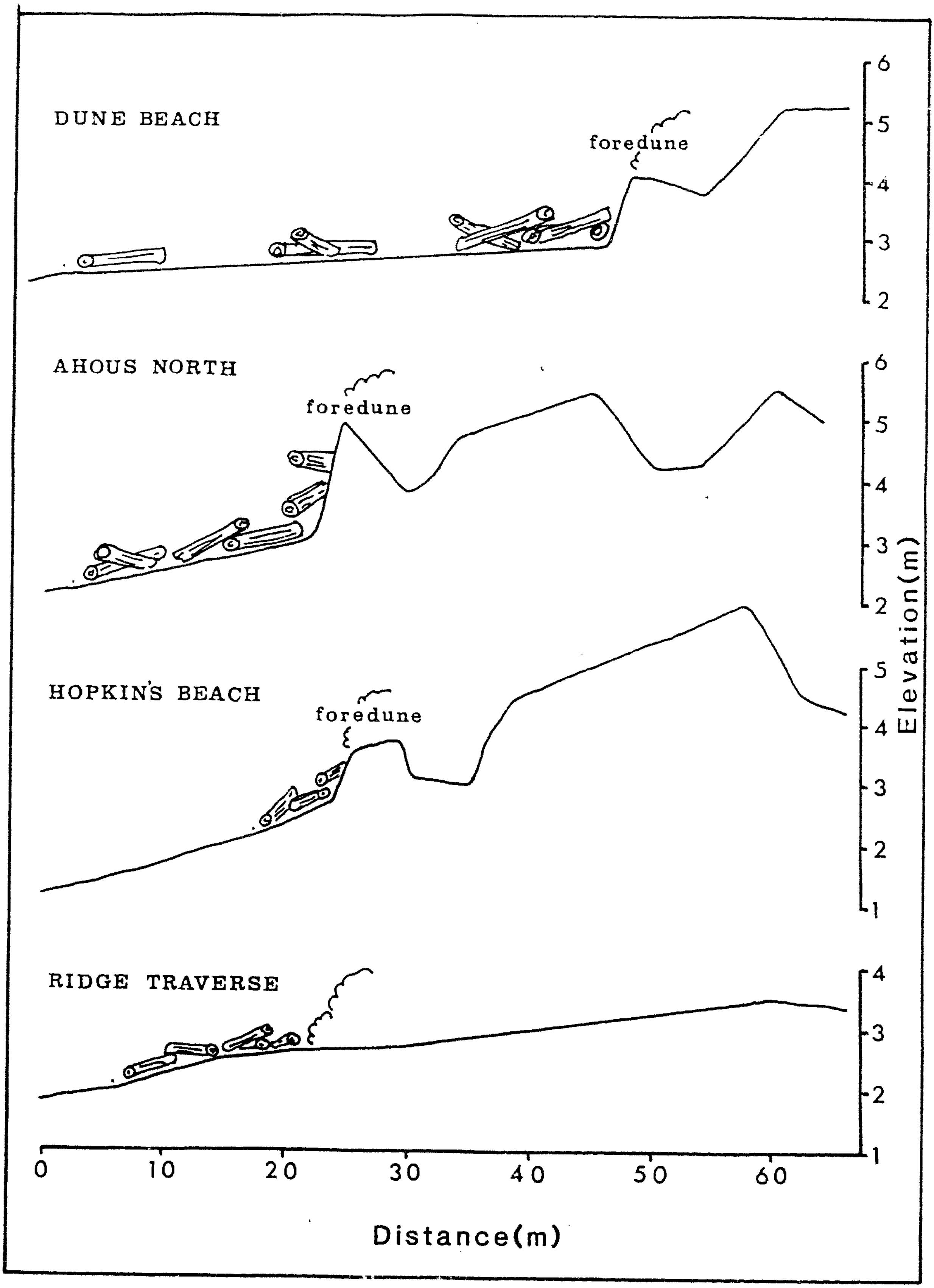


Figure 5.3: Analog Foredune Beaches from the exposed sides of Vargas Island. Note the logs projecting from the foredunes; this shows how an over-estimate of paleodatum occurs.

Where was Mean Sea-level when the Ridges Formed?

The ridge and swale sequence comprising the Ridge System forms an essentially level surface 2.5 km wide (Figure 2.1) which suggests that the ridges advanced during a sea-level stillstand. This stillstand is emphasized by the prominent slope break to the uplands bounding the Ridge System and South Bog, as previously described in Chapter 2. The parallel morphology of the ridges counters the notion that the ridge and swale sequence resulted from rapid co-seismic uplift and subsequent subaerial dune formation, a process that would most likely lead to parabolic dune formation (cf. Bird, 1976) as a result of the ensuing large sediment supply on the exposed platform.

To estimate the height of mean sea-level when the Ridge System developed we can use the wood discovered by Rogers *et al.* (1986). Samples Beta-18336 and Beta-17516 were assumed to represent logline wood by one of Rogers' co-workers (Howes, 1991; pers. comm.). Sample Beta-18336 was obtained from 5.8 m asl in the ridge system. If deposited at the higher-high tide, then a maximum estimate for mean sea-level is 3.8 m asl when the ridges formed. Beta-17516 was obtained from 5.2 m asl. Following the same logic, sea-level was no higher than 3.2 m asl when the ridges formed.

The foredune swale elevation of three modern beaches is roughly 3.0 - 3.7 m asl (Table 5.1). If we subtract this from 6.25 m asl (6.5 m asl less 0.25 m peat accumulation) we can estimate that the ridges formed when mean sea-level was approximately 2.55 - 3.25 m asl.

Thus, three independent estimates of mean water when the ridges formed vary between 2.55 - 3.8 m asl. The latter (3.8 m asl) is probably an over-estimate because the wood of sample Beta-18336 rested at the surface of the sand, and so was deposited on the foredune above the beach-face.

Ahous Lagoon Measured Section

Description

This site is located in the north tributary of Ahous Lagoon, about 50 m upstream from the tributary confluence (Figure 5.1). The geodetic elevations are tied to the tide gauge in the lagoon, and so have a precision of ± 0.10 m. Here, on the left bank on the outside bend of the channel, the cut-bank is 6.5 m high. The Measured Section (Figure 5.4) is in the walls of a large retreating nick point formed by a small seep. The exposure reveals the sediments that constitute the Ridge System (Figure 5.2).

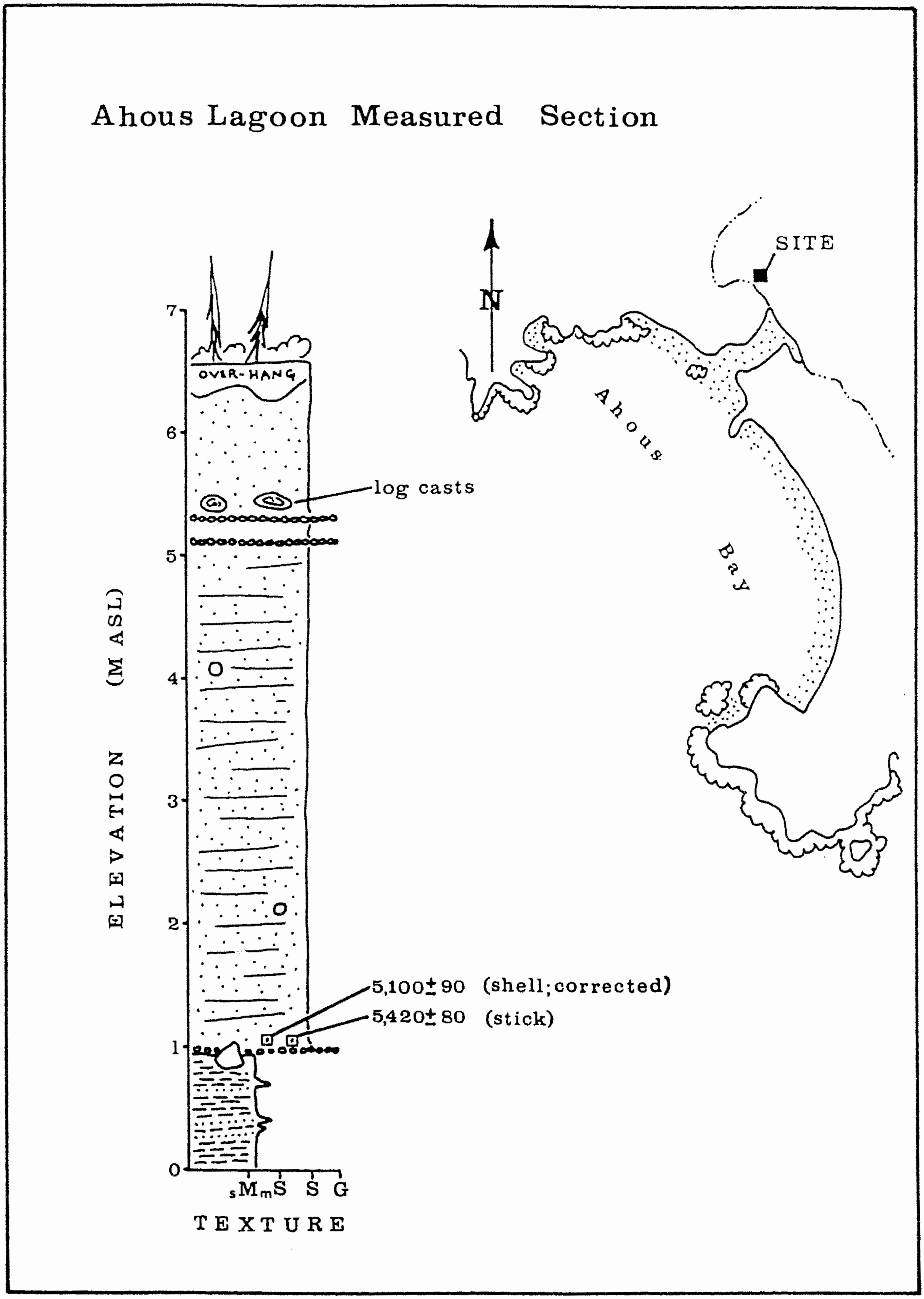


Figure 5.4: The Ahous Lagoon Measured Section. Note the log casts resting on the pebble layers at about 5.3 m asl.

Unit 1: laminated sand and mud

From 0.5 m asl to 0.9 m asl is a blue-grey sandy mud. The mud contains thin (ca. 1-4 mm) interbeds of sandier material. At 0.9 m asl is a sharp contact with the overlying medium/fine sand. The base of the mud was not seen.

Unit 2: medium/fine sand

The first 0.20 - 0.30 m of sand overlying the mud are blue-grey and saturated. In this saturated zone many fossil shells, sticks, and cones were recovered. Directly on the contact is a pebble lag, and a subangular cobble (0.30 m b-axis) was found half embedded in the mud and half projecting into the overlying sand. On the upper surface of this large cobble were many in-situ fossil barnacles.

Approximately 0.25 m^3 of material was excavated from the saturated zone. The sediment was washed through a 1mm sieve in tidewater. A list of the shell materials is given in Table 5.3. Shell materials are listed by tidal tolerance. Clam shell from this zone gave a radiocarbon age of 5,900 ± 90 (SFU-860), or 5,100 ± 90 corrected. A stick from this zone gave a radiocarbon age of 5,420 ± 80 (SFU-857). Seven western hemlock cones and three Sitka spruce cones were recovered.

Shell fauna found in growth position were barnacles, 2 Bodega Tellin (*Tellina bodegensis*), 1 Pacific littleneck clam (*Protothaca staminea*), 1 Chubby Mya (*Platydon cancellatus*), some Purple Olive (*Olivella biplicata*), and many worm tubes constructed of sand and cemented by calcium carbonate. Purple Olive, Bodega Tellin, and Chubby Mya are intertidal species.

The remaining sands extend to the top of the section at 6.5 m asl. The sands are horizontally bedded, emphasized by dark mineral striping, and contain occasional well rounded pebbles. At 5.2 and 5.3 m asl are two distinct, horizontally continuous pebble lags each 0.01 - 0.02 m thick. Immediately above the upper pebble layer, separated horizontally by a few metres, are two 0.20 m diameter holes that are filled with rotten wood. The holes have been exploited by modern roots and ground water through-flow, so the sand around the perimeter of the holes is heavily oxidized. The sands up to 5 m asl are well drained and the color of modern beach sand, but from 5.0 m asl to the top of the section the sands are orange and heavily oxidized and structure could not be seen. The pebble bands are leached white.

Interpretation

The shell material collected from this site, especially the Bodega Tellin and Purple Olive suggest an exposed beach environment, identical to the modern Ahous beach, and not a lagoonal situation, as at present. It is unclear exactly the association of this fauna with mean sea-level at the time of deposition, but the Tellin, Mya, and worm tubes found in-situ, suggest that the 1 m elevation may have been intertidal about 5,100 yr BP. This inference is tentative for two reasons: i) the exact tidal tolerances of these species and the depth to which they burrow is unknown; and ii) the beach at that time would have consisted of a broad shallow wave-cut platform onto which the ridges were prograding. This platform condition was quite unlike the modern beach, and is more similar to that of Bajo Reef off Nootka Island, or the shallows off Estevan Point.

More certain is the implication that a beach-face was landward of this section 5,100 years ago. Eventual beach advance would have buried this hash leading to the accumulation of the sand above.

The pebble layers at 5.1 and 5.3 m asl are similar to lags seen on modern beaches. The holes just above the pebble lag at 5.3 m asl, are one metre below the modern root mat, and are assumed to represent logline material. It is unlikely that these holes developed by piping; it is argued that they are primary features exploited by ground water through-flow. This commonly occurs in forest soils. The position of the logs directly on the pebble lags and planar bedding, lead me to assert that these logs were deposited on the paleo-beach-face between the seaward extent logline and the foredune cut, a zone that has an elevation range of about 0.80 m on the modern beach (Table 5.1), and were not deposited higher in the paleo-foredune. The logs yield a maximum paleo-mean sea-level estimate of 3.3 m asl. Constraining the position of the logs to the paleo-beach face, the sea-level estimate of 3.3 ± 0.10 m asl is associated with a potential positive bias of 0.0 m to 0.80 m; thus, the paleo-datum estimate for this site falls in the bounds 2.4 m asl to 3.4 m asl.

Table 5.3: Shell materials from Ahous Lagoon Measured Section

Species	Freq	Tidal Range	Substrate	Context
Bodega Tellin *Tellina bodegensis*	2	intertidal	exposed sand beach	insitu
Chubby Mya *Platydon cancellatus*	3	intertidal	boring in rock or clay	insitu
Purple Olive *Olivella biplicata*	585	intertidal	sand beaches	some insitu
Sandollar *Dendraster excentricus*	2	intertidal	sandy	hash
Sculptured rock shell *Ocenebra interfossa*	4	intertidal	rocky	hash
Periwinkles *Littorina sp*	35	intertidal	rocky	hash
Finger limpit *Acmaea digitalis*	23	intertidal	rocky	hash
White-capped limpet *Acmaea mitra*	7	int.-20 m	rocky	hash
Hooked Slipper shell *Crepidula adunca*	13	int.-20 m	rocky	hash
Pacific Littleneck *Protothaca staminae?*	6	int.-20 m	sandy bays	hash
Abalone *Haliotis kamtschatkana*	1	int.-30 m.	rocky	hash
Rock entodesma *Entodesma saxicola*	2	mod. shallow	crevices in rocks	hash
Bay mussel *Mytilus edulis*	32	int.-40 m	rocky	hash
Urchin spine *Strongylocentrotus sp*	1		rocky	hash
Purple hinged rock scallop *Hinnites multirugosus*	1	subt.-30 m	rocky	hash
Leafy Hornmouth *Cerastoma foliatum*	2	int.-50 m	rocky	hash
Pink scallop *Chlamys hericius*	2	50-200 m	free swimming	hash
Barnacles	numerous	int.-subt.	rocky	insitu
worm tubes	numerous	int.	sand beach	insitu
Clams *Mactridae*	3	int.-shallow	sand beach	some insitu
other clams	15	int.-shallow	sand beach	hash
Assorted tops and turbans	11	int.-10's m	rocks and weeds	hash
Assorted snails	146	various	rocks and weeds	hash
other limpets	5	various	rocky	hash

Medallion/South Bog Traverse

Description

The profile (Figure 5.5) rises gradually from Medallion Beach to the Medallion Marine Scarp at 4 m asl, 350 m inland. From the crest of the scarp the profile continues as a level surface, at about 6.0 m asl, to CP-1 on the Ridge Traverse. The level surface consists of swamp and open bog, and is informally named South Bog. The stratigraphy in the vicinity of the scarp is referred to as the Medallion Scarp/Raised Beach Complex, and, just landward, the peat accumulation exceeding 4 m depth in South Bog is referred to as the Medallion Subbasin.

Eleven benchmarks (M-1 to M-11) were established on the run from Medallion Beach to the crest of the Medallion marine scarp. Four benchmarks were established along South Bog (SB-1 to SB-4). Due to extremely spongy ground, the closure error (0.16 m) was larger than tolerated (0.05 m) on the South Bog portion of the traverse, so elevations of these benchmarks are not given in Appendix 1, Table A1.2b.

South Bog

South Bog is very narrow and swampy on the north end, widening considerably in the south to form an open bogland 200 m in diameter. Small feeder creeks drain into it from the uplands, while it drains both to the north into the Ridge System and south to Medallion Beach.

The bog is a complex of raised *Sphagnum* hummocks amongst fen. Lodgepole pine and yellow cedar grow on some of the hummocks. Common woody shrubs are *Myrica gale* L. (sweet gale), *Ledum groenlandicum* Oeder (Labrador tea), *Kalmia polifolia* Wang. (swamp laurel), *Empetrum nigrum* L. (crow berry), *Andromeda polifolia* L. (moorwort), and *Vaccinium oxycoccus* L. (bog cranberry), while common herbs are *Drosera rotundifolia* L. (sundew), *Lysichiton americanum* Hulten & St. John. (skunk cabbage), *Sanguisorba officinalis* L. (burnet), *Tofieldia glutinosa* (Michx.) Pers. (False Asphodel), *Trientalis arctica* Fisch. (bog star flower), *Gentiana douglasiana* Bong. (Gentian), violets (*Viola* sp.), cotton grass (*Eriophorum* sp.), grasses (*Graminae*), sedges (*Carex* sp.), rushes (*Juncus* sp.) and *Nuphar lutea* (L.) Sibthorp and Smith (yellow water lily) in deeper water.

Reconnaissance coring with the Hiller corer at intervals along South Bog revealed the depth of peat and associated sediments infilling the basin (Figure 5.5). From 6.0 m asl in the north, the basin deepens to 3.0 m asl between SB-3 and SB-4. The basin floor then rises to 5.0 m asl at SB-4. The two crossectional profiles, at SB-1 and SB-3, show that the

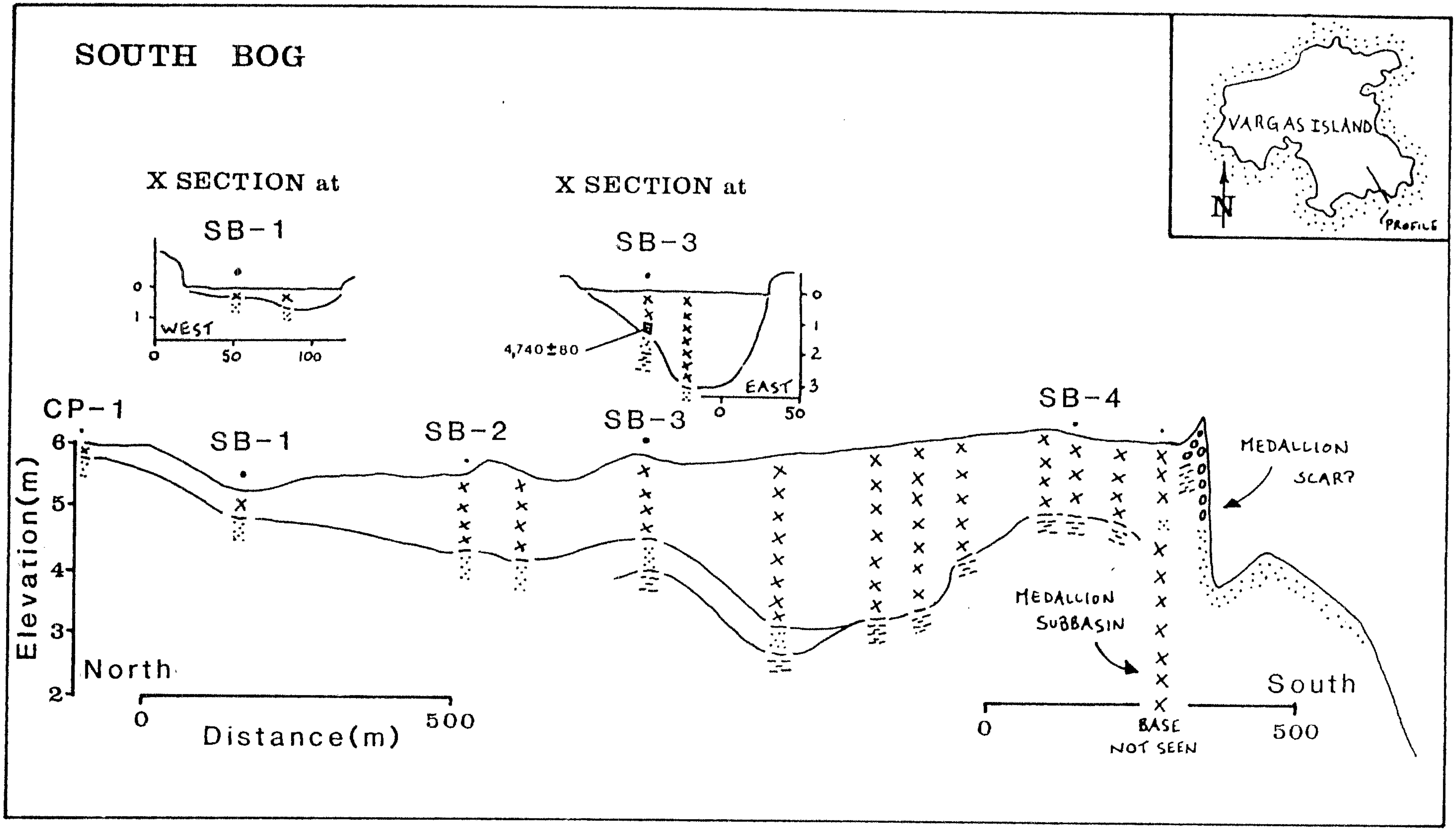


Figure 5.5: Topographic Profile of the Medallion/South Bog Traverse. Note i) the sand wedge filling the South Bog basin from the north; ii) the channel-like cross-sections at SB-1 and SB-3; iii) the basal age of the peat at SB-3 of 4,740±80 yr BP; and the Medallion Subbasin.

narrow part of the bog has a characteristic channel crossection (Figure 5.5). Between SB-4 and the marine scarp the basin deepens to some unknown depth, in excess of 4 m thickness. This deeper basin, informally named the Medallion subbasin, has a northern sill elevation, beneath SB-4, of 5 m asl and a southern sill elevation at the marine scarp of 5.9 m asl. Further coring is needed to properly map the contours of this subbasin and determine if a lower sill exists.

From CP-1 south to the vicinity of SB-3 the basin is floored by sand. This sand pinches out near SB-3 and to the south the basin is floored by sandy mud. Coring in the vicinity of SB-3 revealed that the sands overly the sandy mud. At SB-3, peat contacts the underlying sand at 4.6 m asl, and from this location undisturbed material was retrieved in PVC core and brought back to the laboratory for description and sampling. In the core, the sands were seen to contain vertically oriented, in-situ rhizomes extending at least 0.25 m into the sand (Figure 5.5). These rhizomes have not been identified. From the PVC core, a radiocarbon date on 0.10 m of fibrous peat overlying the sand gave an age of 4,740 ± 80 yr BP (Beta-43484).

Microfossil analyses on five PVC core subsamples, from the zone 4.5 m asl to 4.9 m asl, were conducted by Richard Hebda, the provincial botanist. The two lowest samples spanned the sand/peat contact, while the remaining samples were from the overlying peat and spanned the zone sampled for radiocarbon dating. The description below closely follows Hebda's words: "There was no strong signal of marine conditions in the lower two samples. They contained fine black detritus, possibly pyrite, a potential indicator of marine inundation. There were no saltmarsh plant pollen grains and only one dinoflagellate cyst. The site was likely just at the highest influence of salt water. The sample just overlying the sand contact revealed littoral alder stands surrounded by a western hemlock/spruce forest. The three upper samples revealed the progressive development of a skunk cabbage/shrub swamp, eventually being replaced by a pine dominated bog. The surrounding forest remained dominated by a western hemlock forest, but Sitka spruce became less important" (Hebda, 1991; pers. comm.).

Stratigraphy of the Marine Scarp/Raised Beach Complex

The Medallion Marine Scarp/Raised Beach complex (Figure 5.6) is situated on the southern end of South Bog, about 350 m inland from Medallion beach. The stratigraphy of this feature was revealed in a drainage ditch, dug by homesteaders at the turn of the century, and in a transect of 9 pits extending from the base of the scarp 20 m inland.

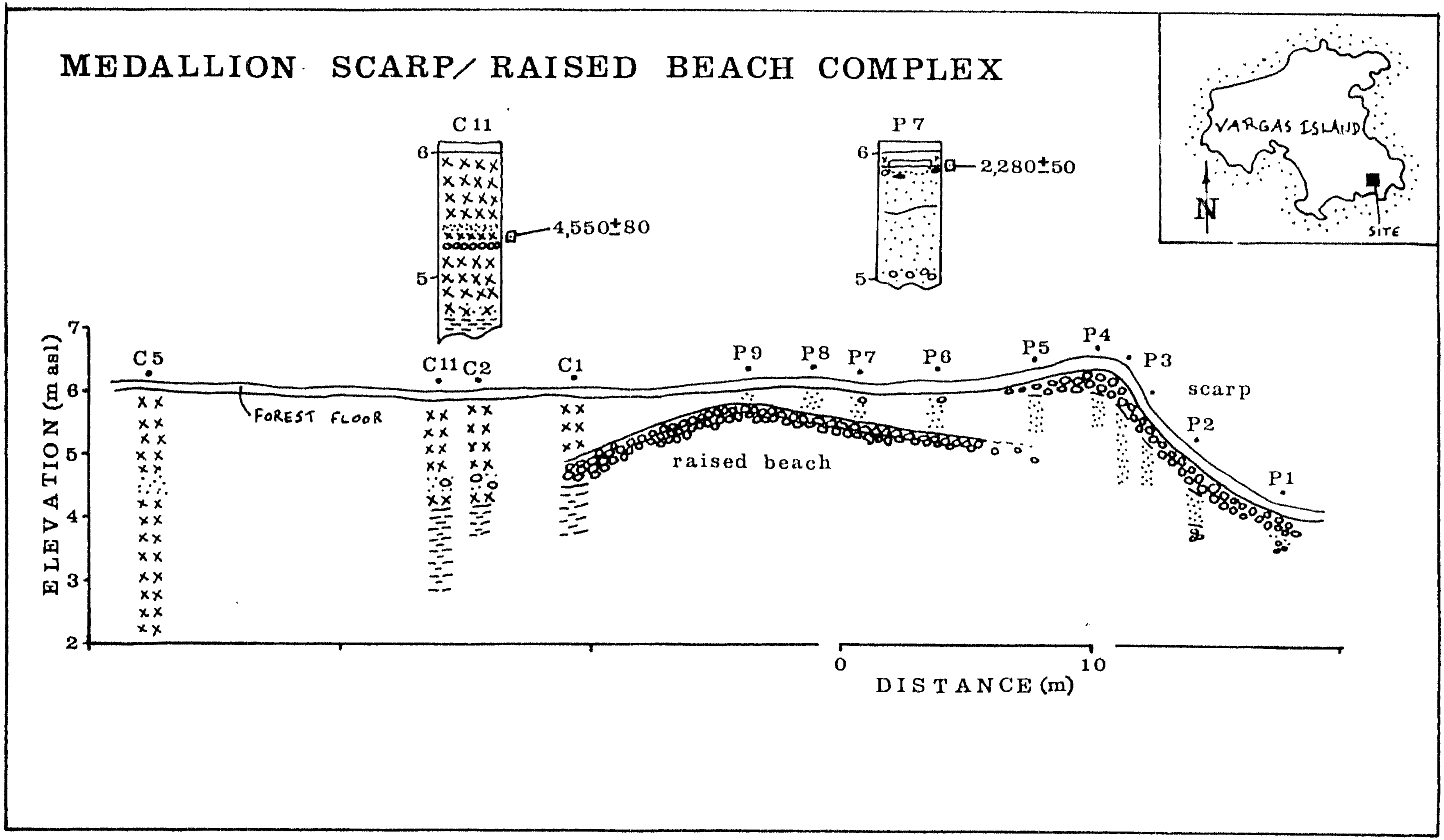


Figure 5.6: Stratigraphic Profile of the Medallion Scarp/Raised Beach Complex. The upper profile of the raised beach was surveyed from a trench; the age on peat at C-11 dates the associated 6 m strandline. The log from P-7 represents logline wood tossed up from the scarp and dates the 4 m strandline.

Unit 1: sandy mud

Blue-grey sandy mud. Some organic staining was observed near the upper contact. Base of the unit was not seen.

Unit 2: pebble gravels and gravelly sand

This unit consists of sandy pebble gravels. The contact with the underlying sandy mud is sharp. The upper surface of these gravels was exposed in the homesteader's trench allowing the surface contour to be surveyed. The surface is shaped like an airplane wing, sloping gradually seaward from the crest at 5.9 m, and dipping more steeply to the landward. Coring with the Hiller corer landward from the gravels, at roughly 5 m intervals, revealed sand stringers that extended into the subbasin. These stringers graded from sandy pebble gravels adjacent to the gravel lens, through pebbly sand, to sand 20 m landward. The sand stringers separate the peat into upper and lower units.

At C-11, 18 m inland from P-9, undisturbed material was retrieved in PVC core and brought back to the laboratory for description and sampling. In the core, 0.10 m of fibrous peat was isolated between a 0.10 m gravel layer below and 0.03 m sand layer above (Figure 5.6). This 0.10 m of peat came from 5.4 to 5.3 m asl, and gave a radiocarbon age of 4,550 ± 80 yr BP (Beta-43480).

Unit 3: medium/fine sands

Tan, medium/fine sands with the occasional pebble. This unit onlaps the gravels of unit 2 on their seaward and the contact with them is distinct. Bedding structure was not noted due to observation lapse.

Unit 4: cobble gravel

This is a sandy cobble gravel. The gravels are thickest at the scarp crest, pinch out landward, and drape over the scarp face, grading into gravelly sand at the scarp base. From the scarp crest, as the gravels pinch out landward, they overly, and grade into, a peaty sand that is distinct from the clean sand of unit 3 below. This stratigraphy is shown in Figure 5.6. A log (dimensions: 1 m * 0.20 m * 0.30 m) was excavated from pit 7. It was resting directly on flat, horizontally lying cobbles at 5.8 m asl, and was half enveloped in the peaty sand. This log gave a radiocarbon age of 2,280 ± 50 (Beta-38876).

Medallion Subbasin Macrofossil Analysis

The vegetation at the subbasin site, consists of a 50 m belt of Western hemlock and red cedar at the southern end of South Bog. Western hemlock, Sitka spruce, and red cedar grow with *Rubus spectabilis* Pursh. (salmonberry), *Malus diversifolia* (Bong) Roem. (Pacific crabapple), and huckleberries (*Vaccinium* sp.) in the swampy area from the scarp to the modern beach.

The two longest cores C-3 and C-5 were selected for analyses. C-5 was 17 m north of the crest of the internal gravel berm and C-3 was 27 m north of the crest. Due to time constraints, I decided to focus my analysis on the zone around the distinct sand stringers of unit 2, ca 4.0 m asl to 5.0 m asl in the selected cores. Note, this is just below the elevation of the sill at SB-4. Ten subsamples from each core were selected from this zone. In addition, four subsamples from C-3, at 1.95 m asl to 2.45 m asl, were analyzed to establish the nature of the woody peat at depth. In C-5, samples from this depth were considered similar to deeper C-3 samples as three hemlock cones and numerous needles were seen to be preserved in good condition.

The results of the analysis are presented in Figure 5.7. A broad comparison can be made between the samples at depth (1.95 m asl to 2.45 m asl) and those from the shallower zone (3.95 m asl to 5.10 m asl). First, the deeper peat is noticeably more woody, with spruce and hemlock dominant in the needle assemblage. Cedar occurs in small percentages in two samples and pine is noted in one. In contrast, while spruce and hemlock still dominate the shallower samples, cedar and pine are consistently present. Also, achenes of *Cyperaceae* are more common in the shallower samples. These observations suggest a transition from mesic to swampy conditions in the immediate locale of the subbasin.

Conifer fossils were present in all samples except those from the modern bog. This contrast suggests that true bog never developed at the Medallion subbasin. The presence of *Rubus* seeds throughout the core strengthens this assertion. This does not rule out the possibility of that the basin was occupied by a small pond, as the modern intertidal sample was dominated by fossil needles, and fossil needles are commonly found in lacustrine deposits (Dunwiddie, 1987; Wainman and Mathewes, 1990).

The zone in C-3 from 4.30 m asl to 4.55 m asl deserves special mention for a few extraordinary occurrences. In this zone there is a spruce needle spike in both cores. Throughout this zone spruce needle concentrations (not noted in Figure 5.7) increase by an order of magnitude, exceeding 1000/100 ml and attaining a maximum of 2500/100 ml. A *Pinaceae* seed spike occurs as well. Most significant is the occurrence of pondweed (*Potamogeton* sp.) seed in this zone in both C-3 and C-5. Pondweed is an aquatic indicator (Mason, 1957). Furthermore, an elytra of the waterlily leaf beetle (*Donacia* sp.) was found in C-3 in the zone 4.35 m asl to 4.55 m asl. The waterlily leaf beetle feeds on the foliage and pollen of waterlilies and skunkcabbage (Niering, 1985).

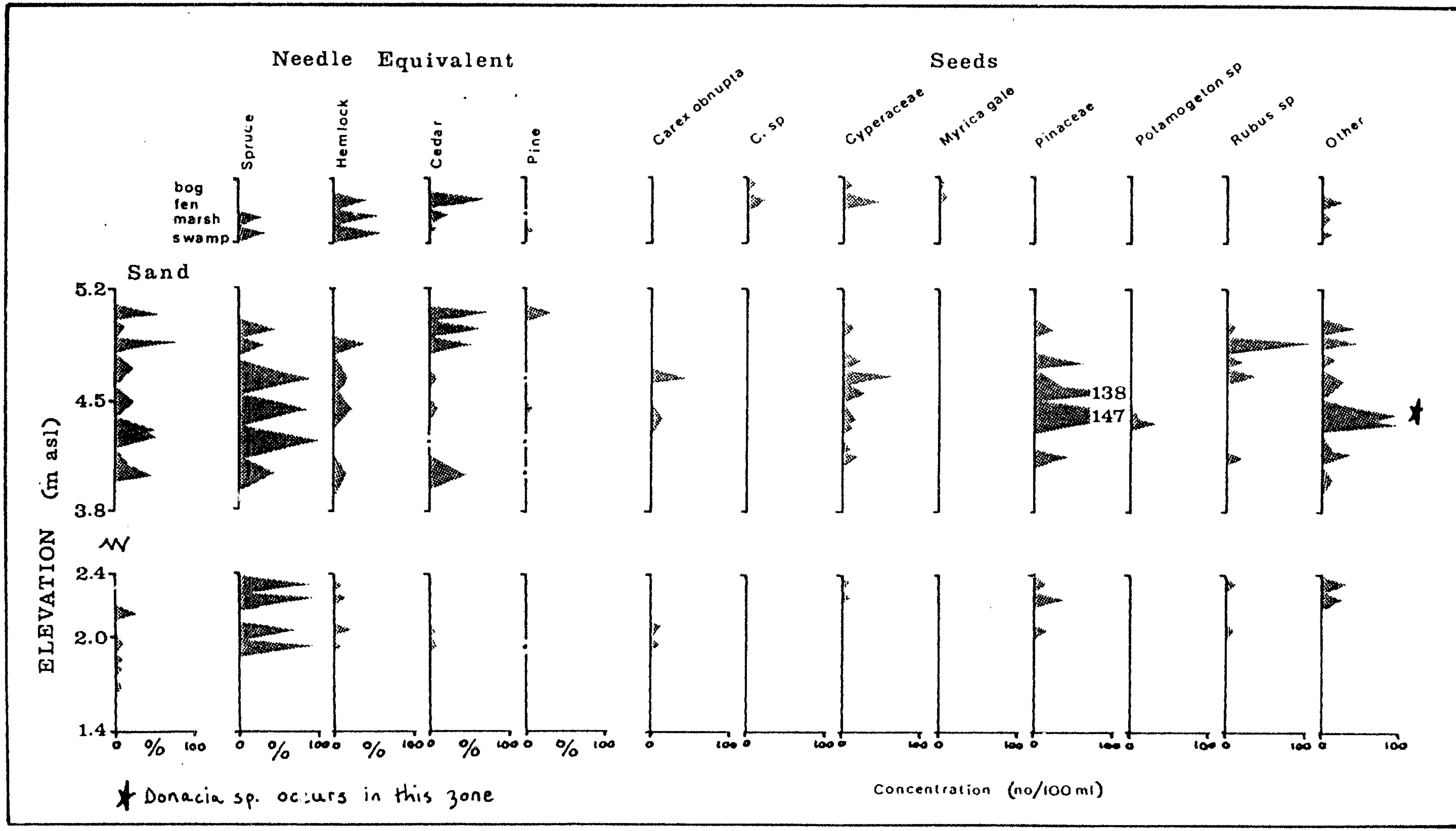


Figure 5.7: The Medallion Subbasin Macrofossil Diagram. Sand (% by weight); Needle Equivalent (% by sum). Note that sample volumes were small - about 35 ml - so the concentrations (no./100 ml) appear more significant than real.

The texture of the sand fraction within the core samples is typical of local beach sand, with a modal grain size of 0.177 mm to 0.250 mm. Pebbles were noted in C-5, 4.40 m asl to 4.90 m asl and in C-3, 3.95 m asl to 4.35 m asl. Examination of these pebbles under the microscope indicated a highly vesicular, volcanic lithology. Thus, the pebbles are of relatively low density, and some were noted to float. Two pieces of unidentified shell fragment were found in screen; one from C-3, 4.55 m asl to 4.75 m asl and the other from C-5, 4.90 m asl to 5.00 m asl. In C-3, 1.95 m asl to 2.45 m asl sand was not visible in the samples, but by C-3, 3.50 m asl to 3.75 m asl sand disseminated in the peat was visible. A discrete sand layer was noted at C-3, 4.75 m asl to 4.85 m asl, but not in C-5. Sand contents vary from 9% to 75% by weight, appearing more constant at depth and more variable higher in the cores.

Interpretation of the Medallion/South Bog Traverse

The sandy mud that is seen to underlie South Bog and the Ridge System is interpreted as post-glacial and possibly early-Holocene marine mud deposited during the post-glacial marine inundation.

In the Marine Scarp/Raised Beach complex, the gravels of unit 2 and the onlapping sands of unit 3 are interpreted as a beach facies. The peat from C-11 (Figure 5.6) indicates a minimum date for the formation of the raised beach about 4,550 yr BP, and indicates a maximum date for the overlying sand. Thus, the sea-level stand that occupied this raised beach spanned the period 4,550 yr BP.

This arrangement of gravel onlapped seaward by sand is identical to modern gravel and sand on the beach on the south west side of Stubbs Island (Figure 5.8). The crest of the Stubbs Island gravels is 2.7 m above mean sea-level. By analogy, we can estimate that mean sea-level was 2.7 m below the crest of the gravels in the Medallion berm. Thus, about 4,550 yr BP, at the time of the occupation of the Medallion Raised Beach, it is estimated that mean sea-level would have stood at 3.2 m asl.

The sand layers in the subbasin are interpreted as the product of periodic wave/aeolian incursions from the paleo-Medallion Beach. This is supported by the presence of vesicular pebbles in core in the shallower samples, as well as the discrete sand layers. The presence of shell also suggests incursion, but birds often carry and drop shell inland. Since no saline indicator species were noted in any of the subsamples analysed, this suggests the basin sill was never exceeded by the higher-high tide. Given a lower sill elevation of 5.0 m asl for the subbasin, an estimate of paleo-mean sea-level of 3.0 m asl is provided. Note the consistency of these two estimates with preceding estimates.

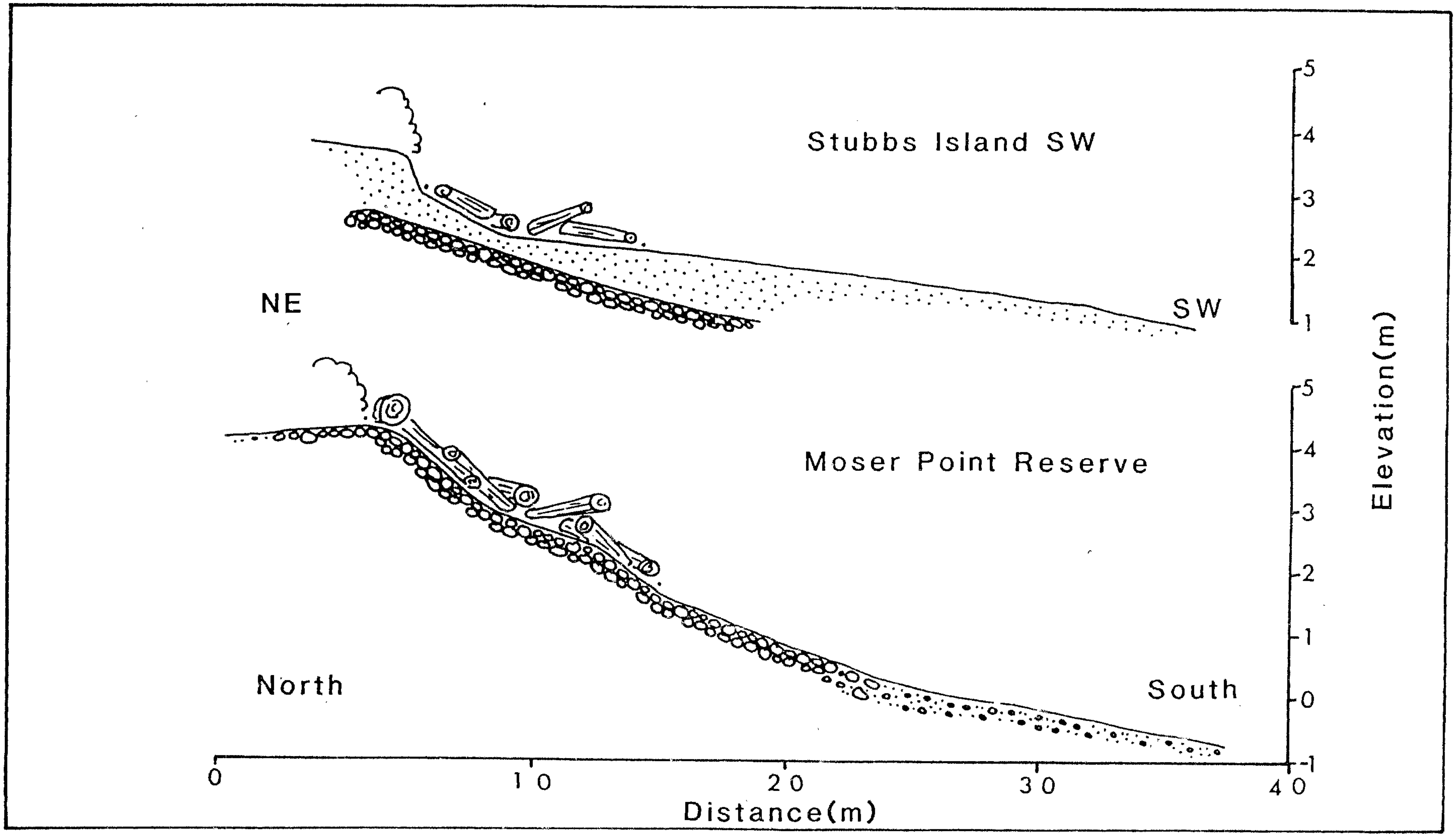


Figure 5.8: Analog Profiles from Stubb's Island and the Moser Point Reserve.

Moreover, if monotonic emergence from the post-glacial marine limit had taken place, or the subbasin was subject to significant wave influence prior to emergence, then we would expect marine muds, or gyttja, to fill the basin (e.g. Village Lake; Hebda and Rouse, 1979), which is not the case. The occurrence of woody peat to 4 m depth in the subbasin suggests that mesic forest conditions existed in the vicinity of the subbasin prior to the formation of the raised beach. The transition to swampy or ponded conditions implies a rise of the local water table in response to sea-level rise.

The sand flooring the South Bog basin, pinching out in the vicinity of SB-3, suggests that the main basin in South Bog was open to tidal influence from the north, but not the south. The channel-like nature of the crossections at SB-1 and SB-3 also suggest a tidal channel existed at some time. Microfossil evidence from the top of the sand wedge and the base of the peat accumulation at SB-3 suggests the possibility of marine inundation, or at least a supratidal setting. Further, microfossil analyses suggest that the initiation of peat infill at SB-3 represents fresh water conditions. The radiocarbon sample from the base of these fresh water peats gave an age of 4,740 yr BP (Figure 5.6). This is a minimum age for the formation of the sand wedge and indicates the cessation of the sand supply, suggesting closure of the lagoon entrance prior to 4,740 yr BP. Subsequent isolation would explain the rise of the local water table in South Bog, as indicated at this site by the transition from near-shore alder stands to pine dominated bog at SB-3, and the appearance of *Potamogeton* sp. and *Donacia* sp. in core in the Medallion subbasin ca. 4,550 yr BP.

The Medallion marine scarp, draped by gravels, is identical in profile, sediment characteristics, and aspect to the Moser Point Reserve Beach (Figure 5.8). Superimposing the two profiles we can estimate that mean sea-level was about 2 m asl at the time of formation of the Medallion scarp. The radiocarbon date from pit 7, a log associated with the scarp gravels gives a minimum age of 2,280 ± 50 yr BP for the formation of this scarp. Thus, from the position of sea-level 2.55 to 3.8 m higher than today, 5,100 to 4,030 yr BP, mean sea-level had fallen to about 2 m asl by 2,200 years ago.

Buckle Berm

Description

This site is located in Buckle Bay, on the property of Neil and Marilyn Buckle, just at the eastern trail-head of the Telegraph Trail (Figure 5.1).

From the modern beach the profile rises in a series of steps to the crest of a cobble gravel berm at 8 m asl, about 40 m inland. This is Buckle Berm (Figure 5.9). The

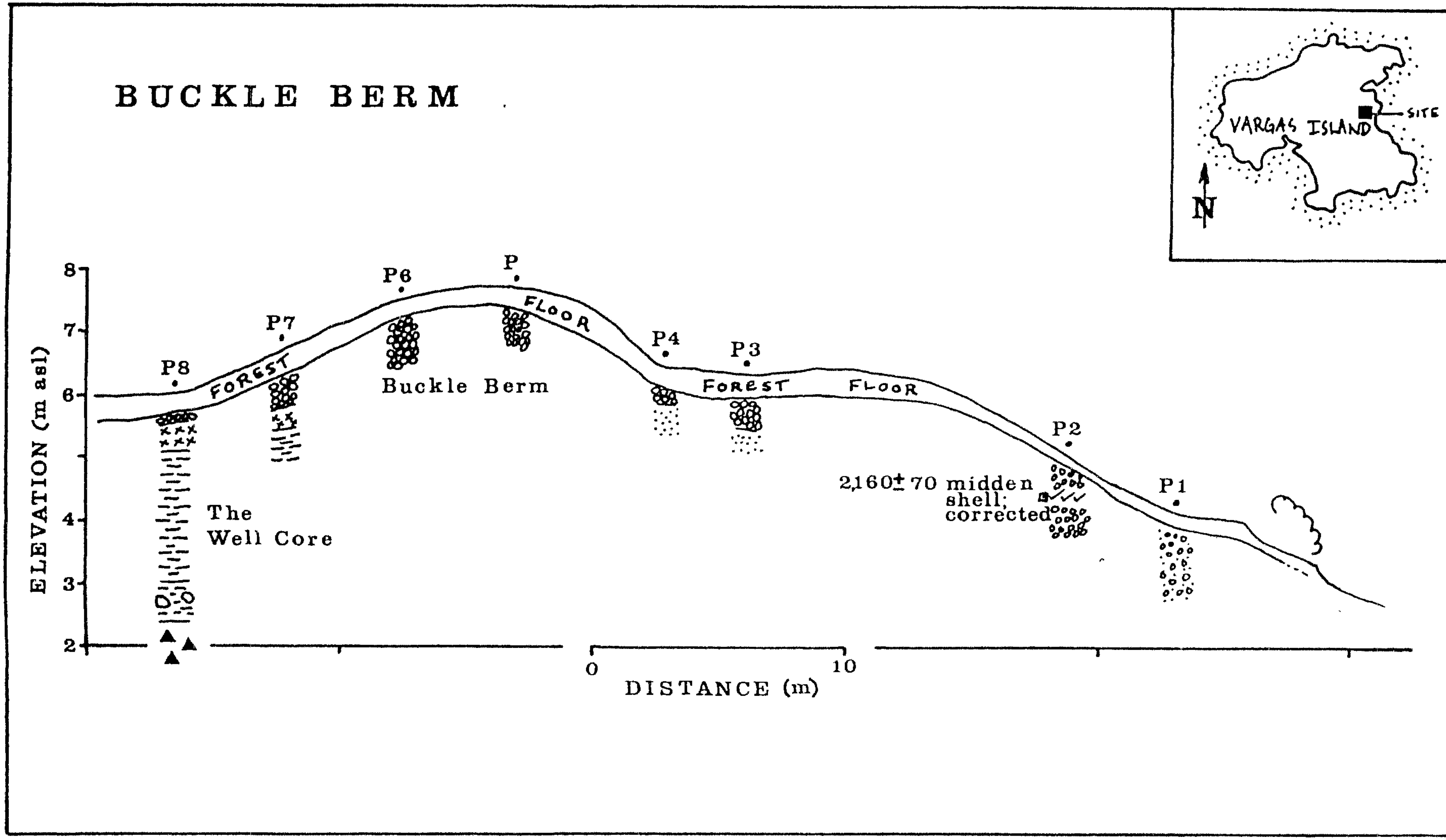


Figure 5.9: Topographic Profile of Buckle Berm.

landward side of Buckle Berm is 6 m asl and forms a swamp 0.5 km^2 in area. The swamp is drained by the creek flowing into Hovelaque's Bay (Figure 2.1).

A transect of 8 pits, spaced at regular intervals inland from the beach, reveals the stratigraphy of Buckle Berm. In pit 2, a 0.20 m thick midden deposit rests, at 4.3 m elevation, on an extremely well sorted pebble layer similar to the modern spring tide berm gravels. The midden contains numerous fire-cracked rock, charcoal, bone, and shell. Clam shell from this midden gave a radiocarbon age of 2,960 ± 70 (SFU-872), or 2,160 ± 70 corrected.

The stratigraphy just to the landward of Buckle Berm, was described by Neil Buckle from dug well holes. The site was later vibracored for verification and sampling. The core is referred to as the Well Core (Figure 5.10). The base of the core at 2.5 m asl is a compact diamict that could not be penetrated with a shovel (Buckle, 1990; pers. comm.).

The Stratigraphy of the Well Core

Unit 1: stoney mud

This unit is about 0.50 m thick. It rests directly on the compact diamict, but the contact was not seen. It is a soft, grey, stoney mud with boulders to 0.70 m b-axis. The unit is rich in shell hash of a variety of species. Many of the clasts have barnacle scars, and some of the shells were found in growth position, with the valves together. These included *Mytilus edulis* Linn. (Bay Mussel), *Hiatella arctica* Linn. (Arctic Rock Borer), and other unidentified clams and snails. A sample of hand-selected barnacle tests gave a radiocarbon age of 12,970 ± 80 (Beta-42922), or 12,170 ± 80 corrected.

Unit 2: muddy sand

This unit is about 1.1 m thick extending from 3.0 m to 4.1 m asl. It is a tan colored muddy sand with small (1-5 mm) orange mottles. A single angular pebble clast (0.02 m b-axis) was noted. The lower contact with unit 1 is relatively sharp, but not distinct, while the upper contact with unit 3 is gradational over a 0.20 m zone.

Unit 3: sandy mud

This unit is about 1.2 m thick extending from 4.1 m to 5.3 m asl. It is a blue-grey sandy mud. Its texture is somewhat variable with depth, some stringers being more sandy. Some fine (< 0.25 mm) dispersed shell fragments were noted. No clasts larger than sand were present.

The contact with the overlying unit is sharp, but the upper 0.10 m is organically stained suggesting relict soil development, yet because of the saturated, mucky character of

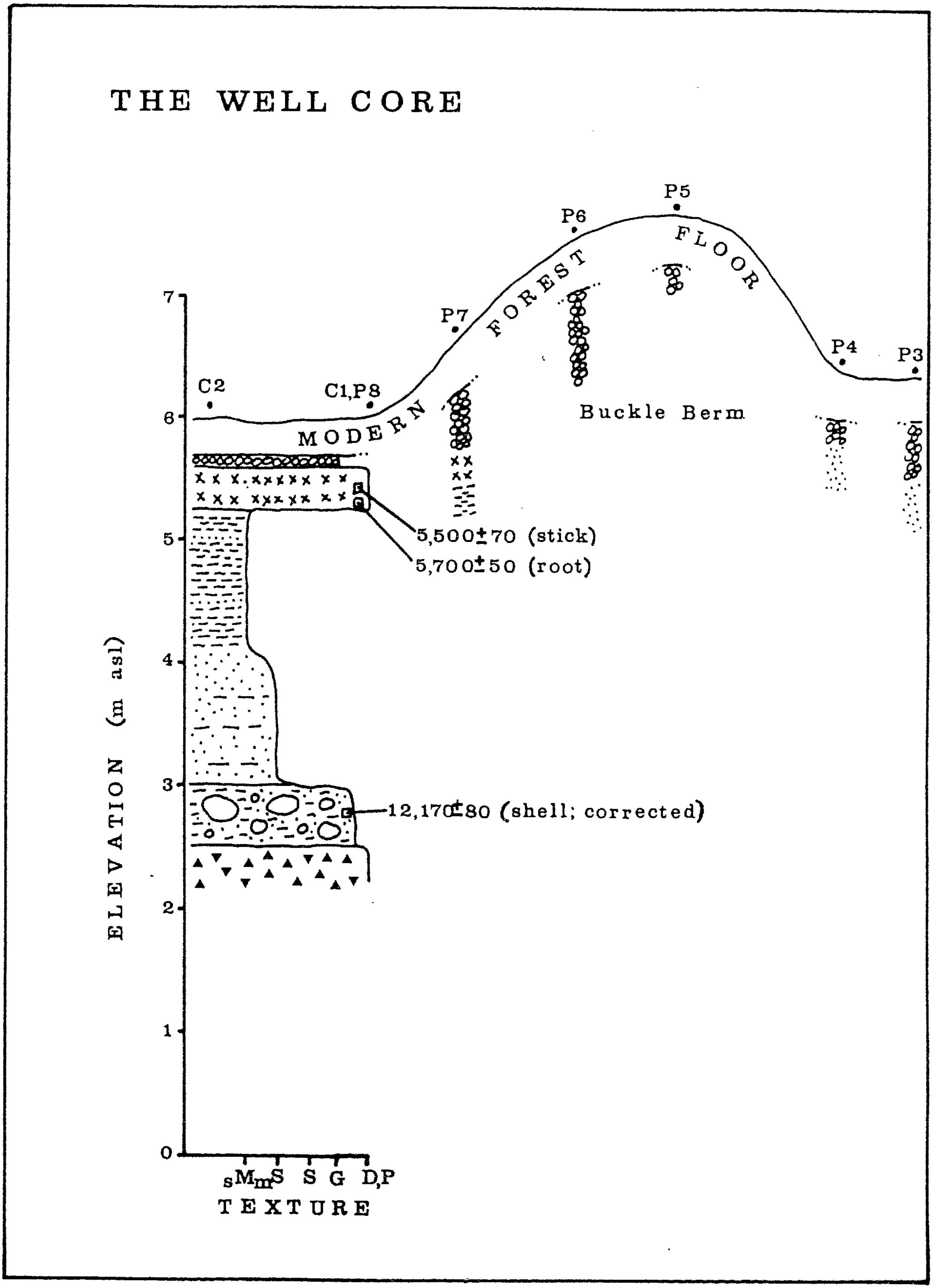


Figure 5.10: Stratigraphic Profile of the Well Core. The stony mud contained *Hiatella arctica*, a glacio-marine indicator species.

the test pits actual soil peds were hard to observe, so the presence of soil structure was not positively ascertained.

Unit 4: woody peat

This unit consists of 0.40 m of woody peat, containing abundant wood fragments, sticks, and charcoal. At the base, at 5.3 m asl and directly overlying unit 3 is a tangle of thick (0.15 m dia.) conifer roots. These have been examined from 2 shovel pits and 2 large 1 m by 2 m pits excavated for wells. The roots appear to be in-situ and have been preserved by an elevated water-table. When one of these roots was removed for a radiocarbon sample the fresh-cut surface was yellow and clean, like modern wood, but within minutes the fresh-cut surface darkened to green, then black. The root yielded a radiocarbon age of 5,700 ± 50 (Beta-38875). A stick 0.20 m above the root layer gave an age of 5,500 ± 70 (Beta-41015).

Unit 5: cobble gravel

Unit 5 consists of about 0.20 m of sandy, clast supported, hard, cobble gravel. Five metres landward the gravel grades into a gravelly sand about 0.10 m thick. This gravel unit is the landward taper of Buckle Berm.

Interpretation of the Buckle Berm Site

Units 1 through 3 in the Well Core are interpreted as glaciomarine lithofacies. The shell rich stoney mud of Unit 1 suggests an ice proximal setting. The fining upward of the sequence of units 1 through unit 3 probably represents a shift from ice proximal to ice distal conditions. Similar interpretations have been made for glaciomarine mud sequences by Clague (1985) and Miller *et al.* (1989). More core would have to be viewed from various sites to define these lithofacies properly in the manner of Powell (1981). The lack of more dropstones (note the single angular pebble clast in unit 2) might be due to i) shallow marine conditions with the iceberg track being confined to the deeper channels, or ii) the rapid decay and retreat of ice upfjord.

The peat of unit 4 at 5.3 m asl, is interpreted as forest floor, and not as flotsam deposited at the base of the gravels. The root mat is extensive and has been viewed in a number of large pits. The organic staining extending into the lower unit strengthens the assertion of a forest floor, but more work needs to be done to confirm this point. This forest floor suggests that the site became emergent, but the timing and extent of emergence is unknown. By 5,700 years ago the high water level had risen high enough, from some lowstand, to kill this vegetation (i.e., to 5.3 m asl), and sometime thereafter the site was transgressed by marine deposits.

A very similar stratigraphic section is described by Forman (1990; his fig. 6) for recent submergence in Spitsbergen. In his section the storm berm overlies 0.50 m of peat, and the crest of the storm berm is 2.6 m above the high tide. Another feature similar to Buckle Berm occurs on the south side of Boat Basin, just west of Hesquiat Lake. This modern feature is also composed of cobble gravels. Although not surveyed, its landward side approximates the higher-high tide and is colonized by beach grass. The crest of the berm is at least a full metre above the high water mark, and the seaward face slopes steeply.

These comparisons suggest that Buckle Berm formed during the same stand of the sea that produced the Medallion Raised Beach and the Ridge System, some 5,100 to 4,030 years ago.

After the formation of the Buckle Berm, sea-level began to fall. The clean, well sorted pebble gravel at 4.3 m asl, upon which the midden lies, is indistinguishable from the modern spring tide pebble berm at 2.3 m asl on Buckle Beach, which suggests that mean sea-level was at or below about 2 m asl, 2,160 years ago. It should be noted that the age and mean sea-level estimate for this feature and the Medallion Scarp are coincident.

The Ratcliffe Terrace

Description

The Ratcliffe Terrace is located at Hovelaque's Bay, about 50 m inland from the modern beach (Figure 5.1). A plan and profile of this site is given in figure 5.11. An array of 19 pits on the top of the Ratcliffe Terrace show the stratigraphy of this feature. The pits were arranged in transects from the outcrop at the terrace edge out onto the terrace surface in an attempt to isolate beach gravels marking the 6 m strandline.

Unit 1: diamict

This is a weathered diamict with angular clasts and a gritty mud matrix. This diamict outcrops on the landward edge of the terrace against the rock outcrop of the upland, but seems to dip steeply beneath the terrace surface.

Unit 2: sandy mud

Blue-grey sandy mud, as described elsewhere. It appears that this unit blankets unit 1. In pit 17, organic staining and sticks were found on the contact between this sandy mud and the overlying sands.

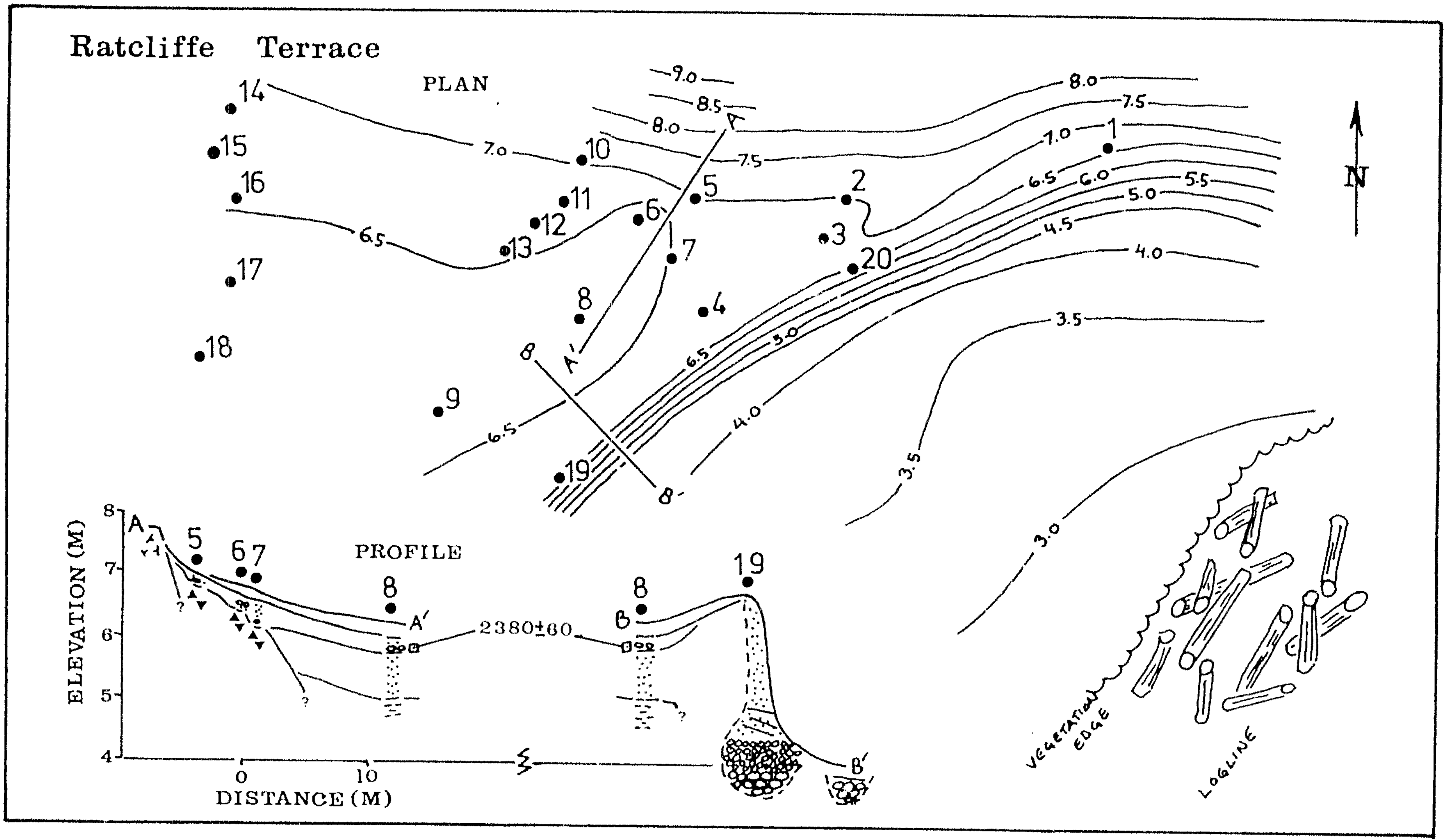


Figure 5.11: Plan and Profiles of the Ratcliffe Terrace. 0.50 m contour interval. Dots with associated number are pits.

Unit 3: medium/fine sands

Medium/fine sands, pinching out against the upland and thickening seaward. Heavily oxidized near the marine scarp, but green and reduced farther inland. Planar bedding was noted in pit 8; bedding was made evident by thin horizontal stringers of rotten wood. The occasional pebble occurs in the sands.

Unit 4: pebble lag and black peat

A lag of well rounded pebbles descends from 6.9 m asl in pit 5 where it rests on diamict, to 6.5 m asl in pit 6, then to 6.2 m asl in pit 9. From pit 6 to pit 9 the pebbles occur as a lag on sand. 6.9 m asl is the highest elevation that rounded pebbles were found in the 4 pits along the terrace edge. In the pits on the terrace the pebbles are associated with a sandy black peat. In pit 8 this black peat is 0.20 m thick. The lower 0.10 m of this peat gave a radiocarbon age of 2,380 ± 60 yr BP (Beta-43479).

Unit 5: sands and cobble gravels

Pit 19 exposes the sediment of the scarp face. Six stadia positions at the base of the scarp range from 3.8 m asl to 4.5 m asl. From 3.7 m asl to 4.8 m asl are the sands and gravels of unit 5. From 3.7 m asl to 4.4 m asl these consist of clast supported cobbles with a gravelly matrix. At 4.4 m asl is a sharp contact with planar bedded sands which extend to 4.8 m asl. The heavily oxidized sand of unit 3 are exposed in the scarp from 4.8 m asl to the crest at 6.6 m asl.

Interpretation

This feature is a constructional terrace formed when the highest wave influence reached about 6.9 m asl. The radiocarbon age of about 2,400 yr BP is young for this elevation. Since this peat underlies 0.30 m of forest floor the material is believed to be contaminated by modern rhizomes. This date is considered unreliable and gives a minimum age only.

When sea-level stood at this elevation, the swamp connecting Hovelaque's Bay and the Buckle Berm site would have been a shallow bay. The Ratcliffe Terrace represents the edge of this bay, probably near its mouth. This bay might have been similar to the shallow bays farther east on Vargas Island. For instance, Mud Bay is about 1.5 km long and completely dries during lower-low spring tides, thus it is 4 m deep at its deepest. The central part of the bay is deepest shallowing gradually to the edges. Lining the edges of the bay are narrow salt marsh communities.

Unit 5 is considered to represent gravels deposited in an undercut at the foot of the 4 m asl marine scarp.

Secondary Sites

Meadow Bay Gravels

Description

This site is located on the trail that leads from Hovelaque's Bay to Meadow Bay (Figure 5.1). Where the trail crosses Meadow Creek there is a 2 m high, 40 m long, section on the right bank (Figure 5.12). The upper surface of the section is at 13.5 m asl, and defines the surface of a terrace remnant. The lower surface is at 11.5 m asl and consists of bedrock. The section is composed of 2 units. The seaward unit, unit 2, is carved into and slightly onlaps the landward unit, unit 1.

Unit 1: compact diamict

This is a massive, matrix supported, very compact grey diamict, with clasts of local and granitoid lithology. The matrix is composed of a gritty mud. In the clast fraction pebbles and cobbles dominate, and boulders are few. Clast roundness varies from subangular to rounded.

Unit 2: heavily oxidized gravels

This unit consists of red-brown, heavily oxidized, well rounded, clast-supported, sandy pebble and cobble gravels. It appears to be composed of lenses and sloping bands of gravel of different textures, from well sorted pebble gravel to coarser cobble gravel. It is not bedded and does not show any fine stratification. The matrix is sandy containing a proportion of fines, giving a greasy feel. In one spot, just above bedrock, are two 0.05 m thick gravel bands with a soft, grey mud matrix. These bands are about 1 m long, and slope down and landward to the rock contact. Just where the trail crosses the creek about 0.40 m of unit 1 material separate unit 2 and bedrock.

Interpretation

Unit 1 is interpreted as a lodgement till. Unit 2 is interpreted as an early post-glacial beach lag consisting of reworked till and ice-contact debris. Some of the fines in the matrix of unit 2 probably result from the translocation of clays, but the sloping mud bands are undoubtedly original materials. A rapid supply of sediment or rapid emergence (or both) is suggested by the immaturity of the matrix materials of unit 2. It is unlikely that unit 2 is

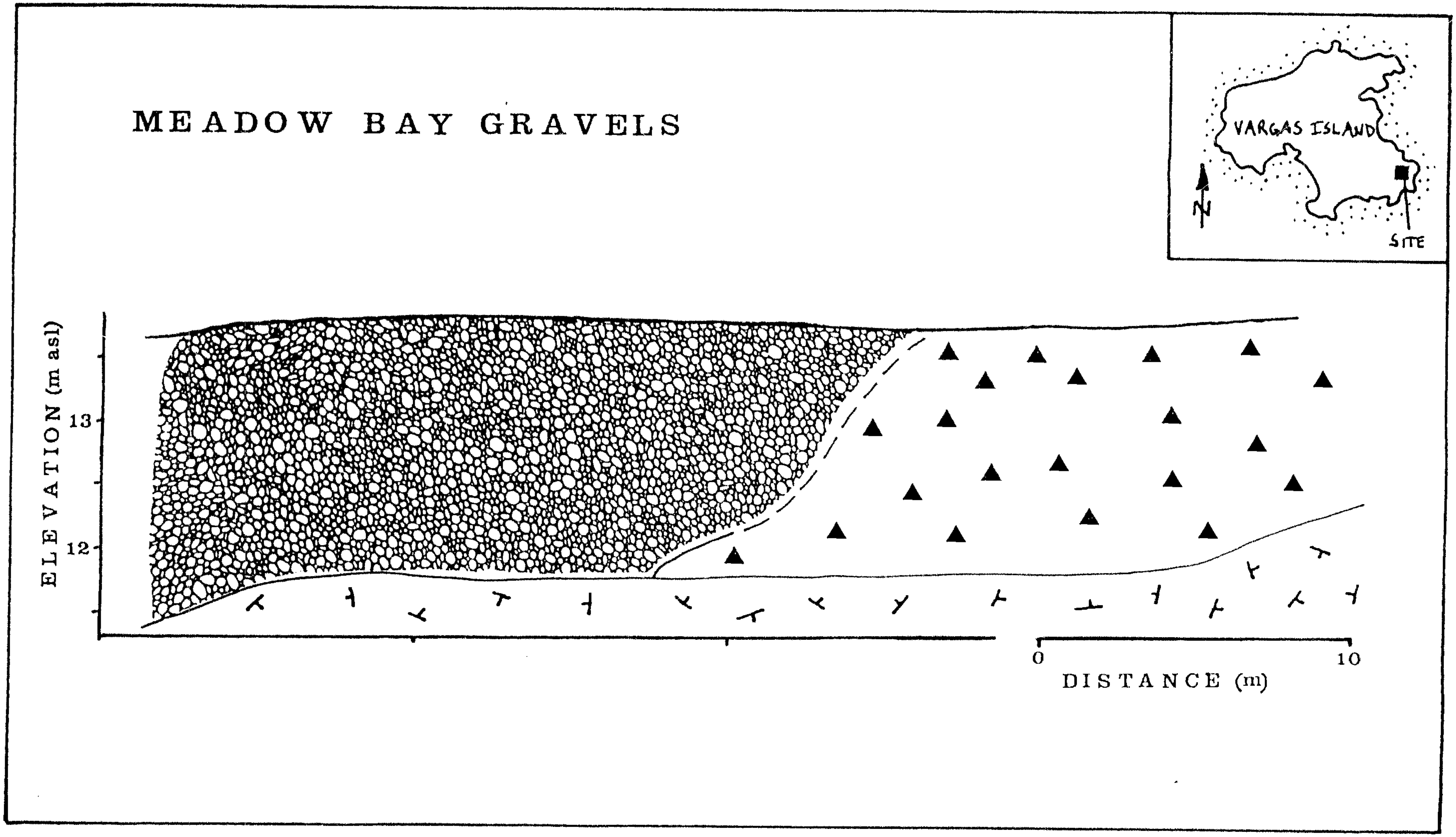


Figure 5.12: Stratigraphic Profile of the Meadow Bay Gravels

glaciofluvial because Vargas Island is isolated by deep channels from the distant fjord heads, so there is no obvious source for glaciofluvial discharge. As well, no fluvial sedimentary structures are present in the section. Unit 2 is interpreted as marine gravel lag, and is the highest marine unit found on Vargas Island.

The Medallion Strandflat

Description

This terrace (Figure 5.13) is located about 50 m inland from the north-east side of Medallion Beach (Figure 5.1). From the back-beach forest a footpath follows the east side of the creek to this feature. The site can be recognized by 6 vertically standing, very rotten posts; probably the remains of a structure built by homesteaders at the turn of the century.

An elevation was carried to this site from benchmark M-2 on the Medallion/South Bog Traverse. Two profiles of the terrace were surveyed each starting from a central benchmark and running perpendicular to one another. Both profiles reveal a 9 m asl terrace inclining slightly over a distance of 40 m to abrupt bedrock cliffs with scarp elevations of 9.6 m asl. A 0.01 m weathering rind occurs on the outer surface of the bedrock scarps. The weathering rind is so thick and soft it could almost be mistaken for compact sediment. The surface above the scarp reflects bedrock topography.

Two pits were dug on the terrace, one close to the bedrock scarp and one near the benchmark. Near the scarp about 0.10 m of sand containing a few well rounded pebbles rests directly on bedrock. At the benchmark sandy mud, likely glaciomarine, occurs and masks the underlying topography of bedrock. That the sandy mud underlies most of the terrace is attested to by its presence in a drainage trench running from the posts to the edge of the terrace.

Interpretation

This site is interpreted as a wave-cut terrace, formed when the higher-high tide was at approximately 9.6 m asl. During the post-glacial marine inundation the sandy mud was deposited masking the topography of bedrock, then during the period of rapidly falling sea-levels the terrace was wave-cut, leaving only a skim of sand near the base of the marine scarp. The heavy weathering rind indicates the considerable age of the scarp.

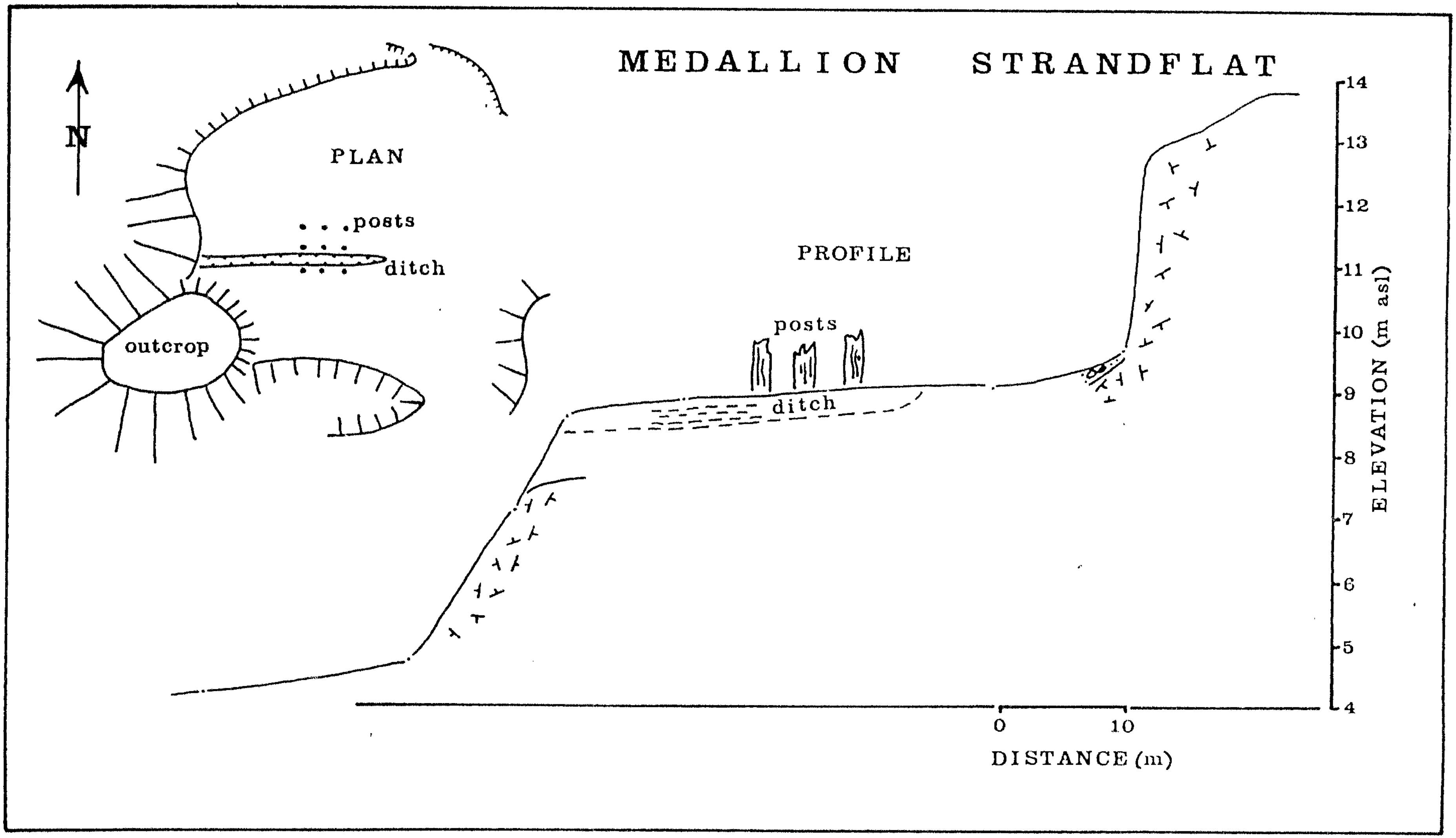


Figure 5.13: Plan and Profile of the Medallion Strandflat.

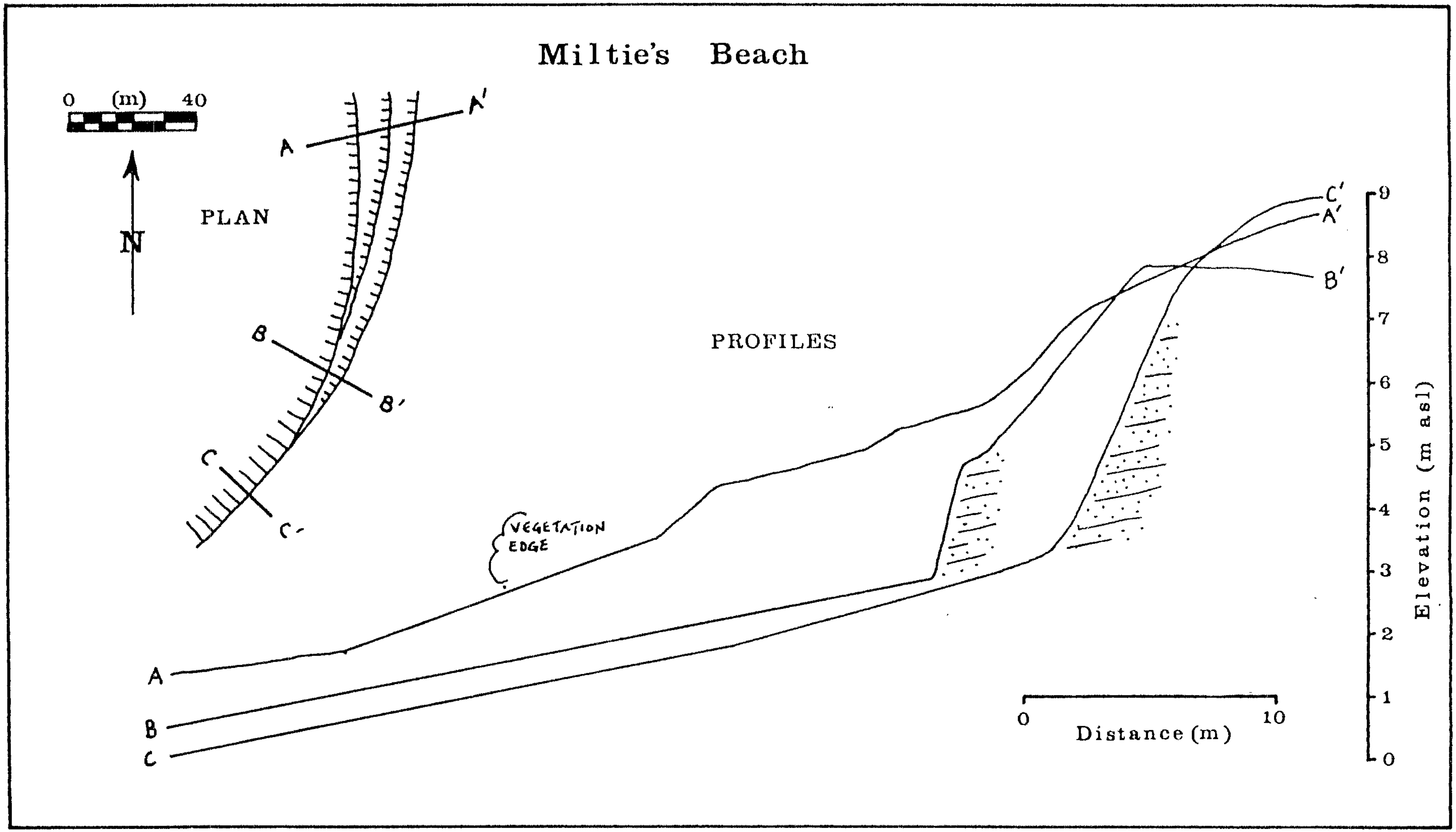


Figure 5.14: Three Topographic Profiles at Miltie's Beach.

Miltie's Beach

Description

This site is located on the north side of Vargas Island, on the prominent northwest facing beach (Figure 5.1). Three profiles (Figure 5.14) along this beach show the progressive erosion of shoreline features. In the northeast profile three scarp elevations are preserved; 3.5 m asl, 4.9 m asl, and 5.6 m asl. The top of the section is at 8.7 m asl. In the middle profile only the 4.9 m asl scarp is preserved, and in the southwest profile all the detail has been eroded away with the profile consisting of a steep scarp with the top at 8.7 m asl. The materials exposed in the section consist entirely of horizontally bedded medium sands.

Interpretation

This site is interesting in that it records a number of sea-level stillstands and it shows how the sea-level record can be progressively lost. The horizontally bedded sands exposed in the section at the southwest end suggest marine deposition related to a higher-high tide of at least 9.0 m asl. About 500 m inland behind this beach is a swamp from which basal peats might give a minimum age of abandonment of the 9.0 m elevation.

Kelsemaht Beach

Description

The Kelsemaht site is located at Yarksis Indian Reserve 11, once the traditional village of Kelsemaht, on the east side of Vargas Island (Figure 5.1). The Kelsemaht Village was occupied until at least 1920, and the Aht still followed a seasonal round until that time (Webster, 1983).

A total of 10 profiles were surveyed at Kelsemaht. Four of these profiles are presented to convey the detail from this site (Figure 5.15). These are the Kelsemaht Cave on the west side of the creek, and 3 profiles at the northeast end of the beach.

At the north-east end a series of marine features were recognized; a terrace remnant at 10 m asl, and marine scarps at ca. 8 m asl, 6 m asl, and 4 m asl. In transect 3, a prominent ridge and swale occurs with the swale elevation at 5.4 m asl on the landward side and 5.0 m asl on the seaward side with the ridge crest at about 6.0 m asl. This ridge and swale is truncated by the 4 m asl marine scarp. The Kelsemaht Cave is a prominent old sea-cliff with a raised cave and terrace at its base. The terrace is at 3.8 m asl, rising to 4.1 m asl where it meets the cliff. The maximum height of gravels in the sea cave is 4.8 m asl.

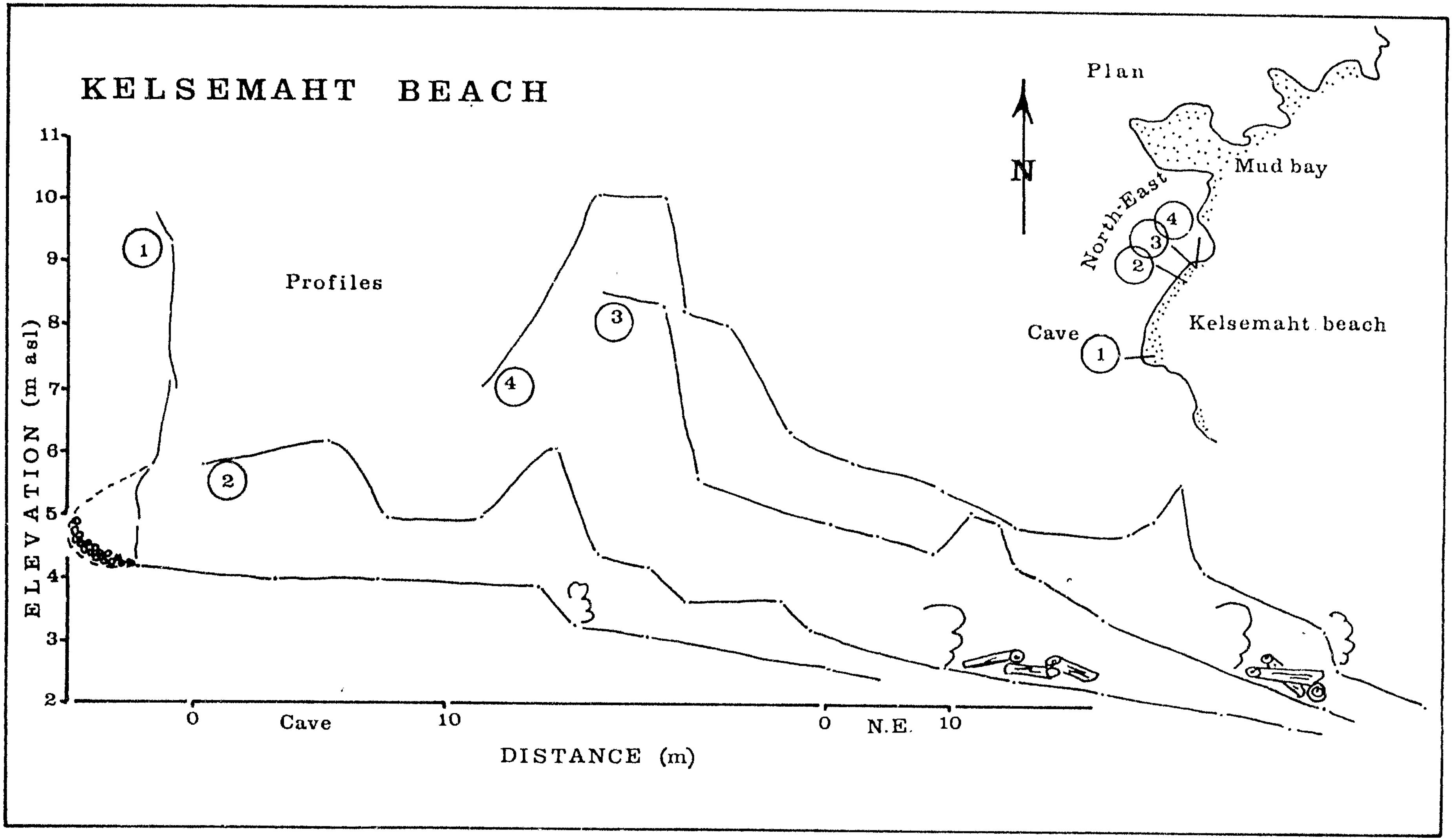


Figure 5.15: Four Topographic Profiles at Kelsemaht Beach.

Interpretation

The uppermost terrace at 10 m asl is the remnant of a wave-cut terrace but the associated scarp elevation was not surveyed. The next lower scarp is at about 8.0 m asl. The next level is the 6.0 m asl scarp in profile 2 and is associated with the upper ridge and swales of profile 3, between 5.5 and 6.0 m asl. These elevations are equivalent to the seaward elevations of the Ridge Traverse, just before the slopebreak to 4 m asl. Finally there is the scarp at 4 m asl which shows up in profile 1, 2, and 3 and at Kelsemaht Cave. A modern sea cave near the Kelsemaht Cave has a maximum height of gravels of 2.3 m asl, suggesting that the Kelsemaht Cave gravels formed when mean sea-level was 4.8 - 2.3 m asl, or 2.5 m asl.

Dick and Jane's Terrace

Description

This site is located near the mouth of the creek draining the northwest side of Vargas Island (Figure 5.1). The beach and back-beach at this site are very wide and are subject to sudden and dramatic erosion due to the combined effect of the creek in flood and winter storm surge (Dick & Jane, 1990; pers. comm.).

From the modern scarp at 2.6 m asl, the profile (Figure 5.16) rises abruptly to the terrace surface at 5.6 m asl. The terrace is about 40 m wide and gently sloping. It ends at a 6.3 m asl marine scarp. The top of this marine scarp is at 10.5 m asl. The ground inland is hummocky and poorly drained, and is underlain by sandy mud and diamict.

The material underlying the terrace is revealed in a road cut at the terrace edge and in a pit dug near the middle of the terrace. In the road-cut is a section about 1 m high and 5 m long. Exposed in it are medium sands with the occasional large pebble. In the pit the sands overlie a sandy cobble gravel with mussel shell. The sand/gravel contact was 1.35 m below surface. The gravels were 0.40 m thick a were lying on sandy mud.

Interpretation

This is a constructional terrace formed when the higher-high tide was roughly at 6.3 m asl. The mussel shell was not sampled so is not available for dating. This site is similar to the Ratcliffe Terrace except that the 4 m asl scarp has been removed by erosion.

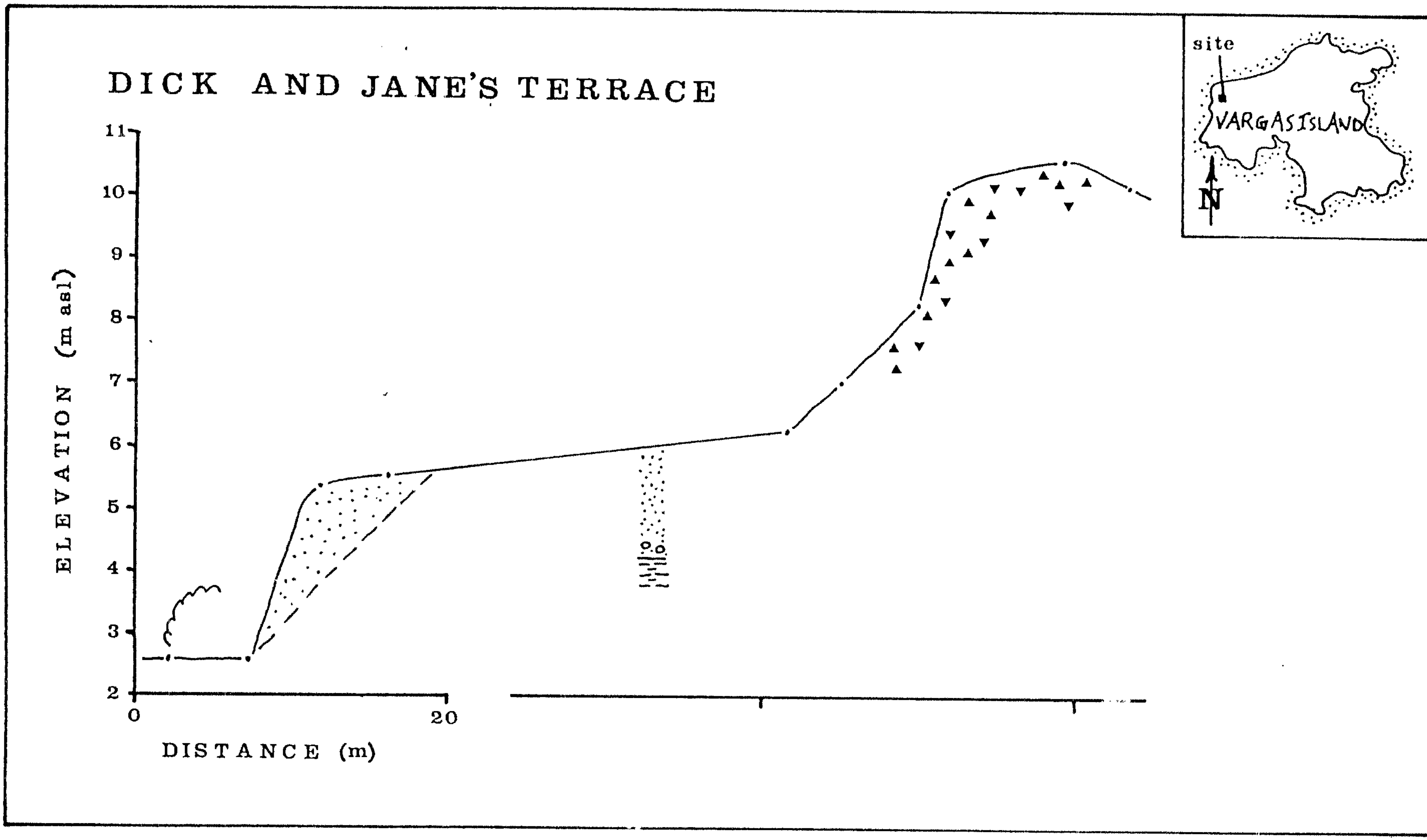


Figure 5.16: Topographic Profile of Dick and Jane's Terrace.

McIntosh Scarp

Description

This site is located at the often-used camp in McIntosh Bay (Figure 5.1). The feature (Figure 5.17) is a large marine scarp, at 5.2 - 5.6 m asl, with diamict exposed in the scarp wall. In the west profile is an additional prominent scarp at 3.8 m asl. The scarp crest is at 11.4 m asl in the west profile and descends to 8.9 m asl in the east profile.

Interpretation

The main marine scarp formed when the higher-high tide was approximately 5.4 m asl. Given that the modern marine scarp is at 2.4 m asl, mean sea-level, at the time of formation of this scarp, may have been about 3.1 m asl. By the same reasoning the lower scarp may have formed when mean sea-level was about 1.3 m asl.

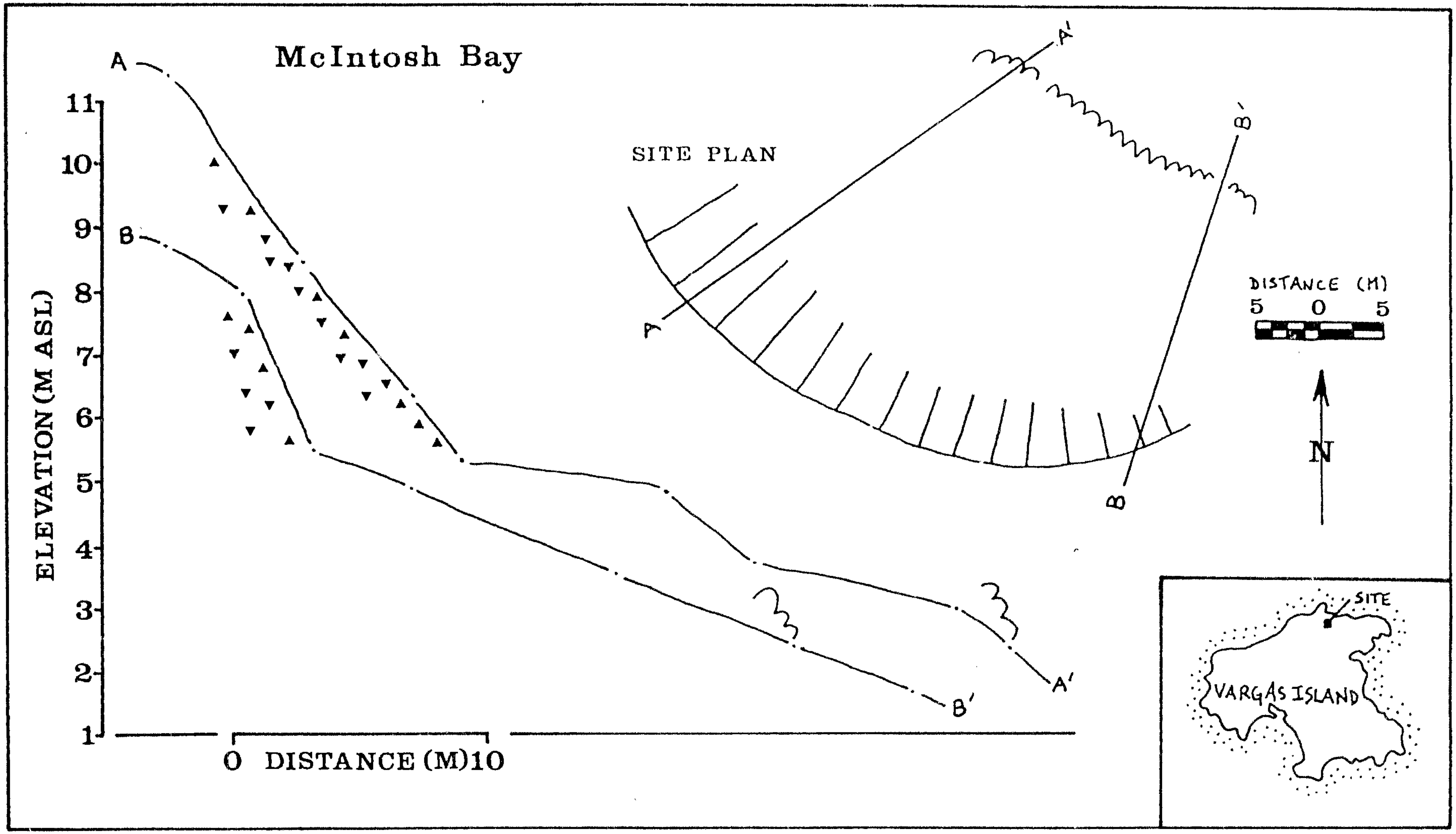


Figure 5.17: Two Topographic Profiles at McIntosh Bay.

CHAPTER 6: Discussion

Introduction: The Vargas Island Sea-level Curve

The sea-level curve developed from Vargas Island (Figure 6.1) extends the sea-level curve for the west-coast of Vancouver Island (Clague *et al.*, 1982) from about 4,000 yr BP to about 5,700 yr BP, and, further, to 7,900 yr BP if new dates on buried stumps from Grice Bay are included. Previous to this research two equally-likely (given available evidence) post-glacial relative sea-level scenarios existed: that of monotonic post-glacial emergence, or that of rapid post-glacial emergence to some early-Holocene lowstand followed by a Holocene submergence and emergence sequence. The significance of the Vargas Island sea-level curve is that it provides solid evidence for the latter scenario, with the mid-Holocene submergence and emergence cycle well constrained by eight new radiocarbon dates (Table A2).

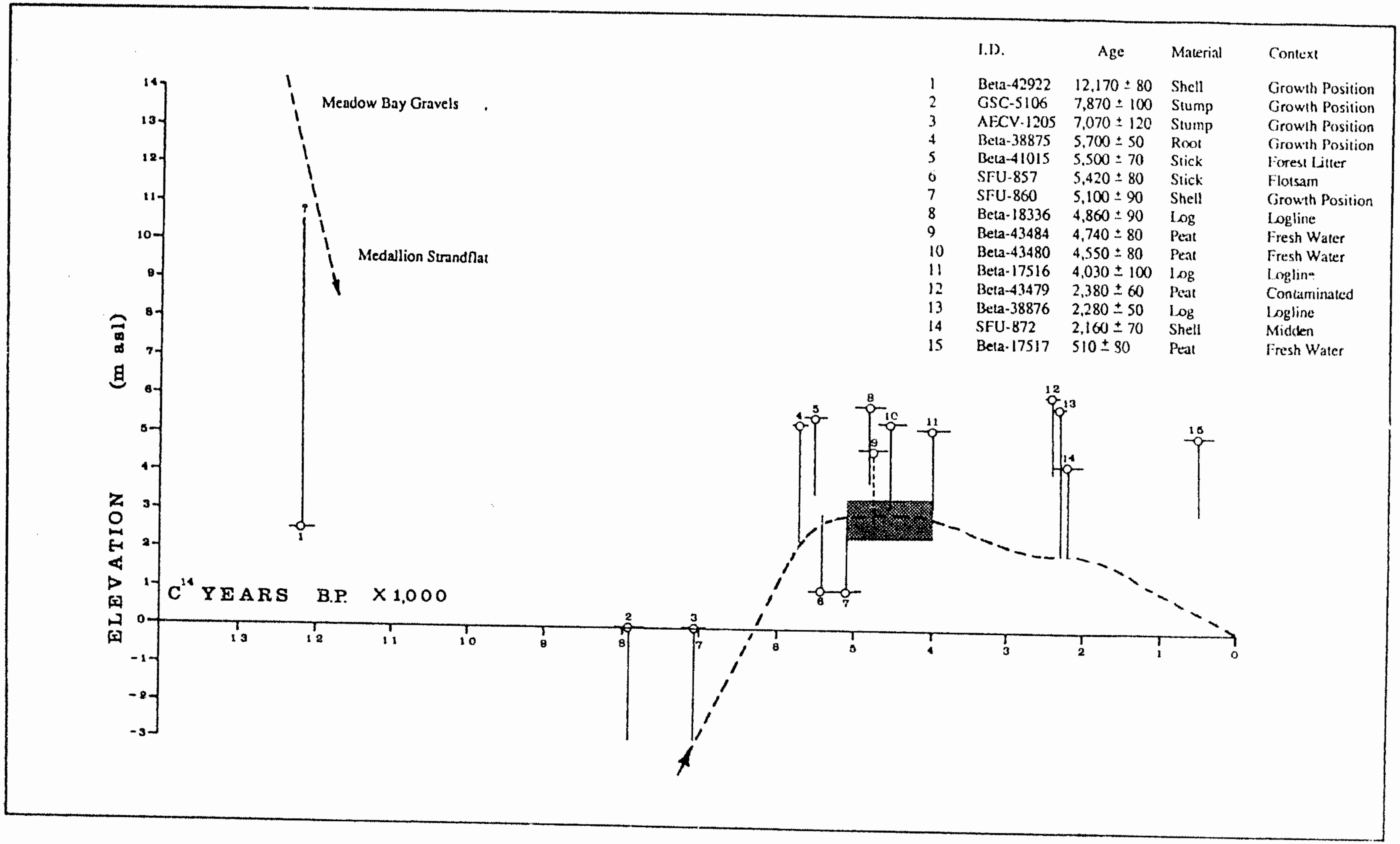
The character and phases of a sea-level curve result from the shifting dominance among geodynamic mechanisms. The Vargas Island sea-level curve can be considered in three parts: i) the early post-glacial emergence; ii) the early- to mid-Holocene submergence; and iii) the late-Holocene emergence. In the remainder of this chapter these three phases of the sea-level curve are individually discussed, then the regional applicability of the curve is examined. Finally, by testing the observed pattern against the pattern of sea-level change predicted by appropriate models from the literature, an attempt is made to resolve the geodynamic processes that led to this particular sea-level pattern.

Deglacial and Early Post-glacial Features

Post-glacial Marine Inundation

Deglacial sediments on the Estevan Coastal Plain are represented by fluvioglacial and marine deposits. Generally, the fluvioglacial deposits are confined to elevations at or above the marine limit (Howes, 1981c). Vargas Island is situated on the outer-coast and isolated from possible post-glacial meltwater sources, and, for the most part, is lower in elevation than the documented regional marine limit of about 30 m asl for Hesquiat Harbour (ibid), or 20 m asl for the outer-coast of the Brooks Peninsula (Howes, 1983b). Thus, deglacial sediments and landforms on Vargas Island are dominantly marine and consist of glaciomarine muds, beach sands and gravels, and wave-cut terraces.

Clague (1985) describes rapid destabilization of glaciers in the Hecate Lowland, near Prince Rupert, by eustatically rising water levels, with retreat by calving occurring so



	I.D.	Age	Material	Context
1	Beta-42922	12,170 ± 80	Shell	Growth Position
2	GSC-5106	7,870 ± 100	Stump	Growth Position
3	AECV-1205	7,070 ± 120	Stump	Growth Position
4	Beta-38875	5,700 ± 50	Root	Growth Position
5	Beta-41015	5,500 ± 70	Stick	Forest Litter
6	SFU-857	5,420 ± 80	Stick	Flotsam
7	SFU-860	5,100 ± 90	Shell	Growth Position
8	Beta-18336	4,860 ± 90	Log	Logline
9	Beta-43484	4,740 ± 80	Peat	Fresh Water
10	Beta-43480	4,550 ± 80	Peat	Fresh Water
11	Beta-17516	4,030 ± 100	Log	Logline
12	Beta-43479	2,380 ± 60	Peat	Contaminated
13	Beta-38876	2,280 ± 50	Log	Logline
14	SFU-872	2,160 ± 70	Shell	Midden
15	Beta-17517	510 ± 80	Peat	Fresh Water

Figure 6.1: The Vargas Island Sea-level Curve. Radiocarbon ages: i) are based on the Libby half-life of 5568 yrs; ii) are corrected for the sea water reservoir effect, where applicable; and iii) show quoted errors from the respective labs. Sea-level indicators are relational, formed above or below sea-level, with the mean sea-level estimate associated with each date derived from modern analog data (see text). Shaded area indicates 2.4 to 3.4 m asl paleo-datum estimate for Ahous Bay Stillstand based on precise survey at Ahous Lagoon Measured Section.

quickly that very little drift material was left behind. A similar deglacial scenario may have occurred in the outer-coast of Clayoquot Sound with fjord glaciers retreating rapidly up Tofino, Bedwell, Herbert, Shelter, and Sydney Inlets (Figure 1.2).

On Vargas Island, the post-glacial marine inundation has been dated at 12,170 ± 80 yr BP (Beta-42922; corrected) from in-situ shell. This shell fauna, containing *Hiatella arctica* and *Mytilus edulis,* among other species, is within a stoney mud at the base of a laminated sand and mud sequence. These glaciomarine materials overly a compact diamict, interpreted as a glacial diamict, probably lodgement till. Thus, deglaciation of the outer-coast began some 12,000 to 13,000 years ago as indicated by this date, and a date of 13,000 yr BP on wood contained within similar marine muds in Hesquiat Harbour (Howes, 1981c).

The blue-grey laminated sands and muds, described in units 2 and 3 from the Well Core at Buckle Bay, and underlying the Ridge System, South Bog, Ratcliffe Terrace, and Dick and Jane's Terrace are similar to the massive moderately compact blue-grey silts and sandy silts described as the typical offshore glaciomarine sediments for the region by Howes (1981c). These are similar to sand and mud lithofacies described by Powell (1981); sandy or silty laminae are due to underflows or interflows from ... "subglacial meltwater discharges when sediment loads are sufficiently high" (p 129). The paucity of dropstones in these units is conspicuous, although at least one angular clast was noted in core, and a large clast was noted in the exposure at Ahous Lagoon. Vibracoring of similar blue-grey muds on the Esowista Peninsula has revealed the occurrence of erratic granitoid dropstones (personal observation), so the glaciomarine origin of these muds is fairly certain. The scarcity of dropstones in Vargas Island sand and mud units may indicate shallow water conditions, with the ice-berg track confined to the deeper channels. In addition, the paucity of dropstones might be due to the rapid retreat and possible grounding of glaciers upfjord. The fining upward of units 1 to 3 at Buckle Berm is interpreted as a shift from an ice proximal to an ice distal condition, and supports this retreat hypothesis.

It is believed that the marine limit in the vicinity of Vargas Island is lower than 32 - 34 m asl, and is probably closer to the marine limit of 19 - 20 m asl for the Brooks Peninsula (Howes, 1983b). Preliminary observations of a prominent slope-break at Radar Hill, on the Esowista Peninsula, suggest that the marine limit may be at 15 m asl in that vicinity (Hutchinson, 1991; pers. comm.). On Vargas Island, the highest observed marine shoreline was represented by heavily oxidized, well rounded beach gravels with an immature matrix. This gravel formed a constructional terrace with a surface elevation of 13.5 m asl.

Post-glacial Emergence

Howes (1981c), working in the Hesquiat area, suggested that ... "marine features above 8 m [above the high tide limit] formed during the last deglacial sequence, whereas those below this height were formed during the Holocene. Marine landforms above 8 m have a continuous vertical distribution; lower landforms occur as a series of steps. The former suggests that sea-level fell rapidly and/or relatively continuously from the marine limit. These higher marine features probably formed during and within a few thousand years of deglaciation when major land-sea shifts were primarily controlled by isostatic and eustatic responses to shifting and dissipating ice loads."

Emergent shoreline features on Vargas Island above about 6 m asl, the highest dated Holocene strandline, are terrace remnants at 13.5 m asl and 10.0 m asl, and strandlines at 9.5 m asl and 8.0 m asl. The most prominent feature is the Medallion Strandflat, a wave-cut terrace with associated strandline at 9.5 m asl. This feature might be associated with horizontally bedded marine sands to about 8.7 m asl at Miltie's Beach. These higher littoral markers appear to be continuously distributed (i.e., no features identified have coincident elevations). Around Vargas Island the topography between 15 m asl and 6 m asl generally reflects bedrock with a blanket of overlying till, and marine mud in topographic lows. The sand and gravel lag representing isostatic emergence (Boulton *et al.*, 1982) was not widespread above 6 m asl. This might reflect a lapse in observation, but it also may suggest that emergence was fairly rapid without much time for littoral modification of emerging topography.

It is unknown how far sea-level fell during this emergent phase, but the occurrence of a forest floor and associated soil development on glaciomarine muds at the Buckle Berm site suggest that mean sea-level fell below at least 3.0 m asl. In-situ stumps rooted on glaciomarine mud at 0 m asl in Grice Bay (Figure 1.2) suggest that mean sea-level fell to least -3 m asl sometime before about 7,870 ± 100 yr BP (GSC-5106). Marine muds overlying terrestrial peat indicate that sea-level was about -9 m asl about 8,000 yr BP in Barkley Sound (Hebda, 1991; pers. comm.).

Early- to Mid-Holocene Submergence

The Evidence for Holocene Submergence

As mentioned earlier, two alternative scenarios for the post-glacial sea level curve exist: i) post-glacial monotonic emergence, or ii) post-glacial emergence to some lowstand, lower than the mid-Holocene level, followed by a submergence/emergence sequence. Stratigraphic evidence was found at two sites on Vargas Island and one in Grice Bay indicating the occurrence of an early- to mid-Holocene submergence. With documented emergence for the late-Holocene, this submergence represents the rising limb of a submergence/emergence cycle, supporting the latter scenario. The height of the Holocene submergence, a lengthy stillstand, is represented by a well developed and regionally extensive strandline at 6.0 m asl. The 6 m strandline is represented on Vargas Island by the Buckle Berm and Ratcliffe Terrace (ca 6.0 - 6.5 m asl), and by marine scarps bounding the Ridge System and South Bog (ca 5.5 - 6.5), Miltie's Beach (ca 5.6 m asl), Kelsemaht Beach (ca 6.0 m asl), McIntosh Scarp (ca 5.4 m asl), and Dick and Jane's Terrace (ca 6.3 m asl).

Holocene submergence is constrained by the death-ages of buried stumps in Grice Bay and a buried root mat at the Buckle Berm site. Considering the modern marine influence extending up to 3 m asl, the death-ages of 7,870 ± 100 yr BP (GSC-5106) and 7070 ± 120 yr BP (AECV-1205) for two Grice Bay stumps, rooted at about 0 m asl, indicate that mean sea-level in Clayoquot Sound had risen to about -3 m asl by about 7,000 yr BP. The death-age of 5,700 yr BP for the Buckle Berm roots, rooted at 5.3 m asl, indicate that mean sea-level had risen to about 2.3 m asl by 5,700 yr BP. The burial of the Buckle Berm forest floor by marine sediments after 5,500 yr BP indicates that mean sea-level had exceeded 2.3 m asl shortly thereafter.

Further, in the Medallion subbasin macrofossil cores, the transition from a mesic forest/swamp assemblage at 2 m asl to an assemblage with aquatic indicators at 4.5 m asl is taken to indicate an elevation of the local water table driven by rising sea-level; conversely, if monotonic emergence had taken place, gyttja would be expected at depth. The forest-litter spikes (spruce needles and *Pinaceae* seeds), and incursions of beach sand at ca 4.0 to 5.5 m asl are the result of the closeness of a beach to the site at the height of submergence. This paleo-Medallion Beach had formed by 4,550 years ago and incursions of sand into the subbasin continued until after this time, an age given by fibrous peat sandwiched between the beach gravel below and sand above.

Datum Estimates for the 6.0 m Strandline

Independent estimates of mean sea-level represented by the 6 m strandline were made from various features at four different sites on Vargas Island (Table 6.1). These estimates of paleo-mean sea-level represented by the 6 m strandline fall between 2.55 and 3.8 m asl. The 3.8 m asl estimate is on wood (sample Beta-18336) that lies on top of sand on the Ridge Traverse. If this sample represents logline wood then it is likely that the log was deposited on the foredune, and is therefore most likely an over-estimate of paleo-datum. The remaining estimates fall between 2.4 and 3.4 m asl, showing considerable consistency.

Table 6.1: Paleo-datum estimates and ages for the 6 m strandline on Vargas Island

Site	Feature	Estimate (m asl)	Age (yr BP)	Quality
Ridge Traverse	foredunes	2.55 - 3.25	- ? -	imprecise
Ridge Traverse	logline	3.8	< 4,860	over estimate
Ridge Traverse	logline	3.2	< 4,030	reasonable
Ahous Lagoon	logline	2.4 - 3.4	< 5,100	confident
Medallion	raised beach	3.2	~ 4,550	reasonable
Medallion	subbasin sill	3.0	~ 4,550	reasonable
McIntosh	scarp	3.1	- ? -	reasonable

Analog estimates from modern foredune swale elevations at Hopkin's Beach, the Dune Beach, and Ahous North suggest the paleo-datum stood between 2.55 m asl and 3.25 m asl. The estimate derived from comparison of the Medallion Raised Beach with the modern Stubbs Island Beach suggests the paleo-datum stood at 3.2 m asl. The 5.4 m asl scarp at McIntosh Bay yields a paleo-datum estimate of 3.1 m asl. Also, given the lowest known sill elevation in the Medallion subbasin of 5 m asl, the fresh-water conditions indicated by *Potamogeton* in core indicate that the higher-high tide did not exceed about 5 m asl during the stillstand, that is the paleo-datum never exceeded about 3.0 m asl.

However, the most confident estimate of paleo-datum for the 6 m strandline is 2.4 - 3.4 m asl from the Ahous Lagoon Measured Section. This survey was a closed traverse with the NK-1 level, and was tied to the lagoon tide gauge, which has an uncertainty of ± 0.10 m and does not suffer positive bias inherent with tideline estimates. The main source of error was the association of the assumed logs with the paleo-logline.

Paleo-Environmental Reconstruction and the History of the Ridges

The stillstand that formed the 6 m strandline, constrained by 8 new radiocarbon dates, lasted from just after 5,500 yr BP to at least 4,030 yr BP. During the early part of the stillstand, sometime after 5,500 yr BP, the sea would have stood at, or washed over, the paleo-tombolo connecting the uplands of northern and southern Vargas Island. Also, there is stratigraphic and morphological evidence that the South Bog was a tidal lagoon during this early part of the stillstand, but a date of 4,740 yr BP on basal fresh-water peat at SB-3 indicates the isolation of the paleo-lagoon sometime before then. This isolation, leading to an elevated water-table in South Bog, is indicated by the transition from nearshore alder stands to the progressive development of a pine bog at SB-3. A date from peat at C-11 in the Medallion subbasin indicates that the sea washed at the Medallion Raised Beach until after 4,550 yr BP. Radiocarbon sample Beta-17516 suggests that the brunt of ridge progradation did not occur until after 4,030 yr BP.

Within the framework of these dates, the 5,100 yr BP age on shell near the base of the Ahous Lagoon Measured Section can be interpreted. If, near the end of submergence, some 5,500 years ago, the sea washed the paleo-tombolo, Ahous Bay would have been a broad wave-cut platform. The upper erosional contact of the glaciomarine mud at 1 m asl in the Ahous Lagoon Measured Section, 430 m inland, indicate, that, with a mean tide level of about 3.3 m asl and a 4 m tidal range, this platform may have been exposed at low spring tides. Thus, the shell fauna, wood, cones, and pebble lag from the base of the Ahous Lagoon Measured Section represent a lower intertidal deposit. This intertidal condition is indicated by species found in-situ, such as Bodega Tellin, Chubby Mya, worm tubes, and Purple Olive. Thus, the paleo-beach advanced across the platform to the position about 430 m inland from the modern beach only after 5,100 yr BP.

Two maximum dates for the age of the Ridge System were obtained by Rogers *et al.* (1986). These are 4,860 yr BP (Beta-18336), 1,320 m inland and 4,030 yr BP (Beta-17516), 1,975 m inland. These dates are contradictory in that the oldest date (Beta-18336) should be farther inland. Thus, either sample Beta-18336 is too old or sample Beta-17516 is too young. The simplest explanation has Beta-18336 representing older reworked logline wood, implying that the majority of ridge progradation did not occur until after 4,030 yr BP.

Late-Holocene Emergence

Although the timing of the onset of emergence is not well constrained, sea-levels began to fall sometime after 4,030 yr BP, but before 2,760 ± 80 (I-9697) yr BP (from Village Lake; Hebda and Rouse, 1979). Further, it is difficult to assess the character of late-Holocene emergence (i.e. episodic or continuous). The 4 m strandline, evident as a scarp on the Medallion Traverse (ca 4.0 m asl), Kelsemaht Cave (ca 4.1 m asl), McIntosh Bay (ca 3.8 m asl), and the Ratcliffe Terrace (4.0 m asl), and as a major slope break on the Ridge Traverse, is a prominent erosional feature suggesting a stillstand, or reversal, during emergence. During this stillstand or minor reversal, considerable loss of the sea-level record between about 6 m asl and 4 m asl might have occurred. Thus, the assessment of the nature of sea-level change between 4,000 yr BP and 2,200 yr BP may be difficult. On the contrary, the gradual slope between the 4 m asl scarp and the modern beach seen on the

Buckle Berm profile, the Ridge Traverse, Medallion/South Bog Traverse, and Kelsemaht Beach suggest relatively continuous emergence from the 2,200 yr BP to present. This is a tentative conclusion which requires further substantiation.

Three estimates of mean sea-level represented by the 4 m strandline were made (Table 6.2). At Buckle Bay a midden deposit resting on extremely well sorted pebble gravels is identical in character to modern spring tide pebble berm gravels on Buckle Beach. The elevation of the modern gravels indicate that the raised gravels were deposited by a 2 m asl stand of mean sea-level. Midden shell gave a corrected radiocarbon age of 2,160 yr BP, thus this shoreline is about 2,160 years old, or younger. Comparison of the Medallion Scarp with a similar, but modern scarp at Moser Point Reserve Beach suggests that Medallion Scarp formed when mean sea-level stood at 2 m asl. A log associated with storm tossed gravels in the Medallion Scarp indicate a maximum age of 2,280 ± 50 yr BP for this feature. Comparison of the height of cave gravels in a Kelsemaht Cave with gravels in a nearby shallow cave at the back of a wave-cut platform suggest that the 4 m asl strandline represents a 2.5 m asl stand of the sea. Thus, the Medallion and Buckle sites suggest that by 2,200 yr BP sea-levels were at or below about 2 m asl.

Table 6.2: Paleo-datum estimates and ages for the 4 m strandline on Vargas Island

Site	Feature	Estimate (m asl)	Age (yr BP)	Quality
Buckle Bay	springtide berm	2.0	≥ 2,160	reasonable
Medallion Scarp	emergence scarp	2.0	< 2,280	reasonable
Kelsemaht Cave	sea cave	2.5	- ? -	- ? -

Regional Extent of Submergence/Emergence Cycle

Three dates on water worn artifacts from a storm berm at Yuquot, 4.5 m above the high tide line, are the oldest Holocene littoral dates previously recovered for the west-coast (see Clague *et al.* 1982; fig 10). The dates (3,590 ± 190 [GSC-1767], 4,080 ± 80 [GaK-2179], and 4,230 ± 90 [GaK-2183]) suggested to Howes (1981c) that the high water level 4,000 years ago was probably 2.5 to 3.5 m higher than present. This 2.5 m asl to 3.5 m asl mean sea-level stand about 4,000 years ago coincides with the mean sea-level estimate of 2.4 m asl to 3.4 m asl, 4,030 yr BP for Vargas Island.

My own observations in Boat Basin revealed a prominent wave-cut terrace with a surface elevation of 6.3 m asl to 7.3 m asl. This survey was carried out with clinometer and tape, and was tied to Tofino predicted water, so the precision and accuracy is unknown. The survey profile was located at the site of the old Boat Basin Post Office. Airphotos of the area (series BC80073, 190 to 204) show a distinct vegetation signature on this wave-cut terrace that extends all the way around the Hesquiat Peninsula and Hesquiat Harbour. On Branch 70 of the Stewardson mainline road, in Boat Basin, the marine scarp is clearly visible in an extensive clear-cut (NTS sheet 92E/8; GR 887815).

Village Lake on the Hesquiat Peninsula is situated on this wave-cut terrace and has an outlet elevation of 3 m asl (Clague *et al.*, 1982). A date of 2,760 ± 80 (I-9697) on shelly gyttja in a core from Village Lake indicates upper-tidal influence at or above 3 m asl 2,800 years ago, providing a minimum age for the wave-cut terrace. A 6 m strandline also occurs on the south-west side of Flores Island (personal observation). I would argue that the Boat Basin and Flores Island strandlines are contemporary with the 6 m strandline on Vargas Island. Finally, recent work in Barkley Sound indicates that a submergence/emergence cycle of similar magnitude and timing occurred south of Clayoquot Sound (Hebda, 1991; pers. comm.).

These data strongly suggest that the submergence/emergence cycle described from Vargas Island can also be documented from as far north as Yuquot, in Nootka Sound, and as far south as Barkley Sound. This submergence, with the associated stillstand spanning 5,100 yr BP to 4,030 yr BP, eroded littoral features of the early-Holocene, and explains the lack of littoral dates from this time period for the central west coast of Vancouver Island.

In order to distinguish the early- to mid-Holocene submergence from the post-glacial marine inundation, this period of sea-level rise and coastal erosion will be referred to as the *Clayoquot Transgression*, an event of regional significance. The associated highstand, lasting at least 1,100 years, will be referred to as the *Ahous Bay Stillstand*, as it is so saliently represented in the landscape by the Ahous Bay beach ridges.

Consideration of Factors Forcing Holocene Sea-level Change Vargas Island

As discussed in Chapter 3, distinguishing the dominant factors (glacio-isostasy, glacio-eustasy, and tectonics) forcing a particular history of sea-level change must involve developing models against which the observed sea-level pattern can be tested.

Isostasy: Comparison with Other Marginally Glaciated Regions

It is generally accepted that from the early post-glacial to about 8,000 yr BP isostatic rebound was the dominant geodynamic process acting in southern coastal British Columbia (Andrews, 1989).

Southern coastal British Columbia experienced late-Pleistocene emergence followed by submergence from the early-Holocene onward. This is the typical Zone I/II response of Clark *et al.* (1978). The west-coast of Vancouver Island is also within the transition Zone I/II as indicated by the early- and mid-Holocene pattern of emergence followed by submergence. Yet, the emergent phase of the late-Holocene is anomalous, and has been ascribed to tectonic forcing by most authors (Clague *et al.*, 1982; Riddihough, 1982; and others).

Thus, the mid-Holocene submergence/emergence cycle described for the west-coast of Vancouver Island invites speculation regarding potential driving mechanisms. This submergence/emergence cycle has no similar counterpart elsewhere in Canada: along the Arctic and Maritime coasts of Canada the Zone I, Zone I/II transition, and Zone II patterns are observed depending on the position of the site relative to the ice margin (Clark *et al.*, 1978; Quinlan and Beaumont, 1981; Andrews, 1989; and references therein).

Interestingly, a similar submergence/emergence cycle occurs in both Norway and Spitsbergen, areas along a passive continental margin. The submergence/emergence sequences (similar in magnitude, but varying in timing) that occurred in these areas are documented only for extreme outer-coast, marginally glaciated regions in settings very similar in character to the position of the outer-coast of Vancouver Island. At this point it is pertinent to note that Clark *et al.* (1978) stated: ... "[The post-glacial relative sea-level] predictions for localities [in ice marginal areas] are the most in error, suggesting that slight modifications of the assumed melting history and/or the rheological model of the earth's interior are necessary" (p 265).

In south Norway (Hafsten, 1983; & references therein), the mid-Holocene submergence, referred to as the Tapes Transgression, occurs at all outer-coast sites (Alesund, Austrheim, Sotra, Bomlo, and Jaeren), while sites in the Oslo Fjord region, about 200 km east, show the characteristic Zone I emergence. The Tapes Transgression is indicated by strandlines to 11 m asl, and occurred between 7,000 and 5,000 yr BP (Hafsten, 1983). These outer-coast areas are currently stable or experiencing slight emergence.

In Spitsbergen (Forman, 1990: & references therein), the mid-Holocene submergence, referred to as the Talavera Transgression, occurs in the outer-coast sites near Isfjorden, while sites to the north, south, and east show Zone I emergence. The Talavera

Transgression is indicated by strandlines no higher than 7 m asl, and occurred between 6,500 and 5,000 yr BP. These outer-coast areas are currently stable or experiencing slight submergence.

Thus, the similarity of the essential pattern (not the timing) of the Clayoquot Transgression with the Tapes and Talavera Transgressions (and associated late-Holocene emergence events), suggests that the mid-Holocene submergence/emergence documented for the west-coast of Vancouver Island may have arisen by some residual glacio-isostatic mechanism such as forebulge collapse (Clark *et al.*, 1978; Quinlan and Beaumont, 1981). Forman (1990) notes: "The cause of the oscillation is unknown, but it might be the result of back migration of glacio-isostatically displaced mantle material" (p 1587). Along these lines Fairbridge comments, ... "it is evident that a transgression such as the Flandrian is not a universal or synchronous phenomenon. In tectonically subsiding areas such as a collapsing marginal bulge ... the transgression may continue [beyond 6,000 yr BP] when the rest of the world is in a regressive mode" (Fairbridge, 1983; p. 219).

Eustasy: Holocene Eustatic Oscillations

Global post-glacial eustatic sea-level rise is generally considered to have ceased by 6,000 yr BP. Thus, if the latter stages of the Clayoquot Transgression, with sea-levels about 3 m higher than today, and/or the late-Holocene stillstand represented by the 4 m strandline were driven by a eustatic process then the timing of these events should correlate with inferred Holocene eustatic fluctuations documented by Mörner (1976), for the North Sea, and Fairbridge and Hillaire-Marcel (1977), for the Hudson's Bay region. Before engaging in such a comparison the theoretical problems of using data derived in such distant areas is recognized, but given that these are supposedly eustatic, ocean-wide changes of sea-level, we can expect at least the timing of such events to coincide if indeed they are driven by similar mechanisms.

The Clayoquot Transgression maximum (Ahous Bay Stillstand) has been constrained to between about 5,100 yr BP and 4,000 yr BP (radiocarbon years) and coincides with Mörner's post-glacial regression maximum, a lowstand known as PR-5 (Mörner, 1976), and a regression maximum indicated by Fairbridge and Hillaire-Marcel (1977), suggesting that the Ahous Bay Stillstand was not likely linked to worldwide changes in ocean volume. On the other hand, the age of the 4 m strandline is coincident to that of a transgression maximum about 2,200 yr BP (Fairbridge and Hillaire-Marcel, 1977). This eustatic rise might explain the prominence of the 4 m strandline, with its abandonment linked to the onset of eustatic sea-level fall.

Assessment of Possible Tectonic Forcing

If subduction along the Cascadia subduction zone off Vancouver Island is seismic, then evidence for jerky uplift might be present in the geological record of the Estevan coastal plain. Geological observations from southern Vancouver Island and the Lower Mainland to date suggest stability for the last 2,000 years (Clague and Bobrowsky, 1989). In contrast, work in Washington and Oregon suggests jerky subsidence (Atwater, 1987; Peterson and Darienzo, 1988 & 1991; Atwater and Yamaguchi, 1991).

The emergence of the west coast of Vancouver Island for the last 4,000 years counters the regional trend of slight submergence, or stability, and has been ascribed to tectonic forcing by most authors. No obvious co-seismic signature was discovered on Vargas Island. In similar extreme out-board areas such as Middleton Island, Alaska, co-seismic emergence has amounted to 40 m over 4,500 years, but closer to the hinge line, post-seismic recovery may offset co-seismic uplift to produce a net emergence of 1 m over 4,500 years (Plafker, 1990). Most uplift rates on active subduction zones fall between 0.5 and 3.0 mm/yr (Lajoie, 1987), and the 3 m of emergence over 4,000 years on Vargas Island is within these known rates. Regardless, in the creeks and sloughs of Vargas Island and surrounding islands, no evidence was noted for recurrent uplift/subsidence phenomenon and Hebda and Rouse (1979) described a gradual isolation of Village Lake on the Hesquiat Peninsula, a basin with 3 m asl outlet sill elevation, for the period after 2,800 yr BP.

A more rigorous test of co-seismic forcing is to compare the timing of the "probably", co-seismic submergence events reported from western Washington and Oregon with the initial emergence and eventual abandonment ages of the 6 m strandline and the abandonment age of the 4 m strandline on Vargas Island. If these ages are synchronous, then these features might be related to the same seismic events, or rupture sequences.

Golder Associates (1988) synthesized the radiocarbon ages (about 100) reported by Atwater (1988) for 8 buried marsh horizons found in estuaries along the Washington Coast from Neah Bay to the Columbia River. The mean death-age of Surface-4 equalled 2,360 yr BP and broadly coincides with the abandonment of the 4 m strandline on Vargas Island. The mean death-age of Surface-8 equalled 4,290 yr BP and falls in the age range of the Ahous Bay Stillstand. Peterson and Darienzo (1991) reported 10 radiocarbon dated marsh top (MT) ages (synonymous with death-age) for Alsea Bay on the central-coast of Oregon. Their MT-6 gave an age of 2,210 ± 80 yr BP (Beta-27185) again coinciding with the abandonment age of 2,200 yr BP for the 4 m strandline on Vargas Island. Their MT-10, which gave an age of 4,510 ± 80 yr BP (Beta-26790), does not coincide with initial emergence or abandonment of the 6 m strandline (i.e., 5,100 & 4,000 yr BP), but the age is not asynchronous with the age range of the Ahous Bay Stillstand.

It is acknowledged that the assumption that these buried surfaces represent co-seismic subsidence events is controversial and alternatively explained by Holocene eustatic fluctuations; if they do indeed represent co-seismic events, then the return period and thickness of estuarine infilling preceding an event can be viewed as an indices for event magnitude. The larger an event then the more likely it was that plate segments ruptured together, and therefore the the more likely is the possibility that such an event would appear in the geological record of south-coastal British Columbia. The return period of Surface-8 and MT-10 cannot be assessed because these are the deepest visible surfaces, so a sense of the magnitude of these events (or event) is not available. The return period, and infilling, preceding Surface-4 and MT-6 are less than the average and might be considered about medium to small sized events.

The broad correlation of the marsh top ages from Washington and Oregon with the radiocarbon ages of the 6 m and 4 m strandlines on Vargas Island does not confirm co-seismic uplift of the strandlines on Vargas Island, but it does not allow this possibility to be ruled out. Furthermore, the potential paleo-seismic record from the west-coast of Vancouver Island cannot be resolved without further detailed research targetting landforms between 7 m asl and present mean water.

Late-Holocene and Contemporary Emergence Rates

With the uncertainty of defining exactly when emergence began after the Ahous Bay Stillstand, only two dates can be used for an estimate of long term emergence rates. These are the two constraining dates on the 4 m strandline. An age on wood of 2,280 ± 50 (Beta-38876) and an age on shell of 2,962 ± 70 (SFU-872) both suggest that mean sea-level stood at about 2 m asl about 2,200 radiocarbon years ago. This yields a long term emergence rate of 0.9 mm/yr. Given the uncertainties in sea-level curve construction and uncertainties in calculating contemporary emergence from tidal data (Clague *et al.*, 1982; Andrews, 1989), it is difficult to compare this long term rate of emergence with the contemporary rates calculated from the tide gauge records at Tofino. Yet, the rate of 0.9 mm/yr for long term emergence is 28% to 44% smaller than the calculated contemporary rates from Tofino of 1.25 mm/yr (Wigen and Stephenson, 1980) and 1.6 mm/yr (Vanicek and Nagy, 1980). These higher rates for contemporary emergence might be statistical artifacts or they might represent pre-seismic crustal strain accumulation (Fitch and Scholz, 1971).

CHAPTER 7: Conclusions and Future Research

Summary and Conclusions

The main result of this research is in providing a sea-level curve for Vargas Island which documents an early- through mid-Holocene sea-level rise (Figure 6.1). This Holocene submergence is shown to be an event of regional significance and it is formally proposed that it be given the name the *Clayoquot Transgression* to distinguish it from post-glacial marine inundation.

The highstand of the Holocene submergence, constrained with eight new radiocarbon dates that indicate a lengthy stillstand, is represented in the landscape by the Vargas Island progradational beach ridges parallel to the modern Ahous Bay. It is formally proposed that this lengthy (about 1,100 years) stillstand be referred to as the *Ahous Bay Stillstand*.

The Marine Limit and Early Post-glacial Emergence

Post-glacial marine inundation has been dated at 12,170 yr BP on Vargas Island from shell fauna containing in-situ *Hiatella arctica* and *Mytilus edulis*. Deglacial sediments are represented by blue-grey laminated sands and muds, which were seen to fine upward indicating a shift from ice proximal to ice distal conditions.

The highest post-glacial sea-level identified was a heavily oxidized, sandy gravel lag deposit at 13.5 m asl. Sometime during the early post-glacial, sea-level fell to below at least -3.0, as indicated by 7,900 year old, in-situ tree stumps in Grice Bay. Post-glacial strandlines are represented by wave-cut and constructional terraces between 13.5 and 7 m asl. These features appear to be distributed continuously with elevation, indicating rapid and/or steady emergence.

The Clayoquot Transgression

Evidence from Vargas Island shows that a mid-Holocene submergence/emergence cycle occurred in Clayoquot Sound. Paleomean sea-level was lower than about 3 m below mean sea-level about 7,000-8,000 years ago as indicated by buried stumps in Grice Bay. By 5,700 yr BP mean sea-level had risen to about 2.3 m asl, as indicated by the death-age of a tree rooted on emergent glaciomarine mud at Buckle Bay. The height of submergence, the Ahous Bay Stillstand is indicated by prominent marine features with strandlines at about 6 m asl. Mean sea-level during this highstand, estimated to fall between 2.4 m asl to 3.4 m asl, was attained sometime between 5,500 yr BP and 5,100 yr BP, and did not start to decline until after about 4,030 yr BP. Thus, the Ahous Bay Stillstand likely lasted at least 1,100 years.

During the Ahous Bay Stillstand the Vargas Island progradational beach ridges formed, with the brunt of ridge advance occurring shortly after 4,030 years ago. Other landforms that formed during the stillstand were numerous prominent marine scarps, a storm berm (Buckle Berm), sandy platforms (Dick and Jane's Terrace, Ratcliffes Terrace), and a gravel beach (Medallion Raised Beach). This stillstand is marked by raised littoral features documented from Nootka Sound to Barkley Sound, and is the most prominent shoreline feature on the central west-coast of Vancouver Island, suggesting that the Holocene submergence and stillstand were of regional significance.

Late-Holocene Emergence

The nature of late-Holocene emergence is poorly understood. The main late-Holocene emergent feature, a marine scarp at 4 m asl which occurs behind Medallion Beach, Buckle Beach, Hovelaque's Beach, and Cow Bay on Flores Island, suggests a stillstand or reversal during emergence. The development of this prominent scarp has eroded much of the topographic signal representing the period from 4,000 yr BP to 2,200 yr BP, yet the gradual profile from 2,200 yr BP to the present beach might be an indicator of more or less continuous emergence for the last 2,200 years. Estimates from the 4 m strandline suggest that 2,200 years ago mean sea-level was about 2 m higher than today.

Forcing Mechanisms

The geodynamic context of the observed sea-level pattern on the west-coast of Vancouver Island is difficult to resolve. In chapter 6 the pattern was compared to models of isostatic (Clark *et al.* 1978), eustatic (Fairbridge and Hillaire-Marcel, 1977), and tectonic (Atwater, 1988; Peterson and Darienzo, 1991) forcing mechanisms in an attempt to rule out certain processes.

The Clayoquot Transgression maximum is younger than the accepted termination age of the post-glacial eustatic sea-level rise. Furthermore, it coincides with a eustatic emergence maximum, so it is unlikely to have a eustatic cause. Co-seismic uplift about 5,100 and 4,000 years ago might have initiated the emergent phase of the late-Holocene sea-level change on Vargas Island. Yet, the Clayoquot Transgression is similar in character to the pattern of Holocene sea-level change documented in outer-coast, ice marginal areas in Norway and Spitsbergen, areas situated on a passive continental margin, and so it is unlikely that end of the Tapes and Talavera Transgressions, as they are respectively referred to, were due to co-seismic uplift. Clark *et al.* (1978) stated that the isostatic response of such ice marginal areas was the most poorly understood, thus it is likely that the Clayoquot Transgression documented for the west coast of Vancouver Island represents an extension of post-glacial eustatic sea-level rise, due to the delayed or prolonged collapse of the marginal forebulge. Thus, both glacio-isostatic and tectonic mechanisms can be considered as geodynamic agents driving the observed submergence/emergence cycle, but both processes are as yet poorly understood and further detailed research needs to be done to resolve the dominant process.

Emergence of the west coast of Vancouver Island during the latter-Holocene is most probably tectonically driven. The critical assessment is whether subduction is occurring a-seismically or seismically. The 4 m strandline on Vargas Island is a prominent emergent feature. Its existence seems to indicate a stillstand or minor submergence event, and its abandonment about 2,200 years ago indicates continued sea-level fall. This timing correlates with the onset of eustatic sea-level fall and with a dated, possibly co-seismic, submergence event in Washington and Oregon. Unfortunately, these two alternative hypotheses can not be discriminated based on the available evidence.

Major difficulties in making paleoseismic comparisons between Vancouver Island, and Washington and Oregon arise from a number of factors: i) the northern, or British Columbia, portion of the Juan de Fuca plate may be segmented, and behaving differently, from that to the south due to the change in the azimuth of the trench in the vicinity of Puget Sound. Thus, strong co-seismic signals evident in Washington and Oregon may not be evident on Vancouver Island due to aseismicity on the northern segment, or the timing of events may be different if the plate segments have ruptured independently; ii) it may be that the southern tip of Vancouver Island, where the most data for the northern segment of the plate has been collected, is situated near the zero isobase for co-seismic deformation with co-seismic movements countered by long-term post-seismic readjustment; iii) due to the unique character of the Cascadia subduction zone (i.e., its youth), the recurrence frequency of thrust earthquakes might be exceedingly large, say greater than 2,000 years, such that the number of events that have occurred since the mid-Holocene may be very few and hard to distinguish in the record; iv) this latter point is confounded by the small amount of emergence that has occurred on the west-coast during the last 4,000 years - three metres of emergence is a narrow band for the resolution of a series of earthquake signals, and post-seismic erosion of the evidence of an event is more than likely along the exposed outer-coast; and finally, v) some or all of the buried marsh horizons in outer-coast estuaries in Washington and Oregon might actually be conditioned by oscillating eustatic sea-levels.

Future Research

Recommendations for future research can be oriented around a few specific goals:

1) Map glaciomarine facies, and marine limit;

i) Pay special attention to the transition from glaciomarine to glaciofluvial materials, and to major slope breaks, to isolate the marine limit. Shell materials are readily available from glaciomarine units for dating the post-glacial marine inundation.

ii) Map the thickness of the glaciomarine sediments and the elevations of the marine limit from outer-coast to inner-coast and develop a glacio-isostatic facies model for Clayoquot Sound (after Boulton *et al.*, 1982). This will help to understand local variations in isostatic response, thus contributing to the study of earth rheology.

2) Constrain the relative sea-level curve more closely. This would entail concentrating on features older than 6,000 yr BP, i.e.;

i) Core Bogs above the marine limit for minimum dates on ice retreat. These bogs also provide a good post-glacial climate record. One such bog is located at the high point of the Ratcliffe/Medallion trail (ca 30 m asl), and is an easy bog to access, with no large feeder creeks, and so will give a good pollen signal for the hypermaritime subzone.

ii) Constrain the early post-glacial emergence by targeting basins between about 20 m asl and 7 m asl. For example, core the bog 500 m inland behind Miltie's Beach for minimum date on abandonment of 9 m asl elevation. Ideally basins should have known bedrock sills. Target the isolation contact in core, using accelerator dating if necessary.

iii) Constrain the early-Holocene lowstand by targeting submarine basins between -3 m asl and about -15 m asl. Look for submerged terrestrial and salt marsh peats, and the isolation contact in core.

iv) Continue with reconnaissance, identification, levelling, and stratigraphic analysis of raised marine landforms, as these classic methods give a feel for the landscape for the best interpretation of the nature of sea-level change. On Vargas Island, the South Bog and Medallion subbasin peat accumulations could be studied in more detail and sampled with proper coring equipment.

3) Continue paleo-seismic research. This would entail studying the late-Holocene emergence in meticulous detail, and would focus on littoral features between 7 m asl and present water;

i) Extend research upfjord with the aim of isolating the hingeline, the zone of no longterm subsidence or uplift. For instance, survey a number of profiles of the mid-Holocene marine scarp around the Hesquiat Peninsula and Boat basin to discover if there is significant landward tilt developed over the last 4 ka. From Estevan Point to the end of

Boat Basin is a 15 km transect, and if extended into Hesquiat Lake, presently tidally influenced during spring tides, an extra 5 km can be added to the transect.

ii) Extend research up the coast to Brooks Peninsula, and beyond, paying special attention to areas around the Nootka fault: Is there a difference in behavior between Explorer and Juan de Fuca plates?

iii) Look for evidence for stepped, or recurrent strandlines, and evidence for strong ground shaking (liquifaction, slumping) in association with certain strandlines.

Appendices

Appendix 1: Survey Data

Table A1.1: Instrument and method used in topographic surveys

Study Site	Inst.	Traverse	Study Site	Inst.	Traverse
Primary sites			*Secondary cont'd*		
Ridge Traverse	Sokisha	closed	Dick & Jane's	NK-1	open
Ahous Lagoon	NK-1	closed	Kelsemaht Beach	Sokisha	closed
Medallion/South Bog	Sokisha	closed	Miltie's Beach	Sokisha	closed
Buckle Berm	NK-1	closed			
Ratcliffe Terrace	Sokisha	closed	*Modern analogues*		
			Ahous North	NK-1	open
Secondary sites			Hopkin's Beach	NK-1	open
Meadow Creek	NK-1	open	Dune Beach	NK-1	open
Medallion Strandflat	NK-1	closed	Moser Point	NK-1	open
McIntosh Cove	Sokisha	closed	Stubbs Island	NK-1	closed

Table A1.2a: Elevations and distance inland of Benchmarks on Ridge Traverse

Station	Type	Elev. (m asl)	Distance (m)	Station	Type	Elev. (m asl)	Distance (m)
CP-2	Stake	2.56	0	BM-11	Stake	6.85	950
BM-23	Stake	3.47	50	BM-10	Stake	7.24	1158
BM-22	Stake	3.54	96	BM-9	Stake	6.48	1214
BM-21	Stake	3.58	136	BM-8	Stake	7.13	1280
BM-20	Stake	3.99	178	BM-7	Stake	7.62	1343
Nail	Root	4.21	297	BM-6	Stake	8.54	1473
BM-19	Stake	5.53	374	BM-4	Stake	6.53	1655
BM-18	Stake	6.52	424	BM-5	Stake	7.67	1590
BM-17	Stake	8.27	530	BM-3	Stake	7.75	1696
BM-16	Stake	4.56	575	BM-2	Stake	7.13	1744
BM-15	Stake	6.79	630	BM-1	Stake	7.47	1820
BM-14	Stake	6.18	690	CP-1	Stake	6.30	1796
BM-13	Stake	7.12	738	BM-A	Stake	7.94	1970
BM-12	Stake	6.94	870	BM-B	Rock	7.50	1998

Table A1.2b: Elevations and distance inland of Benchmarks on Medallion Traverse

Station	Type	Elev. (m asl)	Distance (m)	Station	Type	Elev. (m asl)	Distance (m)
M-1	Nail	3.39	0	M-7	Stake	6.69	308
M-2	Stake	3.20	31	M-8	Stake	6.08	318
M-3	Stake	3.65	90	M-9	Stake	6.15	323
M-4	Stake	4.18	163	M-10	Stake	6.21	328
M-5	Stake	4.53	200	M-11	Stake	6.95	308
M-6	Stake	4.23	230				

Table A1.3: Ahous Lagoon Tide Gauge Observations, June 12 to July 19, 1990

	Elevations (m)				Elevations (m)		
Day	Predicted high tide	Observed high tide	Residual (Pr-Obs)	Day	Predicted high tide	Observed high tide	Residual (Pr-Obs)
June 12	0.80	0.85	-0.05	June 18	0.49	0.41	0.08
June 13	0.83	0.90	-0.07	June 18	1.31	1.17	0.14
June 14	0.86	0.91	-0.05	June 28	1.04	1.16	-0.12
June 15	0.95	0.97	-0.02	July 15	1.19	1.15	0.04
June 16	1.04	1.05	-0.01	July 17	1.31	1.24	0.07
June 17	0.46	0.42	0.04	July 18	1.41	1.30	0.11
June 17	1.16	1.13	0.03	July 19	0.67	0.63	0.04

Mean of the residual equals 0.02 ± 0.07 m, and error associated with observation equals ± 0.03 m

Table A1.4: Difference between predicted tidal height and high tide line

	Elevations (m asl)		
Day	predicted high tide	high tide line	residual (Pr-Obs)
July 17	1.25	1.43	-0.18
July 17	0.40	0.45	-0.05
July 19	1.41	1.45	-0.04
July 19	0.67	0.69	-0.02

Positive bias 0.02 - 0.18 m

Appendix 2: Summary of Radiocarbon Dates

Table A2: New radiocarbon dates from Vargas Island.

Sample ID	C^{14} Age (y B.P.)	Context and Significance
Beta-38875*	5,700 ± 50	Root. 5.3 m asl. Buckle Berm. At contact between glaciomarine mud and peat horizon. Peat buried by marine gravels. Constrains submergence.
Beta-38876	2,280 ± 50	Log. 5.8 m asl. Medallion Scarp. Partially buried in sand, capped by peat. Contemporary with 4 m asl strandline. Constrains emergence.
Beta-41015*	5,500 ± 70	Stick. 5.5 m asl. Buckle Berm. In peat 0.20 m above Beta-38875. Constrains submergence.
Beta-42922	12,970 ± 80	Barnacle shells; 12,170 yr BP corrected** for reservoir effect. 2.5 m asl. Buckle Berm. Shell from stony mud. Unit contained insitu barnacle tests, *Mytilus edulis*, and *Hiatella arctica*. Constrains post-glacial marine inundation.
Beta-43479	2,380 ± 60	Sandy humic peat. 6.1 - 6.15 m asl. Ratcliffe Terrace at pit 8. Peat and pebble lag near top of marine sands. Immediately overlain by 0.30 m of modern forest floor therefore, likely contaminated by modern rhizomes. Minimum age of surface.
Beta-43480*	4,550 ± 80	Fibrous peat. 5.4 - 5.3 m asl. Medallion subbasin at C-11. Fibrous peat between 0.10 m gravel layer below and 0.03 m sand layer above. Contemporary with stillstand.
Beta-43484*	4,740 ± 80	Fibrous peat. 4.7 - 4.6 m asl. South Bog at SB-3. Fibrous peat directly above sand with insitu rhizomes. Contemporary with stillstand.
SFU-857*	5,420 ± 80	Stick. 1.0 m asl. Ahous Lagoon Measure Section. In shell hash at the base of marine sands. On an erosional contact with glaciomarine mud. Constrains age of paleo-wave-cut platform.
SFU-860*	5,900 ± 90	Clam shells; 5,100 yr BP corrected for reservoir effect. 1.0 m asl. Ahous Lagoon Measure Section. Shell from hash at base of marine sands. Constrains age of paleo-wave-cut platform.

SFU-872	2,960 ± 70	Clam shells; 2,160 yr BP corrected for reservoir effect. 4.3 m asl.. Buckle Berm. Shell from midden deposited on paleo-spring-tide gravel berm. Constrains emergence.
Beta-18336*	4,860 ± 90	Wood. 5.8 m asl. Ridge Traverse, 1,320 m inland. On sand at base of peat. Presumed to be logline wood by Rogers *et al.* (1986). Maximum age of surface.
Beta-17516*	4,030 ± 100	Wood. 5.2 m asl. Ridge Traverse, 1,975 m inland. In sand near base at contact with bedrock. Presumed to be logline wood by Rogers *et al.* (1986). Maximum age of surface.
Beta-17517	510 ± 80	Peat. 5.6 m asl. Ridge Traverse, 1,480 m inland. Basal peat on sand. Minimum age of surface.

*: Indicates dates which constrain the mid-Holocene submergence and stillstand.

**: Oceanic reservoir effect; 800 years for the Pacific Northwest Coast (see text).

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