Relative Rise in Sea-level during the Late Holocene at Six Salt Marshes in the Puget Basin, Washington

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RELATIVE RISE IN SEA-LEVEL DURING THE LATE HOLOCENE
AT SIX SALT MARSHES IN THE PUGET BASIN, WASHINGTON

by

Harriet Beale

accepted in partial completion
of the requirements for the degree of
Master of Science

Dean of Graduate School

Advisory Committee

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Harriet Beale
February 18, 2018
ABSTRACT

Relative sea-level changes during the past several thousand years were determined at six marshes in three areas of the Puget Basin, Washington, to identify sites of vertical crustal movement, to provide data on regional relative sea-level history, and to collect data that may be used to monitor a proposed acceleration in the rate of eustatic sea-level rise due to global warming. Six salt marshes were cored to obtain radiocarbon ages of basal fossil peat deposits. Uncertainty in relating fossil marsh plants to elevations of former sea-levels reflects both the vertical range in which the plants occur today at the six study sites, and the responses of salt marshes to the variations in tidal range that accompany rising sea-levels.

Stratigraphic analysis reveals a rise in relative sea-level at all sites. Radiocarbon ages of peat samples indicate a relative sea-level rise of approximately 2-3 ± 0.5 m between 5000 and 3000 years ago and approximately 1 ± 0.5 m between 3000 and 1000 years ago. Relative sea-level in the Puget Basin has probably not risen more than about 1 m in the past 1000 years. These data correlate with those of previous studies from southeastern Vancouver Island and the Fraser Lowland.

No conclusive evidence of either sudden crustal subsidence or differential tilting across the region was
found at the six marshes examined, although slow, steady subsidence of the forearc basin of the Cascadia subduction zone may account for some of the apparent eustatic sea-level rise. Differential crustal displacements of more than 1 m have probably not occurred at Padilla Bay in the past 4500 years, at Quilcene Bay in the past 3000 years, and on San Juan Island in the past 800 years.

Should an acceleration in the rate of sea-level rise from global warming be in progress, erosion observed at marshes on San Juan Island may represent an early signal. Coastal areas with the greatest vulnerability to an accelerated sea-level rise, however, are in southern and eastern Puget Sound, where larger tidal ranges would register the greatest increases in amplitude, so that the highest tides would reach even higher elevations than those otherwise expected for the predicted sea-level rise.
ACKNOWLEDGEMENTS

I would like to acknowledge and thank the people who made this study possible: my advisor, Dr. Maury Schwartz, for his consistent guidance and support; members of my thesis committee Dr. Chris Suczek and Dr. Dave Engebretson, for their interest and their helpful reviews; Doug Canning of the Washington State Department of Ecology for initiating the project and obtaining funding for it; Dr. Brian Atwater of the U. S. Geological Survey in Seattle for providing coring equipment and field instruction, and for being generous with his time and his knowledge; and the Geological Society of America for providing additional funding for field work.

I extend a special thanks to the entire faculty of the Western Washington University Geology Department for their academic and financial support during my graduate career. They have given me an excellent scientific education in a positive learning environment, and have encouraged me personally and shown interest in my progress. I am honored to have learned from them.

Field assistants who volunteered their time were invaluable to the study. Shawn Doan deserves special thanks for helping me with much of the coring during December and January, and for maintaining consistent enthusiasm, goodwill and scientific curiosity. Others who kindly assisted me in
sometimes cold, wet and muddy conditions, included John Costello, Alison Sneed, Jeanmarie Morelli, Mark Pugh, Sue Kahle, Scott Lewis, Crystal Royer, Jerry Wolf, and Sue Lee. I thank each of them for their valuable help.

Employees of the Padilla Bay National Estuarine Research Reserve, the National Park Service at San Juan Island National Historic Park, the University of Washington Friday Harbor Labs, and many state and county agencies provided information and support services. The Washington State Departments of Natural Resources and Ecology allowed the use of their air photos. I also extend my thanks to the property owners who provided access to the salt marshes.

Many others have helped along the way, and I take this opportunity to extend my appreciation to a few of them: Tom Terich, who sparked my interest in Geology; Harvey Kelsey, whose interest and support were most helpful; George Mustoe, who is always a patient teacher; Hugh Shipman of Ecology for stimulating discussions; Bob Bucknam of the U.S.G.S. for including me in a field trip to the southern Sound; and my office partner Kathleen Duggan for her consideration and for providing many helpful suggestions on figures.

I am deeply appreciative of the personal support I have received from patient and understanding friends and family: Jim and Jane Beale, Sarah Phillips, Sue and Ron Kahle, Alice and Jay Shilhanek, Renee Roberts and Carl Root, Cynthia Cornell, Susan Marx, Gordon and Maxine Nicholls, Steve and
Alison Sneed, Debbi Engebretnson, Sue Lee, Bernie Dougan, Kevin Kelley, Roy and Lisa Wright, Jennifer Blomgren, and many others. Several of these good people provided financial and field support in times of need. All of them had unwavering confidence in me and gave their love and friendship freely, and now share in the joy of my success. I extend my heartfelt thanks to them for helping me to reach this goal.
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INTRODUCTION

Relative sea-level changes have occurred in the Puget Basin during the Holocene due to the interaction of tectonic, isostatic and eustatic factors. An understanding of global sea-level history for the Holocene requires the development of local and regional relative sea-level histories that identify and attempt to isolate the effects of site-specific conditions.

The complexity of interrelated factors in the Puget Basin has largely discouraged previous research. However, renewed interest due to the need to determine the regional potential for seismic hazard and to establish baseline data in the event of accelerated sea-level rise induced by global warming has prompted this and other recent studies. The overall goal of this study is to contribute to the data set of late Holocene relative sea-level changes in the Puget Basin.

In this study six Puget Basin salt marshes were examined in order to determine relative sea-level changes at each site during the late Holocene. The study was initiated by the Washington State Department of Ecology Shorelands and Coastal Zone Management Program as part of the Washington State Sea Level Rise Response Project.

Predictions of accelerated sea-level rise due to global climate change are uncertain. An appropriate present-day
response for planning agencies includes inventory and monitoring activities along coastlines and research into the history of sea-level changes in the region. The Sea Level Rise Response Project supported this study primarily to determine if extensively developed areas of southern and inner Puget Sound are more vulnerable to erosion from accelerated sea-level rise than the outer coast, as a result of differential crustal tilting across the region.

Historical changes in relative sea-level suggesting a tilt to the southeast have been proposed by Holdahl and others (1989) from an analysis of precise leveling surveys and tide gauge records. They found that the southern Puget Basin at Tacoma is subsiding at an absolute rate of up to 2.4 mm/yr, while the northern part of the region is relatively stable. Holdahl and others estimate that the northwest coast at Neah Bay is rising at an absolute rate of 2.5 mm/yr and propose a differential crustal tilting that they relate to tectonic activity.

This study of Puget Basin salt marshes has three main purposes. First, the relative sea-level changes determined for the late Holocene may help identify areas of crustal subsidence that are vulnerable to sea-level rise. Evidence of sudden vertical crustal movement would also indicate a region of earthquake hazard and contribute to an understanding of regional structure. In order to meet this objective, marsh deposits were subjected to stratigraphic
A second purpose of the study is to obtain data on regional relative sea-level history for the late Holocene. This was accomplished by stratigraphic analysis and by obtaining radiocarbon dates of peat samples collected from buried salt marsh deposits.

A third purpose of the study is to establish a means of monitoring future sea-level change to predict, for conservation purposes, which marshes are most likely to survive an accelerated sea-level rise. Geomorphic observations and surveys of marsh surface elevations were made to fulfill this purpose.

As the study progressed, regional differences in tidal range became central to an analysis of salt marsh responses to late Holocene relative sea-level changes. Thus, developing an understanding of the role of tidal range as it relates to the three initial research purposes became a fourth and integral goal of the study.
REGIONAL RELATIVE SEA-LEVEL HISTORY

Introduction

Regional relative sea-level changes in the Puget Basin are dependent on the interaction of tectonic, isostatic and eustatic factors. Each of these factors has resulted in relative sea-level changes that have varied in direction, rate and magnitude at different times in recent geologic history. Tectonically-induced vertical crustal movement results from the large-scale motions of crustal plates. Isostatic movement occurs as the crust and underlying mantle adjust vertically to changes in crustal mass. Eustatic changes occur due to changes in the level of the water surface; the changes need not be global, but may be limited to a regional extent. The sum of tectonic, isostatic and eustatic effects determines the relative change in sea-level at any particular time and geographic location.

This section reviews some of the known tectonic, isostatic and eustatic causes of relative sea-level changes in the Puget Basin for the Holocene. Discussion focusses on the extent to which these factors can be isolated in order to better understand their influence on the relative sea-
level changes found in this study.

Tectonic relative sea-level changes

The Puget Basin and adjacent waters are an estuary of the Pacific Ocean lying in a north-south trending structural trough that has been modified by Pleistocene glacial erosion and deposition (Figure 1). The geologic history of the region is dominated by the late Mesozoic and early Tertiary accretion of terranes of varying lithologies. Orogenic processes since the late Cretaceous include contraction, metamorphism, intrusion, volcanism and uplift to form the Cascade Mountains. Puget Sound lies in a fore-arc basin between the Cascade magmatic arc and the accretionary prism that forms the Olympic Mountains (Tabor and others, 1989). Slow, steady crustal subsidence is common in fore-arc basins, and may contribute to a regional relative sea-level rise.

The Puget Basin is located in the Cascadia subduction zone (Figures 2a and 2b). Evidence reviewed in this section indicates that complex compressional and extensional structures underlying the Puget Basin are related to the limited extent and young age of the downgoing Juan de Fuca plate and to the arcuate shape of the plate margin. Tectonically-induced relative sea-level changes may reflect
Figure 1. Map of the Puget Basin and surrounding region showing locations of interest to this study.
Figure 2a. Map of Cascadia subduction zone.  
Figure 2b. Cross-section of Cascadia subduction zone (after Gower and others, 1985).
this structural complexity and thus may vary across the region.

Tectonic activity in the Cascadia subduction zone is segmented, reflecting the independent behavior of the Juan de Fuca plate and the Explorer subplate and Gorda block (Figure 2a) with respect to each other. In a study of magnetic anomalies of the Juan de Fuca plate system, Riddihough (1984) determined that at about 4 Ma the North American-Juan de Fuca relative plate velocity slowed and the Juan de Fuca plate began to break up. At that time, the Explorer plate began to move independently; at about 2.5 Ma the Gorda South Block also began to move independently with respect to the Juan de Fuca plate.

The young and buoyant Explorer subplate is not presently subducting with its own driving forces because it lacks the weight and density of an older plate (Riddihough, 1984). Instead, it is being overridden by the North American plate. The Explorer subplate descends beneath Vancouver Island at a shallow angle because of its low density; Riddihough (1984) suggests that because of this buoyancy, it is underplating and uplifting northern Vancouver Island.

The Juan de Fuca plate is older and more dense than the Explorer subplate and is actively subducting, although at a slow rate. Cross-sections constructed by Riddihough (1984) based on seismic and gravity data show that the Juan de Fuca
plate is subducting at a steeper angle than the younger Explorer subplate, and the former increases its angle of subduction in the region under the north-south trending Puget Basin. Weaver and Baker (1988) found this change in the angle of plunge to be from 11 degrees to the west of the Puget Basin to about 25 degrees eastward.

Spence (1989) noted that the largest recent earthquakes, those of 1949, 1965, and 1976, all occurred in the area of increased angle of plunge, or "slab bend," at depths of 54 km, 59 km, and 60 km, respectively. This area is shown as the "lower seismic zone" in Figure 2b. First seismic motion for these deep events was extensional in the down-slab direction. For earthquakes with shallower focal depths, first-motion axes have been dominantly north-south and are compressional. The structural trough that underlies the Puget Basin may be the result of lithosphere that has been downdropped due to the extensional features at the slab bend.

In the 150 years since non-Indian settlement, the Cascadia subduction zone has not experienced the large-magnitude shallow-thrust earthquakes typical of similar subduction zones such as those near Chile and Japan. The hypothesis of aseismic subduction occurring by slow creep due to a relatively young and warm Juan de Fuca plate (Ando and Balazs, 1979) is countered by Heaton and Hartzell's (1987) suggestion that Cascadia subduction is coseismic, but
Evidence found by Atwater (1987) indicating sudden prehistoric subsidence events in coastal marshes supports the hypothesis of coseismic subduction of the Juan de Fuca plate. These events, suggesting earthquakes of magnitude 8 or greater, have been dated at 300, 1600, 1700, 2700, and 3100 years BP. Vertical crustal movement in the Puget Basin accompanying magnitude 8 coastal earthquakes may overprint the record of structural deformation directly under the Puget Basin.

The margin of the Cascadia subduction zone has an arcuate shape that is most curved along the northwest Washington Coast. Rogers (1983) suggested that, where the dip of the plate steepens, the plate deforms in bulges to accommodate the arcuate plate margin, much like a tablecloth draped over the corner of a table. This model allows for the formation of complex extensional and compressional structures in the region above the slab bend.

Seismic and gravity studies provide evidence suggesting that such structures exist. Wagner and Wiley (1983) mapped faulting in sediments less than 3.4 million years old in seismic reflection studies under central and northern Puget Sound. Gower and others (1985) inferred that large-scale faults cross the sound from east to west and northwest to southeast. In a seismic reflection study, Harding and
others (1988) found evidence for possible downdropped blocks that trend east-west, displacing glacial deposits between Seattle and Bainbridge Island. Seismic data correlate with gravity data from Bonini and others (1974) suggesting a deep basinal structure in this area.

The compressional and extensional structures suggested by seismic and gravity data are obscured in part by Quaternary glacial deposits that blanket most of the region. Puget Sound and waters adjacent to it also cover much of the regional structure.

Evidence of less than 0.5 m of sudden late Holocene crustal uplift has been found at Lynch Cove in southern Hood Canal (Figure 1), and 7 m of uplift has been estimated for a raised marine terrace on Bainbridge Island which has been constrained to a maximum age of 1700 years by radiocarbon dating (Bucknam and Barnhard, 1989).

Gower (1978) mapped both normal and thrust faults affecting Pleistocene and post-Pleistocene deposits near Port Townsend. Gower and others (1985) cite evidence of 3.5 m of vertical displacement along a left-lateral oblique-slip fault at Saddle Mountain, west of southern Hood Canal. Radiocarbon ages of associated drowned tree stumps have dated the event as 1100-1300 years old.

Radiocarbon ages from a peat bog at Shine (Figure 1) suggest a continuous rise in relative sea-level at that northern Hood Canal site for about 4000 years (Eronen and
The structural history of the Puget Basin includes compressional and extensional deformation during the Holocene. The crust may in some places be uplifted and in others downdropped relative to former sea-levels, so that relative sea-level changes determined for this period at different sites in the Puget Basin may vary both in direction and in magnitude.

Diatom assemblages from the same core indicate alternating freshwater and marine/brackish conditions, but the continuous nature of upward peat growth suggests that sudden tectonic movement was not involved.

The structural history of the Puget Basin includes compressional and extensional deformation during the Holocene. The crust may in some places be uplifted and in others downdropped relative to former sea-levels, so that relative sea-level changes determined for this period at different sites in the Puget Basin may vary both in direction and in magnitude.

Isostatic relative sea-level changes

The Puget Basin and the trough under the Strait of Juan de Fuca were pre-existing lowlands into which the Cordilleran ice sheet advanced during the Pleistocene epoch (1.8 Ma to 10,000 BP). Extreme changes in relative sea-level resulted from crustal responses to recurrent periods of ice loading and unloading and from eustatic sea-level variations due to ice sheet growth and decay.

Evidence of early Pleistocene glaciation is largely obscured by the most recent glacial advance, the Fraser glaciation. The Orting, Stuck and Salmon Springs glaciations are all dated at earlier than 800,000 BP (Blunt
and others, 1987). The next known glaciations, the Double Bluff (dated at 100,000 to 250,000 BP) and the Possession Drift (dated at 35,000 to 90,000 BP); left glaciomarine deposits at elevations that indicate relative sea-levels higher than at present (Blunt and others, 1987). Interbedded with glacial sediments are terrestrial interglacial deposits indicating relative sea-levels lowered by isostatic rebound.

The Fraser glaciation began in the Puget Lowland about 18,000 BP with the advance of the Vashon ice (Blunt and others, 1987). Although eustatic sea-level is estimated to have been approximately 100-150 m below current sea-level about 15,000 BP, crustal depression by ice sheets resulted in regional relative sea-levels higher than at present (Komar, 1976).

Vashon glacial recession and the accompanying crustal rebound began about 14,500 BP; deposits of Everson Glaciomarine Drift record the period of floating ice from about 11,000 to 13,000 BP at elevations 200 m above present sea-level near Bellingham (Figure 1). The overlying Sumas outwash deposits indicate emergence of the northern lowland by about 10,000 BP (Easterbrook, 1979).

Post-glacial isostatic adjustment was complete for the most part by 8000 BP (Mathews and others, 1970). Thorson (1989) related what is a relatively rapid rate of post-glacial uplift in the Puget Basin to the low apparent
strength and viscosity of the mantle in a subduction zone. Thorson suggested that residual isostatic rebound due to adjustment of the underlying mantle to changes in crustal mass is not as closely related to ongoing seismicity as is the release of elastic strain stored along faults. These faults either pre-dated Quaternary glaciation or formed along pre-existing zones of structural weakness during repeated glaciations due to the weight of ice sheets. Thorson acknowledged the difficulty of isolating eustatic, isostatic and tectonic factors of relative sea-level change in a subduction zone setting.

Eustatic relative sea-level changes

The eustatic component of relative sea-level change cannot be isolated by applying global rates of sea-level rise. Recent geophysical research has shown that eustatic sea-level does not change uniformly around the globe in response to large-scale shifts of water between continents and ocean basins during glaciation and deglaciation (Morner, 1979; Fletcher, 1988). Clark and Lingle (1979) found that variable effects due to the viscoelasticity of the Earth’s crust and mantle result in changes in the geoid that can be on the same order of magnitude as the overall eustatic change from glacial melting. One of these effects is a pro-
Fletcher (1988) discussed the multiple geophysical, tectonic and climatological reasons for regional and site-specific variations in the rates and directions of eustatic sea-level change. Vertical crustal movement varies with individual plate motions and large-scale plate interactions; variable spreading rates at mid-ocean ridges result in changes in the capacity of the ocean basins. Other factors include sediment-loading and hydro-isostasy (Pirazolli, 1977), and gravitational changes that may relate to the redistribution of crustal mass because of changes in the earth’s rotation (Morner, 1979). Thermal effects on ocean volume and regional and site-specific climatological effects of winds, currents, river discharge and atmospheric pressure may cause eustatic variations, as well (Komar, 1976).

Eustatic sea-level changes, previously defined as simultaneous and global, are no longer well-defined. Morner (1979, p. 537) redefined eustasy as "...simply 'ocean level changes' regardless of causation and implying vertical--global and local--movements of the ocean surface at a particular point."

Fletcher (1988) recommended that research focus on the development of regional relative sea-level curves for which local variables have been identified. Fletcher and others
The tectonic complexity and glacial history of the Puget Basin require that data for many sites within the area be determined before a regional relative sea-level history is defined.

At the completion of essentially all post-glacial rebound in the Puget Basin about 8000 BP (Mathews and others, 1970), shorelines were at lower levels than at present. In the Fraser Lowland, five radiocarbon dates of an organic peat on an alluvial surface indicate that sea-level in the Strait of Georgia was approximately 12 m lower than at present about 7300 ± 120 and 8360 ± 170 BP (Clague and others, 1982). Shipman (1989) cited archaeological and geological studies that are consistent with evidence from Clague (1989) suggesting a rapid relative sea-level rise in the northern lowland until about 5000 BP. Rates of sea-level rise slowed after that, and Clague and Bobrowsky (1990) estimated that for eastern Vancouver Island and the British Columbia mainland shorelines have remained within a few meters of present sea-level for the last 3000 years, remaining within ±0.5 m of present sea-level for the past 2000 years.

Current rates of sea-level rise for the conterminous United States, based on several decades of historical tide gauge data, are estimated at 1.3 ± 0.2 mm/yr (Hicks and
others, 1983). Average global sea-level rise based on tide gauge data is estimated at 10-15 cm for the past 100 years (Titus, 1988).

Rising post-glacial sea-levels submerged the lower parts of river valleys and caused rapid rates of erosion on coastlines of Puget Sound and adjacent waters that are composed largely of unconsolidated glacial material. Most of the depositional coastal features observed today were formed by littoral processes during the past 5000 years, when rates of sea-level rise slowed enough to allow their survival (Komar, 1976). Barriers composed of re-deposited glacial sediments formed lagoons and other protected environments where salt marshes grew. Marshes also formed on prograding river deltas where high sedimentation rates led to the development of tidal flats that dissipated wave energy and allowed salt marsh vegetation to become established.
Introduction

Salt marshes are depositional landforms that form under favorable tidal and sedimentary conditions. This section reviews the tidal and sedimentary factors that contribute to salt marsh development and the vegetation species common to tidal marshes in the Puget Basin. Following this, the causes of sedimentary facies changes in salt marsh stratigraphy are discussed in the context of the Puget Sound coastal environment.

Tidal Range

The tide is a wave with a dominant frequency of 12 hr 25 min. The amplitude of this wave at any one site is the tidal range (Pethick, 1984). A linear relationship has been demonstrated between tidal range and the distance the tide travels across the continental shelf (Redfield, 1958; Cram, 1979). Tidal ranges in the Puget Basin reflect this relationship, increasing with distance from the open sea (Table 1 and Figure 3). Amplification of the tidal range in
Table 1: Puget Basin tidal ranges from Mean Lower Low Water to Mean Higher High Water are listed by average for each region, in order of increasing distance from the Pacific Ocean. The Washington Sound and Strait of Georgia region includes U.S. stations from Padilla Bay north and in the San Juan Islands (data from Canning, 1990).

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<th>Region</th>
<th>No. of stations</th>
<th>MLLW-MHHW (m)</th>
<th>MLLW-MHHW (ft)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Strait of Juan de Fuca</td>
<td>8</td>
<td>2.30</td>
<td>7.54</td>
</tr>
<tr>
<td>2. Washington Sound and Strait of Georgia</td>
<td>9</td>
<td>2.45</td>
<td>8.04</td>
</tr>
<tr>
<td>3. Admiralty Inlet</td>
<td>2</td>
<td>2.59</td>
<td>8.50</td>
</tr>
<tr>
<td>4. Possession Sound and Port Susan</td>
<td>3</td>
<td>3.35</td>
<td>10.99</td>
</tr>
<tr>
<td>5. Saratoga Passage and Skagit Bay</td>
<td>8</td>
<td>3.35</td>
<td>10.98</td>
</tr>
<tr>
<td>6. Hood Canal</td>
<td>4</td>
<td>3.35</td>
<td>11.00</td>
</tr>
<tr>
<td>7. North Puget Sound</td>
<td>10</td>
<td>3.47</td>
<td>11.40</td>
</tr>
<tr>
<td>8. South Puget Sound</td>
<td>8</td>
<td>4.26</td>
<td>13.96</td>
</tr>
</tbody>
</table>

Figure 3. Map of Puget Basin showing the locations of tidal range regions as numbered in Table 1.
continental shelf areas is explained in part by the inversely proportional relationship of tidal amplitude to both water depth and channel width:

\[ A^4 \propto 1/h \quad \text{and} \quad A^2 \propto 1/b \]

where \( A \) = wave amplitude
\( h \) = water depth
\( b \) = channel width

(Stride, 1982)

Tidal ranges thus increase in shallow water and where channels are constricted in width. The tidal range may also be modified by tidal resonance across the continental shelf or by local resonance within a bay. Tidal resonance commonly occurs in a narrow bay where the length and depth of the bay are such that constructive interference of the reflecting waves reinforces the subsequent entering wave of the tide. Resonance can increase the tidal range at the head of a bay over that at the mouth of the bay by four or five times (Pethick, 1984).

Boggs (1987) divided the tidal range environment according to the frequency of tidal inundation; the subtidal zone lies below Mean Low Water (MLW), the intertidal zone lies between MLW and Mean High Water (MHW), and the supratidal zone lies above the MHW level. The subtidal zone is only rarely exposed, the intertidal zone is exposed once or twice daily, and the supratidal zone is flooded only by
Salt marshes generally develop in mesotidal (2-4m) and macrotidal (>4m) coastal areas in which tidal landforms dominate over wind-wave landforms (Pethick, 1984). A larger tidal range allows a greater vertical range for salt marsh development and a more pronounced zonation of salt marsh vegetation. As tidal range increases, salt marsh plant species migrate to higher elevations and extend this vertical range (Ranwell, 1972).

Salt Marsh Sedimentation

**Sediment sources and accumulation**

Tides provide most of the energy that transports sediment to tidal flat and salt marsh depositional settings. Sediment may be of coastal origin, transported from eroding bluffs by longshore drift, or from fluvial sources, which include deltaic deposits and fluvially-derived bay muds remobilized during storm events (Frey and Basan, 1985).

The horizontal velocity of the tidal current is zero at low tide, reaches a maximum at half-tide and decreases again to zero at high tide. Thus, as the water moves in across the tidal flat, sediment is entrained by increasing velocity until the zone is reached at which half-tide occurs, and beyond this zone the velocity decreases toward the shore. As

21
the velocity decreases, sediment settles out; the zone of deposition where salt marshes develop is landward of this boundary (Pethick, 1984). Although this boundary between erosion and deposition is not always distinct, tidal flat elevations increase beyond it in an onshore direction.

There is also a net movement of individual sediment particles landward from the seaward point where they enter the system (Frey and Basan, 1985). A particle is entrained and transported landward until the velocity decreases below the particle's settling velocity. Because the grain is carried forward by the current in the thin layer of viscous flow near the bed as it falls, a settling lag occurs. The grain is therefore deposited further shoreward than the point where its settling velocity is reached. As the tide turns, the ebb-tidal flow will require more energy to re-entrain the grain and suspension will occur later in the tidal cycle. As a result, the particle will be suspended on the ebb-tide for a shorter time, and will gain distance in an onshore direction. The increased rates of deposition on the upper tidal flat result in vertical accretion to elevations at which salt marsh vegetation can be established (Pethick, 1984).

Maximum deposition of suspended sediment occurs at the turn of the tide when the current velocity is zero. Deposition of suspended sediment is enhanced by electrolytic flocculation of clay-sized particles where fluvial
freshwater enters salt water. Clay flocculation generally accounts for less sediment deposition than does the formation of aggregates of clay and organic compounds, which are larger than clay floccules and have higher settling velocities (Frey and Basan, 1985). River flooding contributes further sediment to deltaic depositional environments. Macroinvertebrate fecal material and pelletization of clay and silt by suspension feeders are also significant depositional processes.

Sediment accumulates on vegetated surfaces due to the damping effects of plants and organic debris, as well as to increased clay flocculation near salt-emitting plants. Substrate cohesion is increased on the vegetated marsh by roots and by algal, bacterial and diatom films (Frey and Basan, 1985).

Grain size distribution

Grain sizes on unvegetated tidal flats generally decrease from the lower to the upper tidal flat. The upper tidal flat grades from sand-sized particles to increasingly finer silt and clay particles in an onshore direction from the boundary between zones of entrainment/erosion and deposition (Frey and Basan, 1985). This is the result of decreasing velocities from the middle to the nearshore part of the tidal flat, as discussed in the previous subsection. Salt marshes are prograding landforms, and thus their
depositional sequences have a fining-upward pattern from lower intertidal sands to the mud, silt and peat of the supratidal zone (Harrison, 1975).

Sediment deposition on the salt marsh occurs when tidal channels are overtopped, flooding the marsh. Grain sizes on the vegetated marsh generally decrease with distance from tidal channels (Pestrong, 1972), and grains are not as well-sorted as those on tidal flats (Bouma, 1963).

Peat is "an accumulation of partially decomposed and disintegrated plant remains which have been fossilized under conditions of incomplete aeration and high water content" (Bell, 1981, p.60). Salt marsh peat is a fibrous peat with an extremely high void ratio; it is subject to compaction under its own weight of up to 44% of its original thickness in 6000 years (Bloom, 1964). Kaye and Barghoorn (1964) estimate the porosity of peat to be from 75% to 90% depending on the silt and clay content. Incorporated into peat deposits are drift logs and other debris that have accumulated on former marsh surfaces.

Rates of vertical accretion

Rates of vertical accretion on salt marsh surfaces depend on the net effects of sediment deposition, erosion and compaction. Rates of sediment deposition vary seasonally; they are higher in the summer when the erosive effects of storms decrease. Maximum rates of vertical
accretion occur on young marshes after lateral limits to marsh growth on the margins of the estuary have been reached (Frey and Basan, 1985). Vertical accretion rates are asymptotic over time, slowing as the upper marsh approaches the elevation of the highest tides (Pethick, 1981).

Vertical accretion rates in young marshes (less than 100 years old) have been determined to be as high as 10 cm/yr and depend primarily on rates of sedimentation, while older marshes (more than 200 years old) stabilize at rates of accretion of less than 0.001 cm/yr and are dominated by vertical plant growth (Pethick, 1984). As the marsh ages, its slope flattens between its landward edge and the zone of maximum accretion, and subsequent sedimentation under conditions of stable sea-level results in seaward growth (Ranwell, 1972).

During conditions of rising relative sea-level, rates of vertical accretion on mature marshes are determined by the ability of upward plant growth to keep pace. Rates of relative sea-level rise on salt marshes are a function of eustatic sea-level rise, crustal subsidence, compaction, and tidal range. Harrison and Bloom (1977) found that rates of vertical accretion in a 10-year period on mature marshes in Connecticut correlated with the amplitude of tidal ranges; higher rates of vertical growth were associated with the greater tidal ranges.
Sedimentary structures

Sedimentary structures typical of salt marsh deposits include discontinuous parallel and wavy laminations, lenticular bedding, fine laminae with distinct contacts, and frequent disruption of laminae by root structures and bioturbation (Elliot, 1986). Frey and Basan (1985) noted that dessication cracks and salt crystal molds are commonly found on the low marsh and in salt pans. In some marsh deposits, seasonal variations between horizons of storm-derived sediment and summer organic growth have been recognized (Redfield, 1972).

Structures found in low marsh deposits more closely resemble those of tidal flats than do those of the high marsh. These may include flaser bedding and small-scale ripple cross-stratification. In thinly-vegetated areas of the low marsh, indicators of subaerial exposure such as raindrop imprints and foam marks may be preserved (Boggs, 1987).

Salt marsh drainage

Drainage systems on vegetated surfaces are more complex and meandering than those on unvegetated tidal flats (Pestrong, 1972). Drainage patterns are typically dendritic (Boggs, 1987) and tend to retain their position as the marsh grows upward unless depositional conditions change.
Natural levees are common, and eroded slump blocks often heal, so that drainage systems on salt marshes tend to be stable over extended intervals.

Salt marsh drainage systems as described by Frey and Basan (1985) develop extensively during stages of youth, stabilize in position during mature stages, and become partially or totally filled on older marsh surfaces. On older marshes, salt pans may form at former drainage headwaters, and tidal creek drainage is gradually superseded by surface runoff processes.

Salt marsh vegetation

Zonation of vegetation

Salt marsh plant species are zoned according to elevation with respect to sea-level (Figures 4 and 5) (Hinde, 1954; Chapman, 1960; Ranwell, 1972). Salt marshes originate on upper parts of tideflats when sediments accumulate vertically to elevations at which the surface is exposed for enough time every 24 hours that pioneer plant species become established. Vertical accretion promotes the growth and survival of additional species. Species selection is controlled by multiple variables including elevation, salinity, eH and pH, insolation, turbidity, wave energy, topography, tidal range and sediment supply (Frey
Figure 4. Diagram showing the zonation of salt marsh surfaces in relationship to sea-level. The diagram is drawn to U.S. Atlantic Coast tidal datums (from Titus, 1988).

Figure 5. Photo of zoned vegetation at Third Lagoon, San Juan Island.
The most important of these interrelated variables are elevation and salinity. Halophytic (salt-tolerant) species are zoned from the bare tidal flat to the upland according to the number and duration of submergences they undergo during the tidal cycle. In most salt marshes the zonation is controlled by elevation, although disturbed drainage patterns may modify the zones.

Macdonald (1977) described this zonation for the mixed tidal regimes of the United States Pacific coast, dividing the environment into the low marsh, extending from approximately Mean Lower High Water (MLHW) to Mean Higher High Water (MHHW), and the high marsh, from approximately MHHW to Extreme High Water (EHW). The low marsh is frequently submerged for more than six hours at a time and is never continuously exposed for more than fifteen days. The high marsh is submerged for less than six hours daily for several days and then is continuously exposed for several weeks to several months. Pioneer species of the low marsh propagate by the expansion of rhizomes, root-like underground stems that stabilize the sediment and survive periods of tidal submergence (Northwest Environmental Consultants, 1975).
Table 2. Plant species observed in Puget Basin salt marshes (Raedeke and others, 1976; Disraeli and Fonda, 1979; Burg and others, 1980; Seliskar and Gallagher, 1983; Kunze, 1984; Granger and Burg, 1986).

<table>
<thead>
<tr>
<th>Scientific name</th>
<th>Common name</th>
<th>Marsh position</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Agrostis alba</em></td>
<td>creeping bentgrass</td>
<td>high</td>
</tr>
<tr>
<td><em>Atriplex patula</em></td>
<td>orache</td>
<td>high</td>
</tr>
<tr>
<td><em>Carex lyngbyei</em></td>
<td>Lyngbye’s sedge</td>
<td>low</td>
</tr>
<tr>
<td><em>Cuscuta salina</em></td>
<td>dodder</td>
<td>low</td>
</tr>
<tr>
<td><em>Deschampsia caespitosa</em></td>
<td>tufted hairgrass</td>
<td>high</td>
</tr>
<tr>
<td><em>Distichlis spicata</em></td>
<td>saltgrass</td>
<td>low</td>
</tr>
<tr>
<td><em>Eleocharis palustris</em></td>
<td>spike-rush</td>
<td>low</td>
</tr>
<tr>
<td><em>Eleocharis parvula</em></td>
<td>spike-rush</td>
<td>low</td>
</tr>
<tr>
<td><em>Elymus mollis</em></td>
<td>limegrass</td>
<td>high</td>
</tr>
<tr>
<td><em>Glaux maritima</em></td>
<td>milkwort</td>
<td>low</td>
</tr>
<tr>
<td><em>Grindelia integrifolia</em></td>
<td>gum plant</td>
<td>high</td>
</tr>
<tr>
<td><em>Hordium brachyantherum</em></td>
<td>meadow barley</td>
<td>high</td>
</tr>
<tr>
<td><em>Jaumea carnosa</em></td>
<td>jaumea</td>
<td>low</td>
</tr>
<tr>
<td><em>Juncus balticus</em></td>
<td>baltic rush</td>
<td>high</td>
</tr>
<tr>
<td><em>Juncus lesueurii</em></td>
<td>salt rush</td>
<td>high</td>
</tr>
<tr>
<td><em>Potentilla pacifica</em></td>
<td>pacific silverweed</td>
<td>high</td>
</tr>
<tr>
<td><em>Plantago maritima</em></td>
<td>seaside plantain</td>
<td>low</td>
</tr>
<tr>
<td><em>Puccinellia pumila</em></td>
<td>alkaligrass</td>
<td>low</td>
</tr>
<tr>
<td><em>Rumex occidentalis</em></td>
<td>western dock</td>
<td>high</td>
</tr>
<tr>
<td><em>Salicornia virginica</em></td>
<td>pickleweed</td>
<td>low</td>
</tr>
<tr>
<td><em>Spergularia canadensis</em></td>
<td>sandspurry</td>
<td>low/high</td>
</tr>
<tr>
<td><em>Spergularia macrotheca</em></td>
<td>sandspurry</td>
<td>low</td>
</tr>
<tr>
<td><em>Spergularia marina</em></td>
<td>sandspurry</td>
<td>low</td>
</tr>
<tr>
<td><em>Scirpus acutus</em></td>
<td>hardstem bulrush</td>
<td>high</td>
</tr>
<tr>
<td><em>Scirpus americanus</em></td>
<td>three-square</td>
<td>low</td>
</tr>
<tr>
<td><em>Scirpus validus</em></td>
<td>bulrush</td>
<td>low</td>
</tr>
<tr>
<td><em>Trifolium wormskjoldii</em></td>
<td>marsh clover</td>
<td>high</td>
</tr>
<tr>
<td><em>Triglochin maritimum</em></td>
<td>arrowgrass</td>
<td>low</td>
</tr>
</tbody>
</table>
Puget Basin salt marsh species

Plant species commonly found in Puget Basin salt marshes are listed in Table 2. Many of these species grow on both the low and the high marsh but are listed with the marsh position at which they generally occur.

Species that pioneer the mudflats include *Triglochin maritimum*, *Salicornia virginica*, and *Spergularia canadensis* (Mason, 1957; Northwest Environmental Consultants, 1975; Seliskar and Gallagher, 1983; Kunze, 1984). Kunze (1984) observed that *Glaux maritima* is frequently a pioneer on sandy-gravel deltas of Hood Canal.

Salinity, temperature and other controls on species survival vary from site to site, so that species that dominate on the low marsh are frequently found as minor species in high marsh assemblages. Examples include *D. spicata* and *T. Maritimum*, both of which are common species from low marsh to high marsh elevations, as are *Atriplex patula*, *Jaumea carnosa* and *Salicornia virginica*. *Deschampsia caespitosa*, a high marsh species, is observed growing on many low marsh surfaces (Kunze, 1984).

High marsh diversity increases above the elevation of MHW due to decreasing soil salinity, while low marsh assemblages are generally limited to fewer species (Macdonald, 1977). Chapman (1960) noted that diversity in Pacific Coast salt marshes increases toward the south, implying that temperature is a primary factor controlling
diversity.

Barbour and others (1973) grew \textit{D. spicata}, \textit{S. virginica} and \textit{J. carnosa} at different salinities in greenhouse environments and found that the plants grew as well, if not better, at lower salinities and in dryer soil than they did in the typical low marsh conditions. The authors concluded that these species are capable of survival in many freshwater environments but are unable to compete with other species. They apparently flourish in salt marsh environments because they are salt-tolerant and have no competitors.

Sedimentary Facies Changes

Introduction

The underlying stratigraphy of salt marsh deposits often reveals variations in sedimentary facies. Depositional changes in Puget Basin salt marshes may occur as a result of:

1) a site-specific relative sea-level rise,
2) local variations in tidal range,
3) gradual or sudden tectonic movement, or
4) steady or uneven rates of eustatic sea-level change.
Site-specific relative sea-level rise

Compaction of sediment or ponding of water on the marsh surface may cause a site-specific relative sea-level rise that is neither tectonic nor eustatic in origin. A compaction or ponding event may include the following:

1) autocompaction of peat under its own weight;
2) compaction caused by seismic shaking;
3) compaction caused by the weight of drift logs covering the marsh;
4) compaction caused by the weight of a forest after a change from salt marsh to freshwater conditions;
5) ponding when tidal channels are blocked by drift logs or other debris;
6) ponding during flooding, extreme high tides, or storm surges.

If compaction lowers salt marsh elevations to tidal flat conditions, a layer of clay is deposited over the peat. Ponding of water allows fine-grained material to settle out of suspension, forming a thin silt or clay layer. Such clay layers may resemble those caused by tectonic or eustatic sea-level changes.

Variations in tidal range

A change in tidal range may result in a lateral shift in depositional environment within the intertidal zone. Tidal ranges vary locally with the shape of the basin and
sediment supply. These factors may change as a result of ongoing erosion under conditions of stable sea-level, and are likely to change more rapidly under conditions of rising relative sea-level.

Tidal ranges also increase at the high tides of the 18.6 yr lunar nodal cycle. Kaye and Stuckey (1973) correlated this cycle with minimum and maximum tidal ranges at Boston, New York, and Charleston for the period 1920-1970. Pethick (1984) attributed salt marsh erosion in Britain to this phenomenon in the years just before and after the peak tides of 1982.

Changes in tidal range are followed by migration of marsh vegetation zones and changes in the elevations of marsh surfaces. A breached berm may flood a lagoon so that low marsh or tidal flat conditions supersede a high marsh environment. An increased sediment supply may result in basin shallowing, increasing the tidal range so that salt marsh vegetation moves up in the intertidal zone and tidal flat clay is deposited over marsh vegetation. In an application of Walther's Law of vertical succession of facies, stratigraphic order may reflect the changing sediment and salinity patterns caused by altered tidal ranges, masking the actual change in relative sea-level.

**Tectonic subsidence**

Seismically-induced tectonic subsidence results in
sudden marsh submergence. Atwater (1987) has located buried conifer forests in protected estuarine salt marshes on the Washington coast that he interprets as evidence of sudden late-Holocene subsidence events related to coseismic subduction of the Juan de Fuca plate. Other workers have found buried high marsh surfaces on the north and central Oregon coast that have yielded radiocarbon dates corresponding with those of subsidence events described by Atwater in southwestern Washington (Grant and MacLaren, 1987; Darienzo and Peterson, 1987; Nelson, 1987).

Evidence for multiple sudden subsidence events in these studies includes repeated sequences of tidal flat deposits grading upward to high marsh peat deposits that are overlain by a sediment-capping layer that workers interpret as being of storm or tsunami origin. Distinct parting contacts are observed between high marsh peat deposits and overlying tidal clay. Microfossil assemblages have been shown to correlate with stratigraphic interpretations at several of the Oregon sites.

A vertical displacement that removes the marsh from the range of halophyte growth results in a change up to freshwater or down to tidal flat conditions. A small vertical displacement that does not result in a change in depositional environment or a relative sea-level rise from the slow, steady subsidence of the fore-arc basin are difficult to recognize in the stratigraphic record. In this
study, the maximum vertical range observed for the occurrence of an indicator plant species is 1.49 m. Vertical displacements of less than that amount may not be noticeable in a stratigraphic analysis of marsh deposits.

**Eustatic sea-level rise**

A late-Holocene regional relative sea-level rise has been documented for the southern British Columbia coast by Clague and Bobrowsky (1990), but variations in the rate of rise have not been determined. In the absence of evidence for coseismic vertical crustal movement, those data may represent a regional eustatic sea-level rise. This regional eustatic sea-level rise may include a component of relative sea-level rise due to regional subsidence of the fore-arc basin.

The depositional record of eustatic sea-level rise may vary. If eustatic sea-level rises slowly, the marsh keeps pace by upward growth, and no clay layer forms. If the rise occurs in pulses, thin layers of clay may be deposited. These layers represent a period of tidal flat conditions that persist until sediment re-accumulates to higher elevations.

A rapid eustatic sea-level rise may result in periods of erosion that may not be obvious in a cross-section of marsh deposits. The profile of an eroding marsh is an
irregular surface of small vegetated plateaus and islands (Redfield, 1972; Ranwell, 1972). If a marsh erosion surface aggrades slowly, lenses of peat, silty peat and clay form that resemble the lenses formed during a marsh transition from a sparsely vegetated tidal flat to an established marsh.

Conclusion

Puget Basin salt marsh stratigraphy may be subject to multiple interpretations of depositional history. The complex coastline of the inland waters is affected by variable wind and wave patterns and has an abundant sediment supply from glacial deposits. The Puget Basin shoreline experiences frequent changes in depositional environments as shores erode and sediment is shifted. Changes in the sedimentary facies of marsh deposits may also reflect compaction or ponding events, most commonly as a result of drift logs from the heavily forested upland that cover the marsh or block tidal channels.

Alternate hypotheses must be ruled out before depositional changes observed in salt marsh deposits are attributed to tectonic or eustatic factors. Only if the changes are observed over a large areal extent or are accompanied by strong supportive evidence can conclusions be drawn regarding tectonically or eustatically-induced relative sea-level changes.
METHODS

Study Site Selection

Selection criteria

A number of tidal marshes in the Puget Basin were examined as potential study sites. Criteria for selection included:

1 - Environmental conditions are relatively unaltered by human activities. Many salt marshes have been diked, drained, dredged, or filled for agricultural and other uses. Nearby development alters marsh circulation and may result in erosion that destroys part of the depositional record.

2 - Full tidal circulation exists so that vegetation zones reflect current sea-level.

3 - There is proximity to an existing tide-gauge for future monitoring.

4 - A noncompressible substrate can be located under the marsh deposit at depths of less than 5 m. Dated samples of marsh deposits must be obtained at or very near this substrate to minimize the effects of compaction.

5 - There are minimal impacts from rivers with high sedimentation rates.

6 - The sites are located with respect to each other so
that the relative sea-level changes determined will provide meaningful data for the Puget Basin.

Reconnaissance

A reconnaissance of salt marsh sites was made during July and August of 1989. A number of sites were examined and researched in Jefferson County, northern Kitsap County, eastern Clallam County, Island County, San Juan County and Skagit County. On-site examination for obvious marsh alteration was followed by research of historical records to identify sites with potentially disrupted marsh stratigraphy. Marshes examined and found unsuitable for study are listed in Appendix I.

Marshes selected for study

The following marshes were selected as best meeting the criteria for study (Figure 6):

1 - Third Lagoon and Westcott Bay, San Juan Island

San Juan Island was determined to be tectonically stable during the historical period in a study by Holdahl and others (1989) of tide gauge records and precise leveling surveys. These sites provide data for the northwest part of the study area. The Friday Harbor tide gauge is nearby.

There are few salt marshes of appreciable size on the island due to a relatively narrow tidal range and to predominantly bedrock shores rather than bluffs of
Figure 6. Map of the Puget Basin and study site locations with nearby geographic features for each site.
unconsolidated glacial material. Many coves that might otherwise protect marshes are exposed to strong winds of long fetches, preventing marsh growth. Streams drain small watersheds and do not produce deltas on the order of those on the mainland.

Third Lagoon at American Camp National Historic Park was selected because the marsh in the eastern lagoon appears to be relatively unaltered. The uplands were logged in the early years of European and American settlement, but logging roads and skid trails lead to nearby Jakle’s Lagoon (Kunze, 1984) and the marsh at Third Lagoon appears to have been undamaged. Tidal exchange within the lagoon is complete. The substrate of the lagoon can be traced into the upland.

A marsh at the head of Westcott Bay was also selected for sampling. Uplands have been logged and developed for agriculture, but the marsh configuration appears unchanged from early maps. Tidal circulation in the marsh is good, and the substrate of the marsh deposits can be traced into a gently sloping eastern upland. Because marsh deposits on San Juan Island are consistently less than 0.5 m thick, it was impossible to obtain basal samples at two significantly different elevations for any one site. Two marshes were sampled in order to provide a better representation for this part of the Puget Basin.

2 - Donovan Creek, Linger Longer Road and Fishermans Point, Quilcene Bay, Jefferson County
Three marshes on Quilcene Bay in the northern part of the Hood Canal area were selected for study to provide data points on the Olympic Peninsula. The tidal flats of this north-south-trending bay are almost completely exposed at minus tides.

Part of a salt marsh has been preserved behind a sandspit at Fishermans Point on the southern tip of the Bolton Peninsula. Tidal exchange here is good. Deposits under the marsh are 3-4 m deep.

The marsh on Linger Longer Road formed in part as a deltaic deposits of the Big Quilcene River, although artificial levees have removed the fluvial influence. Deposits are 1-2 m thick.

The Donovan Creek site has shallow deposits about 1 m thick as well as a deeper part (3-4 m) behind a former berm. Tidal circulation has been disturbed by a road built on fill across the seaward margin of the marsh. The Port Townsend tide gauge is nearby. Basal peat samples from Donovan Creek at 3-4 m depth provide comparison with those from Fishermans Point, while samples from the shallower part of the Donovan Creek site can be compared with those from the marsh on Linger Longer Road.

3 - Sullivan-Minor marsh, Padilla Bay

A marsh adjacent to Bay View Ridge on the northeast shore of Padilla Bay was selected in order to obtain a data point on the mainland. Non-Indian settlement on mainland
shores in the Puget Basin occurred rapidly after 1840 and most tidal wetlands were altered for agricultural, port, or residential development. The Sullivan-Minor marsh was diked in the late 1800s but was recolonized after a few decades and since then has had full tidal exchange (Granger and Burg, 1986). This site was selected, in part, to assess the usefulness of an altered marsh in studies of relative sea-level change, as there are few unaltered tidal wetlands remaining on the eastern mainland shore.

Glacial sediment underlying the marsh can be traced to Bay View Ridge and is within coring range. The nearest existing tide gauge is at Friday Harbor.

Field Methods

Introduction

Relative sea-level changes at the study sites were determined from salt marsh stratigraphy and from radiocarbon dates on fossil peat samples. A brief history of the research that developed these methods is reviewed in the first part of this section, followed by a description of the methods used in field work.

Field work was divided into two phases:

1. Salt marshes were cored to determine underlying stratigraphy and to collect samples for radiocarbon dating.
Marsh depositional histories were reconstructed from cross-sections based on core data. Relationships of dated samples to former sea-levels are based on the identification of fossils of salt marsh plants.

2 - Present marsh surfaces were surveyed to determine the elevation of dated samples and to establish a baseline for monitoring future changes in sea-level.

Previous studies of relative sea-level history from salt marsh stratigraphy

Salt marsh stratigraphy and radiocarbon dating of plant fossils have been used to determine relative sea-level histories in previous studies. Jelgersma (1980) has used this method in several decades of research to determine Holocene shorelines in the Netherlands and the adjacent North Sea. The relative sea-level history of the Atlantic coast of the U.S. was reconstructed using these methods. Redfield and Rubin (1962) dated buried peat from a marsh at Barnstable, Massachusetts. Kaye and Barghoorn (1964) discussed the autocompaction of peat in their study of marsh deposits in Massachusetts. Bloom (1964) quantified the effects of compaction of peat in a salt marsh study in Connecticut. A regional curve of Holocene relative sea-level change compiled for Delaware Bay was based on a number of similar studies (Kraft and others, 1987; Fletcher and others, 1990).
Studies of Holocene relative sea-level changes on the west coast of North America include the work of Atwater and others (1977) in San Francisco Bay. Data points determined for the Puget Basin include the work of Eronen and others (1987) and Bucknam and Barnhard (1989). Clague and others (1982), Clague (1989) and Clague and Bobrowsky (1990) have dated tidal marsh deposits as a method in determining the relative sea-level history of the southern British Columbia coast.

Coring methods

Salt marshes were cored using a Eijelkamp hand corer to maximum depths of 5 m in order to determine underlying stratigraphy and to obtain basal peat samples for radiocarbon dating (Figure 7). Cores were examined for thicknesses of individual layers and the nature of the contacts between the layers. Layers were described in terms of sediment size, color, type of debris, sedimentary structures, vegetation species when possible, growth/nongrowth position of plant fossils, and where peat was compacted. Appendix II is a listing of core data used to construct cross-sections.

Principles of coastal geomorphology were used to reconstruct the depositional history of each marsh. Present-day net shore-drift directions were determined using geomorphic indicators suggested by Jacobsen and Schwartz.
Figure 7. Photo of sample core showing layers of gray clay and oxidized peat in the corer. The background is a low marsh surface of pickleweed.
Stratigraphic cross-sections generated from core data were analyzed for evidence of depositional changes. Buried tidal flat and salt marsh surfaces are assumed to be geomorphic indicators of the intertidal and supratidal environment based on the established relationship between sea-level and marsh vegetation zones.

*Distichlis spicata* and *Triglochin maritimum* are the indicator plant fossils used in this study to establish the relationship of a dated sample to former sea-level. These fossils are generally well-preserved and easily identified in Pacific Northwest salt marshes (B. Atwater, personal communication, 1989). Use of the fossils as indicators of former sea-levels is based on the assumption that the plants grew at approximately the same levels within the tidal range as they do today.

Eight basal peat samples were obtained for radiocarbon dating where the base of the marsh deposits could be traced into the upland. The one exception was at Fishermans Point, where the substrate could not be traced to the upland because fill covers the shallow part of the site. Bedrock is believed to underlie that part of the marsh (P. Spencer, written communication, 1989). Gravel at the base of the sample core and nearby cores is composed of shale similar to nearby bedrock exposures of the Bolton Peninsula siltstone (Spencer, 1983).

Large samples of peat (100-200 grams) were collected
due to the high sediment and water content relative to carbon content. Sufficient peat for each sample was collected by coring six to eight samples within a radius of 0.5 m of the original core. The vertical intervals sampled were 2 cm thick at both sites on San Juan Island and 5 cm thick at all other sites. Peat samples were taken from just above a basal layer of clay or sandy clay, at the first appearance of identifiable salt marsh fossils in growth position. This horizon is typically a silty peat layer that probably marks the transition from a sparsely-vegetated lower low marsh to the upper low marsh where *D. spicata* often dominates.

A sample dating an ash layer at Padilla Bay (see Field Observations) was compiled from 19 cores taken at the site. Grass stems and small pieces of wood within or in contact with the ash layer were collected and combined as one sample for dating, based on the assumption that the ash layer is a constant time horizon.

Historical maps, surveys, and air photos were examined for geomorphic trends during the past 50-100 years. This information was used in the analysis of marsh depositional histories. These sedimentation and erosion trends were then used to assess the response of each marsh to possible future sea-level rise. Assessments are based on the assumption that when sea-level rises, salt marshes will migrate inland only where the adjacent topography has a gentle slope and if
the rate of sea-level rise is slow enough that the marsh is not drowned (Titus, 1988).

**Surveying methods**

Each marsh was surveyed for three purposes:

1. to determine the elevation of dated samples;
2. to record the elevation of current marsh surfaces with respect to National Geodetic Vertical Datum of 1929, and to compare this with the nearest location for which tidal datums and N.G.V.D. elevations have been correlated;
3. to determine the vertical range of indicator plants.

Core sites were surveyed to determine sample elevations and to construct marsh profiles. Survey loops were closed with errors limited to 0.003 m (0.01 ft) at all sites except Third Lagoon. There the error is within 0.03 m (0.1 ft), because the high-tide water level was used as a turning point. Elevations referred to N.G.V.D. and tidal datums are given in both meters and feet; all other data are given in metric units only.

Elevations of high and low marsh surfaces were surveyed based on field-checked revisions of vegetation zones mapped in previous reports (Northwest Environmental Consultants, 1975; Kunze, 1984; Granger and Burg, 1986) for all marshes except Donovan Creek, where marsh surfaces were mapped and surveyed based on field observations. Marsh vegetation was observed from August 1989 to May 1990, and therefore does
not reflect plant assemblages limited to June and July. Maximum vertical ranges of indicator species were surveyed for all sites to establish the range of error in relating indicator fossils to former sea-levels.

The benchmarks used were referenced to the National Geodetic Vertical Datum of 1929. Although global sea-levels are estimated to have risen about 15 cm (5.9 in.) since 1929, the N.G.V.D. of 1929 is a terrestrial reference scale fixed to Mean Sea Level (MSL) as measured in 1929 and can be used to compare vertical accretion in marshes regionally and nationally (Titus, 1988). The elevations surveyed in this study may be used as baseline data for estimating vertical accretion in the event of an accelerated sea-level rise.

Tidal datums are based on tidal range, which varies from site to site. Mean Lower Low Water (MLLW) is the reference for bathymetry and tidal datums. In this study marsh elevations were determined by leveling to N.G.V.D. benchmarks, while correlations of N.G.V.D. with tidal datums are based on past surveys at the nearest tide-gauge location for which correlations have been made (Canning, 1990). The relationship of marsh surfaces to N.G.V.D. is therefore precise, while the relationship between marsh surfaces and the tidal datum scale is only approximate. Measurements of tides at each site were used to estimate the relationship between tidal datums at the tide gauge and true tidal datums at the study site.

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FIELD OBSERVATIONS

Third Lagoon, San Juan Island

**Geomorphic setting** (Figures 8 and 9)

The marsh lies in the eastern part of a lagoon that is sheltered by two sandspits (Figure 6). The spits are fed by eroding bluffs of glacial sediment on the east and west. Glacial sediment of Vashon recessional outwash (Eddy, 1975) forms low bluffs (approximately 1.5 m high) east of the lagoon, overlying the sheared rocks of the Lopez Structural Complex that are exposed in beach outcrops. The glacial deposits cover the contact between these rocks and the Upper Mesozoic Constitution Formation which may underlie part of the marsh site (Brandon and others, 1988). The bedrock is not exposed west of the lagoon; the eroding glacial bluff there rises to a height of approximately 15 m. Glacial erratics lie on the beach below this bluff.

Third Lagoon is one of a series of four lagoons that have formed along the southern shore of Griffin Bay. Longshore drift is bi-directional, but the easterly drift dominates slightly due to a greater fetch. Drift logs have accumulated in the western marsh and to a lesser extent in the eastern lagoon.

Freshwater contributions from the upland are minor. The primary source of marsh sediment is fine-grained
Figure 8. Third Lagoon. Map of vegetation (after Kunze, 1984) and location of core transect A-A'.

KEY

- Upland
- Berm
- Brackish
- High marsh
- Low marsh
- Tidal flat
- Core
- Sample core

Griffin Bay

Figure 8. Third Lagoon. Map of vegetation (after Kunze, 1984) and location of core transect A-A'.
Figure 9. Third Lagoon. Photo of study site looking southwest; the marsh is in the left part of the lagoon.

Figure 10. Third Lagoon. Photo of high marsh vegetation advancing into the upland. The tree on the left has died.
Marsh vegetation (Figure 8)

The marsh is dominantly a low marsh *Salicornia virginica* and *Distichlis spicata* community. Vegetated islands and tidal creek levees rise above an unvegetated surface of mud. Some of the islands are eroding and now only support *S. virginica*. A large part of the marsh is now an unvegetated tidal flat overlying peat deposits.

The lagoon is fringed by a narrow high marsh. High marsh and brackish marsh vegetation are mixed at the upper limit of the high marsh, and marsh vegetation is advancing into the forested upland (Figure 10). Cedar and hemlock trees are dying while the more tolerant spruce trees survive.

Tidal range (Figure 11)

The narrow range of elevations for marsh vegetation zones in the San Juan Islands is increased at Third Lagoon by the advancement of high marsh plants into the upland. The upper limit of the high marsh shown in Figure 11 is uncertain because of mixed vegetation patterns. The *D. spicata* range does not overlap into this mixed zone.

The tidal range of Griffin Bay is amplified because of the effect of shallow tidal creeks during the flow into the
Figure 11. Third Lagoon. Marsh elevations (N.G.V.D.) and tidal datums at Friday Harbor (1934-41). Elevation ranges are in meters.

Figure 12. Cross-section along transect A-A’ of Figure 8. Sample age is in radiocarbon years BP.
lagoon. A high tide measurement converted to N.G.V.D. at 0.6 m (1.9 ft) for Friday Harbor was surveyed at an elevation of 0.9 m (2.8 ft) in the eastern lagoon.

**Stratigraphy (Figure 12)**

Gray intertidal clay overlies gravel at the substrate of the lagoon. The sparsely vegetated, occasionally bioturbated clay grades into an oxidized peat layer rich in fossils of *D. spicata*. The dated sample was taken from silty peat near the base of this layer, at about 38-40 cm in depth, at the transition with the underlying gray clay. The surface brown silty peat layer has a distinct contact with the underlying oxidized peat, suggesting a rapid transition. However, plant fossils in growth position extending across the contact indicate that this transition was not sudden.

The cross-section shows the irregular profile of an eroding marsh (Figure 13), a pattern which is not observed at the contact between the surface and the underlying oxidized peat layer. The substrate of glacial material can be traced into the upland that approaches the lagoon at a moderate slope from the south and east.

**Depositional history**

The sparsely vegetated basal clay over gravel was deposited very near to low marsh elevations and developed into an oxidized, frequently exposed *D. spicata*-rich marsh.
Figure 13. Third Lagoon. Photo of eroding marsh surface at low tide showing remnant islands of low marsh vegetation.
at only 5-10 cm above the substrate. A radiocarbon age obtained for a sample from the base of this oxidized layer suggests that the original tidal flat became a salt marsh environment about 900 ± 80 years ago.

The cross-section (Figure 12) reveals peat and sand deposits at depth near the northeast shore that suggest an early fringing marsh. The lens of silty peat overlain by clay under a present channel suggests that this channel has maintained its position during marsh deposition. Continuous upward marsh growth and the unaltered tidal channel position indicate an undisturbed depositional history until the deposition of the surface layer of silty peat. This layer suggests that the marsh is unable to maintain an equilibrium with relative sea-level rise.

Compaction of peat may be responsible for the depositional changes to silty peat; compact peat was observed in several cores. However, the eroding surface and advancing vegetation suggest that the rate of relative sea-level rise may have increased. The surface erosion indicates that the low marsh would not be likely to survive an accelerated rate of sea-level rise and the high marsh would migrate to the eastern upland to survive as a fringing marsh.
Westcott Bay, San Juan Island

Geomorphic setting (Figures 14 and 15)

A marsh has formed at the head of Westcott Bay where fine sediment has accumulated behind a sand and gravel bar. The bay is shallow and tidal flats covering most of the bay are exposed at low tide. Sediment is derived from Sumas Stade glacial deposits (Eddy, 1975) that are eroding along the shore of Westcott Bay. Longshore drift brings the sediment to the marsh primarily from the southeast and, to some extent from the southwest, in this protected bay.

Freshwater drainage in the area has been altered. A stream enters the marsh from an artificial pond in the forested northern upland, and a stream from the eastern pasture has been channelled and dammed for a livestock pond. The stream alterations do not, however, appear to have significantly changed marsh dynamics. An historical map shows the marsh in the same general configuration at the time of non-Indian settlement (Richards, 1899).

Marsh vegetation (Figure 14)

A low marsh of *Distichlis spicata*, *Salicornia virginica* and *Jaumea carnosa* dominates the site. The pioneer species in the lagoon is *Triglochin maritimum*. Vegetation is generally continuous but sparse, and there are several unvegetated salt pans. Narrow strips of high marsh and
Figure 14. Westcott Bay. Map of vegetation (after Kunze, 1984) and location of cross-section A-A’.
Figure 15. Westcott Bay. Photo of study site looking northeast. The lagoon is exposed by a low tide.

Figure 16. Westcott Bay. Photo of dying upland vegetation where the high marsh vegetation zone is advancing inland.
Brackish marsh vegetation border the low marsh adjacent to the freshwater streams and in transition to the pasture. The narrow strip of high marsh at this transition is not mappable at the scale shown in Figure 14.

Brackish marsh vegetation is encroaching on the wooded upland in the northeast part of the marsh (Figure 16). Several small and medium-sized evergreen trees in this area have died as a result.

**Tidal range (Figure 17)**

The narrow range of marsh elevations reflects the relatively narrow tidal range of the San Juan Islands. However, a high tide field measurement at this site indicates a broader tidal range than at Friday Harbor. A high tide referenced to N.G.V.D. of 0.6 m (1.9 ft) at Friday Harbor was surveyed at 0.9 m (3.1 ft) at Westcott Bay.

At Westcott Bay, the greater distance inland, the shallow bay and the marsh position at the head of the bay all contribute to the amplified tidal range. The tidal range here is also larger than at Third Lagoon, where the relative proximity to open waters and the absence of significant tidal resonance result in a narrower tidal range.

**Stratigraphy (Figure 18)**

The cross-section reveals gray clay over nearshore beach gravel grading upward through silty peat to an
Figure 17. Westcott Bay. Marsh elevations (N.G.V.D.) and tidal datums for Friday Harbor (1934-41). Elevation ranges are in meters.

Figure 18. Westcott Bay. Cross-section along transect A-A' of Figure 14. Sample age is in radiocarbon years BP.
oxidized surface layer of peat. The clay is unvegetated at the base and sparsely vegetated near the gradational contact with the silty peat. Glacial deposits underlying the marsh deposits can be traced from the sample core to the southeastern upland pasture. Nearshore basal gravel was not traced further into the marsh as the very compact clay was impenetrable by hand coring methods.

The dated peat sample came from the transition from vegetated clay to peat at a depth of 38-40 cm, the deepest level at which *D. spicata* fossils were found. Twenty centimeters of clay separated the sample from basal gravel.

**Depositional history**

The gradational sequence of unvegetated to vegetated clay to silty peat and then to a surface peat layer indicates that the marsh evolved from a tidal flat with no apparent disturbance. The eastern upland was probably initially flanked by a gravel beach from which the berm now protecting the marsh gradually extended to the northwest. A radiocarbon age from the base of the silty peat horizon indicates that the marsh developed from a tidal flat in the lagoon sheltered by this berm about 520 ± 70 years ago.

Although marsh vegetation is advancing into the forested upland, the low marsh does not appear to be eroding. The advancing vegetation may indicate a relative sea-level rise, yet the low marsh is surviving by upward
growth. Air photos do not show appreciable changes in marsh size for the past 25 years, and changes in vegetation zones are not discernible. A marsh in nearby Garrison Bay (Figure 6) is also encroaching on the upland but is eroding at the shoreline, perhaps due to a less protected setting and a poor sediment supply.

The uniform thickness of these marsh deposits and those at Third Lagoon, as well as at marshes on Garrison Bay, Mitchell Bay and Davison Head (Figure 6), suggests that a relative sea-level rise of less than 0.5 m has occurred since late Holocene tidal ranges and sedimentary environments on San Juan Island became suitable for marsh development and survival. The radiocarbon date at the base of this deposit at Third Lagoon is several hundred years older than that at Westcott Bay. The age difference for two samples from similar depths suggests that compaction may have occurred at Third Lagoon. However, the older age at Third Lagoon can also be explained by erosion at the surface, by differences in tidal range, or by radiocarbon dating errors.
Fishermans Point, Quilcene Bay

Geomorphic setting (Figures 19 and 20)

The marsh lies behind a hooked sandspit at the base of a steep, forested cliff at the southwest tip of the Bolton Peninsula (Figure 6). The cliff is composed of glacial sediments overlying the Eocene Bolton Peninsula siltstone identified by Spencer (1983) as an informal member of the Twin Rivers Group. The steeply dipping, poorly-indurated siltstones are eroded to form a wave-cut platform on the southern shore of the peninsula (Figure 21).

Spencer believes that the intertidal area around and including the spit is built on a wave-cut platform of these easily eroded rocks (P. Spencer, written communication, 1989). Cores at the rear of the marsh reached the substrate at 167 cm below Mean Sea Level, approximately 4 m from the surface. Gravel recovered from the substrate in cores is siltstone and shale similar to that feeding the spit from the eroding point.

Exposures of Bolton Peninsula siltstone extend north of Fishermans Point in the cliff along Quilcene Bay. Basaltic rocks of the Crescent Formation across the bay indicate that the contact between these rock units lies under Quilcene Bay.

The stratified glacial sediments observed in roadcuts and overlying the rocks along the shoreline are identified
Figure 19a. Fishermans Point. Map of vegetation and study site (after Northwest Environmental Consultants, 1975).

Figure 19b. Map of the southern part of the high marsh with core transects A-A' and B-B'.
Figure 20. Fishermans Point. Photo of study site looking north.

Figure 21. Fishermans Point. Photo of rapidly eroding siltstones at the southern end of the Bolton Peninsula.
by Gower and others (1985) as pre-Fraser continental ice sheet deposits. These sediments strike N30W and dip 33 degrees to the NE, are locally deformed, and are cut by small high-angle reverse faults striking N35W that are upthrown from NE to SW. Two small reverse faults in an exposure of the Bolton Peninsula siltstone in the tidal creek entrance to the lagoon strike N30W and N40W and dip shallowly to the NE. This evidence suggests that Holocene vertical crustal movement at this site is a possibility. Gower and Spencer both suspect a fault in Quilcene Bay.

The original marsh behind the spit has been partly covered by fill, and the spit is eroding at the narrow segment south of the hook. Surveys, maps and photos show a broad berm with a road on it until 15 to 20 years ago. An abundant sediment supply from the eroding point appears to be uninterrupted, so the erosion may be due to changing directions of wave energy as the tip of the peninsula erodes and leaves that part of the spit more vulnerable. Without protection or nourishment, the spit may be breached in the future. The vulnerability of the eroding spit and the steep slope of the bluff suggest that this marsh would not survive an accelerated sea-level rise.

**Marsh vegetation** (Figure 19)

The plant succession moves from lagoon tidal flat into a *S. virginica*-dominated low marsh and then to a high marsh
that fringes the lagoon and covers a larger area in the lee of the point. Grass adjacent to the area of fill is at the elevation of the high marsh and marks an early transition to upland vegetation which is probably due to freshwater runoff.

Tidal range (Figure 22)

At this site Triglochin maritimum grows only on the high marsh and not as a pioneer on tidal flats. The range for the low marsh is thus less than it is at marshes on Linger Longer Road or Donovan Creek where *T. maritimum* extends the lower limit of the low marsh, and at Donovan Creek where the tidal range is amplified by a location at the head of the bay.

The tidal range at Fishermans Point is amplified by the shallow tidal channel and lagoon across which the water travels. A tide of 1.0 m (3.3 ft) at Zelatched Point was surveyed at 1.4 m (4.6 ft) at the Fishermans Point marsh, with measurements referenced to N.G.V.D.

Stratigraphy (Figure 23)

Cross-sections show that the deeper deposits under the high marsh grade from basal clay over gravel upward to peat. The clay layer at the base of A-A' is unvegetated or sparsely vegetated. The clay grades upward into the silty peat and peat deposits, with vegetation in growth position
Figure 22. Fishermans Point. Marsh elevations (N.G.V.D.) and tidal datums for Zelatched Point (1965-66). Elevation ranges are in meters.

Figure 23. Fishermans Point. Fence diagram showing cross-sections along A-A' and B-B' of Figure 19b. Sample age is in radiocarbon years BP.
and fragments of wood that were incorporated as debris at the transitional contacts. Peat deposits begin at a deeper level near the spit than at the bluff. The basal clay of cores 1 and 2 contains horizontally-oriented grasses.

Abundant fossils of *T. maritimum* and *D. spicata* are found at the horizon of the dated sample, just above the transition to silty peat. Lenses of peat near the base of B-B' are also rich in salt marsh plant fossils. The peat is compact in deposits below about 0.5 m above MSL. A mid-level clay lens at approximately MSL has transitional contacts with overlying and underlying peat and is sparsely vegetated.

Several thin clay layers interrupt the upper layers of peat and contain vegetation in growth position. Contacts are dominantly transitional for all but one of these layers. The 1-15 cm thick clay layer at an elevation of approximately 1 m above MSL is separated from underlying peat by a distinct parting surface in 5 of 11 cores. Gravel overwash and a thicker clay deposit occur at a similar elevation near the spit. Charcoal was observed above and/or below this layer in 7 of 11 cores. Other occasional parting contacts were noted but none was repeated at a consistent elevation.
Depositional history

The deposits at this site record a relative sea-level that has risen about 3 meters since the original marsh became established. A sample from the base of peat deposits at the rear of the marsh yielded a radiocarbon age of 2950 ± 70 years ago. The marsh probably originated under the area now covered by fill and advanced to this part of the early lagoon at about that time.

The thicker clay at depth near the lagoon indicates a gradual extension of the marsh in that direction. Thicker clay at depth near the bluff and slightly lower surface elevations suggest that a marsh fringed the spit while tidal mud was deposited near the bluff in tidal channels, similar to the present configuration.

Lenses of peat within basal clay in B-B' (Figure 23) probably developed as islands of an early marsh expanding into the lagoon. The mid-level lens of clay may represent a former tidal channel or local subsidence caused by compaction of underlying peat.

The 1-15 cm thick clay layer associated with a parting contact may represent a subsidence event due to compaction from seismic or other effects. Alternatively, associated charcoal may represent a fire that could have temporarily lowered the surface elevation on the dry marsh, allowing greater tidal inundation and deposition of clay. However, gravel overwash near the spit suggests a storm-damaged berm.
as a probable cause. This hypothesis explains the parting contacts with horizontal grasses, and the deposition of burned wood transported by storm waves.

The general upward growth of the marsh associated with a gradual relative sea-level rise appears to be unbroken by any other event, and stratigraphic evidence suggests the most likely cause of the clay layer with notable parting contacts is a temporarily breached or damaged berm. The gravel overwash occurs near the part of the berm that is presently eroding and vulnerable to breaching.
Donovan Creek, Quilcene Bay

**Geomorphic setting** (Figures 24 and 25)

Thorson (1989) believes that during glacial recession of the Puget Lobe, about 14,000 to 13,500 years ago, the linear valleys between Quilcene Bay and Discovery Bay to the north functioned as a major spillway draining a huge lake impounded in Puget Sound by the retreating ice sheet. Thorson's Leland Spillway controlled the drainage of post-glacial Lake Bretz until lake levels were lowered, the crust rebounded, and the Puget Lobe ice dam broke up.

Deltaic deposits of an early Quilcene River that drained into Lake Bretz now lie above sea-level. These late Fraser deposits as well as pre-Fraser deposits that mantle the Bolton Peninsula (Gower and others, 1985) provide the sediment that built a broad tidal flat at the head of the bay onto which prograded the deltas of the Big Quilcene and Little Quilcene Rivers.

The marsh has developed where Donovan Creek drains the southern part of the former Leland Spillway into the north end of Quilcene Bay (Figure 6). The partly buried berm of a former barrier beach in the eastern part of the marsh suggests that the head of the bay may have filled in behind nested berms similar to those currently seen at Tarboo Bay, the next bay to the east (Figure 26). Barrier landforms trending upstream are formed in these bays by longshore
Figure 24. Donovan Creek. Map of vegetation and location of core transects A-A' and B-B'.

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Figure 25. Donovan Creek. Photo of eastern part of study site at the head of Quilcene Bay, looking northeast. The creek enters the bay from the left, and the buried berm is just right of this in a lighter color than the green marsh.
Figure 26. Photo of the heads of Quilcene Bay (left) and Tarboo Bay (right) with nested berms and spits trending upstream at Tarboo Bay. North is at the top of the photo.
drift from south to north. Wave energy from the south is high due to an approximately 50 km fetch the length of Quilcene Bay, Dabob Bay and Hood Canal.

A small freshwater stream draining the eastern slope has probably redistributed sand and gravel of the berm. There may therefore be just one former berm, although a former berm from a lower sea-level may be buried by marsh deposits. The deeper deposits probably fill a former channel of the creek.

Before non-Indian settlement in the 1900s, this marsh and those at the mouths of the Big Quilcene and Little Quilcene Rivers had merged and were prograding southeastward, filling in the northwest part of the bay (Figure 27). Parts of the larger wetland are now behind levees, and a road built on fill crosses the Donovan Creek marsh where it meets the bay. Maps show a change from a road on a trestle to a road of fill between 1947 and 1953, marking the approximate date that full tidal influence was disrupted. Tidal exchange with the bay now occurs by way of two large culverts under the road, which have altered sedimentation patterns and tidal circulation.

Marsh vegetation (Figure 24)

Vegetation patterns indicate that construction of the road has disrupted the natural biota, and survival of the salt marsh may require more adequate drainage. Patches of
Figure 27. Historical map of Quilcene Bay as observed before non-Indian settlement (U.S. Ex. Ex., 1841a).
brackish marsh vegetation near the culverts and in the western marsh near the road are probably due to ponding of water behind the road. The brackish marsh is encroaching on the evergreen and alder trees in the western upland. The present high marsh includes a thinly vegetated meadow of *Distichlis spicata* on the eastern marsh, while identical elevations in the southwestern part have multiple species or brackish influence.

The marsh surface is very compact, and parts of it have been used for pasture. A channelled freshwater stream from the eastern upland is bordered by brackish marsh vegetation. High marsh species grade to dune vegetation on the partly buried berm deposits.

*Tidal range* (Figure 28)

The tidal range is amplified by the marsh location at the head of a long, narrow bay and the shallow tidal flats that cover most of the northern bay. A high tide of 0.6 m (1.9 ft) at Zelatched Point was measured at 2.5 m (8.1 ft) at the beach south of the road, with measurements referenced to N.G.V.D. This extreme amplification is probably due to a component of tidal resonance in Quilcene Bay, and the tidal range is even further amplified when the water enters the marsh.

The wider tidal range is reflected in the wide range of elevations for the low marsh. The range for *D. spicata* and
Figure 28. Donovan Creek. Marsh elevations (N.G.V.D) and tidal datums for Zelatched Point (1965-66). Elevation ranges are in meters.
Triglochin maritimum together is 0.51 m, but for T. maritimum alone it is 1.49 m because it extends from the pioneer position on tidal flats to the high marsh.

**Stratigraphy (Figures 29 and 30)**

The marsh deposit is divided into a deeper eastern part and a shallower western part. Cross-sections A-A' and B-B' show the deeper eastern marsh deposits to be dominantly gray intertidal clay with lenses of peat and some sandy overwash. The blue-gray clay is unvegetated at the base but contains sparse grasses in growth position and lenses of silty peat at higher levels. The dated sample in cross-section B-B' comes from a compact lens of sandy, silty peat containing D. spicata fossils in growth position. The substrate of glacial material is traced into the gently sloping eastern upland except for a gap at the base of the former channel where the clay is too compact to core by hand.

Surface peat layers were more compact than most of the deeper marsh deposits. This is probably due to land use as pasture.

The western marsh deposits overlie beach sand and fine-to-coarse gravel which were deposited by relatively high wave energies. Cross-section A-A' shows beach sand and sandy clay grading upward to vegetated clay, silty peat, and the present marsh peat. The dated sample in A-A' contains
Figure 29. Donovan Creek. Cross-section along core transect A-A' of Figure 24.

Figure 30. Donovan Creek. Cross-section along core transect B-B' of Figure 24.
fossil *T. maritimum* which is abundant at this horizon. A thick clay horizon seen in B-B’ at mid-level under the western marsh contains grass roots and stems in growth position and multiple 1-2 cm thick lenses of peat and silty peat. Contacts between horizons are transitional.

**Depositional history**

The deeper deposits under the eastern marsh probably accumulated in a former stream channel protected by a berm about 3170 ± 90 radiocarbon years ago. The sandy, silty peat sample that returned this date came from the deposit of a briefly established marsh. Peat horizons near the buried berm in A-A’ (Figure 29) suggest a tidal flat lagoon fringed by a marsh.

The shallower western marsh deposits began forming over beach sands about 930 ± 60 radiocarbon years ago, probably at the margin of the advancing delta of Donovan Creek. The clay horizon with peat lenses seen in B-B’ (Figure 30) continued as a sparsely vegetated tidal flat near low marsh elevations until more recently. This part of the marsh may have formed in the lee of a bar or shoal not located by coring, or buried by the road fill, so that the inner lagoon filled in more slowly. Lenses of peat deposits in the clay layer probably represent islands of marsh vegetation that grew at a slightly lower elevation than the established marsh.
Linger Longer Road, Quilcene Bay

**Geomorphic setting** (Figures 31 and 32)

The marsh is on the margin of what was once a larger marsh that covered the northwestern shores of Quilcene Bay (Figure 27). The sediment was in part a deltaic deposit of the Big Quilcene River, which is now artificially channelled to the north of the marsh (Figure 6). Eroded glacial sediment of both pre-Fraser and Vashon origin is brought by longshore drift from bluffs to the south. Storm waves entraining tidal flat sediments may also nourish the marsh.

To the west a steep bluff rises above the marsh. To the north and northwest former wetlands are behind levees and beyond tidal influence.

An artificial berm, built on what may have been an offshore shoal, extends through the marsh from the base of the steep upland, reducing wave energy to the north part of the marsh. Many drift logs lie on the southern exposed marsh and to a lesser extent at the rear of the northern marsh.

Tidal exchange is by way of several tidal creeks. Freshwater input is minimal. The surface of the tidal flat along the shoreline south of the marsh has been partially reworked by dredging for the oyster industry, but the northern marsh has not been significantly affected.
Figure 31. Linger Longer Road. Map of vegetation (after Northwest Environmental Consultants, 1975) and location of core transect A-A'.
Figure 32. Linger Longer Road. Photo of study site looking west. The marsh is on the lower left and tidal flats are exposed by low tide.

Figure 33. Linger Longer Road. Photo of high marsh and house at the end of the berm.
Marsh vegetation (Figure 31)

The low marsh Triglochin maritimum is advancing into the bay as sediment from longshore drift increases tidal flat elevation. High marsh vegetation covers most of the protected northern marsh and is continuous except for one salt pan (Figure 33). Brackish marsh vegetation at the rear of the marsh may receive freshwater input from the road.

Tidal range (Figure 34)

The tidal range at this marsh is less than that at Donovan Creek, where tidal resonance has a pronounced effect. A high tide at the Linger Longer Road marsh was measured at 0.81 m (2.67 ft) for a predicted height of 0.69 m (2.27 ft) at Zelatched Point, after conversion of tidal measurements to N.G.V.D.

The range for the low marsh includes T. maritimum at the lower limit. The range for the occurrence of Distichlis spicata and T. maritimum together extends from the upper low marsh to the upper high marsh.

Stratigraphy (Figure 35)

Beach sand and gravel underlie the nearshore deposits. The substrate of the marsh rises steeply to meet the road along the base of the bluff east of the marsh. Sandy clay overlies the basal gravel, except where a clay deposit appears to have accumulated in the lee of a 0.5 m
Figure 34. Linger Longer Road. Marsh elevations (N.G.V.D.) and tidal datums at Zelatched Point (1965-66). Elevation ranges are in meters.

Figure 35. Linger Longer Road. Cross-section along A-A' shown in Figure 31. Sample age is in radiocarbon years BP.
topographic rise. The rise may have been a shoal or a berm built by longshore drift from the south. A peat horizon overlying sandy clay in contact with the rise indicates that an early salt marsh was established on this topographic high.

The dated sample containing *T. maritimum* fossils was taken from the base of a silty peat layer grading into the peat deposit of the early marsh. Deposits of silty peat and sandy clay overlying the peat grade into the present marsh surface.

Several partings occur in association with thin layers of sand. One parting contact at about 50 cm depth occurs in 5 of 14 cores and is associated with decreased vegetation just above the parting in 3 of 5 cores. Other cores show no apparent depositional changes at this elevation. Parting contacts associated with overlying layers of sand occur in other cores but are not consistent across the marsh; they occur only in cores near the bayward margin and along the tidal creek. Small pieces of wood debris are incorporated into the cores at all levels.

Depositional history

The marsh developed as a prograding deltaic feature modified by high wave energy and longshore drift from the south. The early marsh probably formed over the topographic high at about 630 ± 60 radiocarbon years ago, as suggested
by a sample taken from the silty peat grading into the deposits of the early marsh.

This early marsh was probably covered by sand and clay during a storm event. Until that event an area of low marsh and sandy tidal flat appears to have existed between this rise and the shore, suggesting the presence of a protective bar. Gradual upward growth with relative sea-level rise culminated in the present marsh surface.

Parting contacts probably reflect storm events. The fetch at this site is about 50 km from the southern end of Hood Canal, and residents of the house on the berm (Figure 33) report strong winds and waves during storms. Because the partings are not consistent across the marsh it is unlikely that they result from vertical crustal movement.
Padilla Bay, Skagit County

Geomorphologic setting (Figures 36 and 37)

The Sullivan-Minor marsh at the base of Bay View Ridge has developed on a wave-cut terrace eroded during post-glacial sea-level rise. Siegfried (1978) mapped Bay View Ridge as re-worked beach gravels of Everson glaciomarine drift overlying Vashon till. Siegfried mapped three terraces on the ridge that formed during post-glacial isostatic rebound.

Longshore drift is from south to north (Keuler, 1979), and is modified just north of the site by the drainage from Leary Slough (Figure 6). Sediment nourishing the marsh is derived from Leary Slough to the north and from eroding bluffs to the south. Exposures of sediment identified by Siegfried (1978) as both Everson glaciomarine drift and Vashon till lie along the shoreline of these eroding bluffs. Sediment was derived from distributary channels of the Skagit River delta, as well, until drainage into Padilla Bay was cut off by early non-Indian settlers.

The delta was diked for agricultural purposes beginning in the 1850s. The northern salt marsh sits behind what remains of an artificial levee built by early farmers (Figure 38). The altered marsh was farmed for several decades before it was recolonized by halophytic vegetation following a levee failure early in this century (Granger and
Figure 36a. Padilla Bay. Map of study site.

Figure 36b. Padilla Bay. Map of northern marsh showing vegetation (after Granger and Burg, 1986) and core transect A-A'.

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Figure 37. Padilla Bay. Photo of study site looking east.

Figure 38. Padilla Bay. Photo of northern marsh surface looking north along the artificial berm.
This northern salt marsh is partly protected by pilings that curve bayward at the southern part of the marsh, suggesting the limits of what may have been a natural berm or shoal extending north from the protruding bluff to the south.

To the south of the salt marsh there is a brackish marsh that is protected by artificial berms. There is some tidal influence by way of a ditch that follows the base of the bluff. Farther south, a small forested residential community sits behind the beach in the lee of the protruding bluff and is protected by seawalls and pilings.

Drift logs cover the northern part of the recolonized northern salt marsh. A freshwater creek enters the marsh from the bluff. Tidal creeks are probably not in their original drainage pattern; a rectangular pattern suggests they are left from the period of agricultural use.

**Marsh vegetation (Figure 36b)**

*Salicornia virginica* dominates the low marsh in the center of the northern marsh where surface elevations are lowest. Most of the northern marsh is otherwise dominated by a *S. virginica-D. spicata* association with minor *Atriplex patula* at a transitional level between the low and the high marsh. High marsh vegetation occurs primarily in the southern part where the elevation is slightly higher.
Brackish, transitional marsh vegetation grows at the base of the bluff where freshwater drainage enters the marsh. The drift logs at the extreme northern end have a sparse growth of low marsh *S. virginica* between them.

**Tidal range** (Figure 39)

Marsh surfaces lie at lower elevations at this marsh than they do at the other study sites. Present marsh surface elevations may not reflect the natural conditions because artificial ditches and levees have altered tidal circulation. The vertical range for *D. spicata* alone includes the range for *T. maritimum*.

A high tide of 1.5 m (4.8 ft) for the north end of the Swinomish Channel where it enters Padilla Bay was measured at 0.4 m (1.3 ft) on the beach in front of the residences at the study site. Tidal heights are referenced to N.G.V.D. The tidal range at the Swinomish Channel is probably higher due to a longer expanse of shallow tidal flat that separates it from the entrance to the bay. A component of tidal resonance due to the shape of the bay may also account for the 1.1 m difference in tidal height.

**Stratigraphy** (Figures 40, 41 and 44)

The deepest clay layer shown in cross-section A-A' (Figure 40) across the northern marsh is inferred in places because of difficulty in penetrating the thick layers of
Figure 39. Padilla Bay. Marsh elevations (N.G.V.D.) and tidal datums for the northern Swinomish Channel at the entrance to Padilla Bay (1955). Elevation ranges are in meters.

Figure 40. Padilla Bay. Cross-section along A-A' of Figure 36b. Sample ages are in radiocarbon years BP.
gray intertidal clay. Maximum depths of cores are indicated where they did not reach the substrate. Where cores did penetrate this clay they encountered lenses of compact peat up to 30 cm thick, but generally 10-20 cm thick, containing fossils of *D. spicata*.

The substrate slopes steeply away from the bluff and is overlain by sand and gravel. Forest debris of twigs and pine cones occurs in these cores. The deeper dated sample taken from above this substrate contained fossil *D. spicata*; the shallower sample contained both *D. spicata* and *Triqlochin maritimum*.

Mid-level layers of peat separate the three intertidal clay layers. Contacts between these layers are dominantly transitional. Occasional parting contacts lie at varying elevations. Stratigraphy near the bluff may represent slump structures generated either by seismic activity or by slopewash. Layers shown as undifferentiated organic material are dark brown and have a friable texture, suggesting they were derived from the bluff.

Cores were taken outside of the northern marsh in order to determine the depositional history of the site. Cross-section B-B' (Figure 41) shows the stratigraphy under the beach south of the marsh, where at least twenty-four nearly-eroded tree stumps in growth position line the beach for about 150 m (Figure 42). An ash layer lies 70-100 cm below the organic horizon containing both the stumps and the
Figure 41. Padilla Bay. Cross-section along core transect B-B' of Figure 37.
Figure 42. Padilla Bay. Photo of eroded tree stumps on the beach in front of the residences. Three tree stumps are in the foreground in front of field assistant Shawn Doan.

Figure 43. Padilla Bay. Photo of exposed peat eroding on the beach in front of the northern marsh, showing the contact with the underlying clay layer containing the ash.
living forest, suggesting that the stumps and the living forest are of the same generation of trees. A boring of a living spruce tree behind the beach gives a minimum dendrochronological age of 140 years.

The pale gray ash layer forms an approximately 1-cm-thick horizon in a layer of sparsely-vegetated gray intertidal clay. Fossil grass stems and leaves lie horizontally in contact with the ash layer.

The southern limit of cross-section B-B' is traced onto the sand and gravel beach at the protruding bluff. The only core that penetrated to 5 m revealed intertidal clay with two interbedded sand layers. The ash layer shallows at the northern limit of the cross-section, lapping onto a shallow sand and gravel deposit that crosses the beach and extends into the northern salt marsh near the curved pilings. Silty peat grading upward to peat fringes this sand and gravel deposit. Cores following the ash horizon bayward onto the tidal flat indicate that it continues for approximately 20 m offshore. The sand layer underlying the clay layer that contains this ash continues at least 30 m bayward under tidal clay after the ash layer disappears.

The eroding peat layer at the surface is exposed in several places along the beach (Figure 43). No identifiable salt marsh biota was found in any core of this layer, but the sediment has a porous, matted and organic texture and an oxidized color that suggests it formed by vertical plant
growth. Abundant wood debris is associated with this layer.

Cross-section C-C' (Figure 44) across the southern brackish marsh also contains the ash layer. The substrate slopes away from the bluff at a steep slope similar to that in the northern marsh. Micaceous sand at 5 m below the surface grades upward to blue-gray intertidal clay which is occasionally vegetated. Overlying this, a mid-level layer of gray micaceous sand grades upward to the clay layer in which the ash was deposited. The upper clay horizon contains vegetation in growth position. The D. spicata-bearing layer of surface peat is interrupted by a thin layer of clay with sparse vegetation in growth position.

Outside of the northern marsh no significant buried organic horizon was found except near the residences. Two cores not shown in cross-section encountered a sand horizon 0.7 to 0.9 m thick overlying a basal dark brown organic layer. There is a sharp contact between layers. Intertidal blue-gray clay up to 1 m thick overlies the sand. These two cores reached a base level of sand and gravel at approximately 4 m in depth. Debris in the buried organic layer includes wood and conifer needles; no marsh biota were noted.

Depositional history

Marsh stratigraphy suggests that an offshore shoal or spit extended northward from the protruding bluff,
Figure 44. Padilla Bay. Cross-section along transect C-C' of Figure 36a.
protecting the southern part of the site. Offshore bars and shoals observed in air photos taken at low tide could represent relict platforms of an early feature that migrated landward. Eroding peat deposits seaward of the present levee are evidence that the marsh was once wider than it is now. Sand layers at depth shown in cross-section B-B' (Figure 41) may represent former shoals.

An early map (U.S. Ex. Ex., 1841b) does not show an outer spit but does show a marsh fringing the bluff. The rate of sedimentation from the south was higher before Skagit River drainage was diverted from Padilla Bay to Skagit Bay by pioneer farmers. More recently, construction along the Bay View State Park shore has disrupted sediment transport from the south, but the eroding bluff continues to nourish the site with some sediment.

The line of curved pilings marks the approximate northern limit of the ash layer in cores, at a point where the ash layer lies in silty peat that suggests a fringing marsh and shallows to lap onto sand and gravel deposits. Fine horizontal grasses and distinct contacts between the ash and clay indicate that the ash was deposited in a low energy environment. The shallowing ash layer, the sand and gravel deposits that cross the beach into the marsh, the fringing peat deposits and the line of curved pilings probably mark the northern limit of the shoal or berm where it widened or curved landward.
The lack of an ash layer under the northern beach and marsh may indicate a less protected environment at the time of ash deposition, so that higher wave energy might have washed the ash away. Alternatively, the ash may have fallen on the mid-level peat layer (Figure 40), resulting in poor preservation of the ash after compaction.

Organic matter found in contact with the ash is dated at 2750 ± 90 radiocarbon years BP, which falls within the time span of the Mt. St. Helens P series (Sarna-Wojcicki and others, 1983). In a microprobe analysis of the glass in the ash, the ash does not correlate well with the St. Helens P series (S. Cornelius, Washington State University, written communication, 1990). Good correlations are made with St. Helens reference ashes T and Jb, dated at 200 and 11,000 years old, respectively. Neither of these ages seems likely, given the stratigraphic position of the ash layer and the 90-year standard deviation of the radiocarbon date. The fossil organic fragments in the radiocarbon dating sample included intertidal plant fossils in growth position.

The best correlation for the Padilla Bay ash found in the microprobe analysis is with ash Mb-2, an ash from McClellan Bog, located southwest of Mt. St. Helens. The Mb-2 ash is undated and has no stratigraphic control (S. Cornelius, written communication, 1990). Until further work is done on the Mb-2 ash, the radiocarbon date of 2750 ± 90 BP is assumed to be more reliable for the Padilla Bay ash.
layer. However, the ash may belong to a previously unidentified ashfall.

The original northern marsh at Padilla Bay was established where sediment deposited by Leary Slough mixed with sediment brought by northward drift. This part of the site has alternated between a salt marsh and a tidal flat, with what appear to be two marsh-wide changes in depositional conditions in the past 4000 years.

Dated samples from the northern marsh indicate that the lower layer of peat lenses formed in the intertidal zone about 4170 ±70 radiocarbon years BP. The peat lenses represent either the occasional localized marshes of an emerging wetland or the vegetated islands of an irregular, eroding marsh surface.

A radiocarbon age for silty peat lying 1.6 m higher than these lenses stratigraphically indicates that most of this part of the site emerged from tidal flat to salt marsh about 3750 ± 80 years ago. The two mid-level layers of clay in the northern marsh may represent either periods of relative sea-level rise or changes in the depositional setting caused by a shifting berm or shoals to the south.

The original southern marsh probably developed from south to north, gradually extending from the protruding bluff. It was most likely protected by a shoal or berm that migrated landward, as previously discussed.

The buried organic deposits near the residences may
correlate with one of the buried peat layers in the northern marsh. Elevations cannot be compared because they may have been altered by compaction evident in the peat deposits. Forest debris in the organic layer suggests that the buried organic horizon may be a freshwater deposit, but this debris may also be derived from the nearby bluff. These buried organic deposits are probably coeval with the peat lenses at depth in the northern marsh, and the lack of an organic horizon at depth in the marsh between them (Figure 44) indicates that marsh vegetation at both ends of the site joined to form a continuous marsh at a later time.

The absence of salt marsh fossils in this deep organic layer, and at the base of the surface organic layer containing the forest and eroded stumps, does not necessarily indicate they formed after a sudden transition to freshwater conditions. Buried salt marsh deposits under forests on the Washington coast have a similar absence of preserved fossils which is attributed to a rapid rate of organic decomposition in the acid conditions of a forest soil (B. Atwater, personal communication, 1990). The deeper organic layers at Padilla Bay could therefore represent salt marsh deposits that evolved to freshwater conditions. If this deep peat layer correlates with one of those in the northern marsh, then overlying clay horizons represent a relative sea-level rise.

The eroded tree stumps along the southern beach and the
Eroding peat along the northern beach suggest that a rise in relative sea-level has occurred in the past several hundred years. Old-growth cedar stumps in the living forest near the residences currently sit at low elevations in wet soil behind protective bulkheads. Although a relative sea-level rise is possible, erosion on the beach may be related to the reduced sediment supply from the south. Alternatively, the underlying peat horizons may have undergone gradual compaction due to the weight of the forest as the trees grew to a large size.

The existing salt marsh is unlikely to survive an accelerated rate of sea-level rise. Marsh vegetation could expand across the residential area if access were allowed, but the steep bluff along the bay would prevent further migration inland.
INTERPRETATIONS

Stratigraphy

Introduction

Stratigraphic evidence for the marshes examined indicates a rise in relative sea-level at all sites. The elevation of present marsh surfaces above basal salt marsh peat deposits is evidence for this rise. No conclusive evidence of sudden vertical crustal movement was observed at the six marshes studied.

Cross-sections are often subject to more than one interpretation of depositional history in the Puget Basin because of variable wave and sedimentation patterns as well as local and region-wide differences in tidal range. Alternate interpretations of relative sea-level changes based on core data, cross-sections and geomorphic observations are presented in this section for each of the study sites.

San Juan Island

On San Juan Island, marshes at Westcott Bay and Third Lagoon have evolved with relatively undisturbed upward growth. A recent relative sea-level rise is suggested at Third Lagoon by an eroding surface and by a marsh-wide
surface horizon of silty peat over an oxidized peat deposit. The inability of the marsh at Third Lagoon to survive this rise in the apparent absence of evidence for a change in the rate of sedimentation suggests that the rate of sea-level rise may have increased. Advancing marsh vegetation zones at both sites and at Garrison Bay on San Juan Island may be the result of salt-water intrusion, although vegetation on shorelines adjacent to the marshes appears unaffected. Erosion of marsh surfaces supports the hypothesis of a relative sea-level rise.

Quilcene Bay
Marshes at Quilcene Bay developed at an earlier date than those on San Juan Island, probably because the larger tidal ranges in the Hood Canal area reached amplitudes sufficient for salt marsh evolution at an earlier date. Deposits at Quilcene Bay are thicker than those on San Juan Island and contain evidence of changing depositional conditions that cannot be conclusively attributed to any one agent of change.

Evidence in these marsh deposits suggests that storm activity may have generated one of the depositional changes at both Fishermans Point and Linger Longer Road. Many Puget Basin tidal marshes are exposed to storms of high wind and wave energy which can leave parting contacts and associated sand layers such as those seen in the deposits at Linger
Longer Road and Fishermans Point. Other peat deposits overlain by clay horizons may be the result of compaction due either to the weight of overlying sediment or to seismic vibrations.

Short-term accelerations in the rate of eustatic sea-level rise might also be responsible for stratigraphic changes. Site-specific marsh responses to regional eustatic sea-level rise, including adjustments in tidal range, could account for some of the variations observed in the depositional record.

Significant amounts of vertical crustal displacement are probably not a factor in depositional changes at Quilcene Bay. The stratigraphy of deposits at Donovan Creek does not correlate with a nearly marsh-wide parting contact at Fishermans Point. If tectonic subsidence did occur at Fishermans Point, then a small amount of displacement is indicated by the rapid recovery of salt marsh vegetation.

**Padilla Bay**

Relative sea-level has risen 3 to 4 m at Padilla Bay in the past 4000-5000 years. Eroded tree stumps on the beach and the low elevations of the living forest suggest that an acceleration in the rate of relative sea-level rise may have occurred in the past several hundred years. This rise may be due to the compaction of organic layers by the weight of the forest and by human activities, as well as to facies...
changes resulting from significant reductions in sediment input since non-Indian settlement.

This study did not find conclusive stratigraphic evidence for sudden tectonic subsidence at Padilla Bay. Buried organic layers in marsh deposits in both the northern and southern marshes may represent subsidence events, but shifting shoals and bars offshore or changes in rates of sedimentation might also be responsible. Further study that includes pollen and microfossil analysis might identify the organic deposits as down-dropped freshwater horizons. However, the elevations of peat fossils dated as about 3800-4900 years old relative to those of the other study sites precludes a significant amount of differential vertical displacement between sites. Radiocarbon ages constrain differential tectonic displacement to a maximum of about 1 m.

Radiocarbon Ages

Introduction

The graph in Figure 45 shows the present N.G.V.D. elevations of dated samples and their radiocarbon ages after conversion to sidereal years. Assuming that late-Holocene marshes formed at intertidal ranges similar to those found today, the elevations shown represent differences in relative sea-level. In this section, a discussion of levels
Table 3. Indicator fossils, average elevation and radiocarbon ages for samples. Radiocarbon ages are calibrated to sidereal years to account for past variation in atmospheric radiocarbon (Stuiver and Reimer, 1986).

<table>
<thead>
<tr>
<th>SAMPLE</th>
<th>SITE</th>
<th>FOSSIL</th>
<th>AVERAGE ELEVATION</th>
<th>AGE IN $^{14}$C YRS BEFORE A.D. 1950</th>
<th>AGE IN SIDEREAL YRS BEFORE A.D. 1950 (2σ RANGE)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SJT-1</td>
<td>Third Lagoon</td>
<td>Distichlis</td>
<td>57 cm</td>
<td>900 ± 80</td>
<td>817 (970-670)</td>
</tr>
<tr>
<td>SJW-2</td>
<td>Westcott Bay</td>
<td>Distichlis</td>
<td>70 cm</td>
<td>520 ± 70</td>
<td>535 (670-470)</td>
</tr>
<tr>
<td>QDC-3</td>
<td>Donovan Creek</td>
<td>Distichlis</td>
<td>-117 cm</td>
<td>3170 ± 90</td>
<td>3396 (3636-3145)</td>
</tr>
<tr>
<td>QDC-4</td>
<td>Donovan Creek</td>
<td>Triglochin</td>
<td>81 cm</td>
<td>930 ± 60</td>
<td>910 (960-720)</td>
</tr>
<tr>
<td>QFP-5</td>
<td>Fishermans Point</td>
<td>Triglochin</td>
<td>-75 cm</td>
<td>2950 ± 70</td>
<td>3161 (3359-2886)</td>
</tr>
<tr>
<td>QLL-6</td>
<td>Linger Longer Rd.</td>
<td>Distichlis</td>
<td>61 cm</td>
<td>630 ± 60</td>
<td>646 (687-530)</td>
</tr>
<tr>
<td>PB-7</td>
<td>Padilla Bay</td>
<td>Distichlis</td>
<td>-320 cm</td>
<td>4170 ± 70</td>
<td>4821 (4869-4458)</td>
</tr>
<tr>
<td>PB-8</td>
<td>Padilla Bay</td>
<td>Triglochin</td>
<td>-161 cm</td>
<td>3750 ± 80</td>
<td>4148 (4419-3873)</td>
</tr>
</tbody>
</table>

Figure 45. Sample average elevations and radiocarbon ages after conversion to sidereal years.
of accuracy estimated for the radiocarbon ages and for the elevations shown in Figure 45 is followed by an interpretation of the rates of relative sea-level rise suggested by these data.

**Dating accuracy**

Table 3 lists the fossil indicator species, average elevations and radiocarbon dates for the eight basal peat samples. Variations in radiocarbon ages may occur if the plants dated are C-4 plants, that is, they are enriched in carbon-13 relative to carbon-12. Rhizomes that function as C-4 plants have resulted in radiocarbon ages that are inaccurately young by 150 to 200 years (Stuckenrath, 1977). *Distichlis spicata* is among the rhizomes found to follow the enriched C-4 pathway in photosynthesis (Bender, 1971), so that this error is possible for all samples except QDC-4, which did not include any *D. spicata*.

In Figure 45 the ranges in age are shown at two standard deviations, which are asymmetrical because of conversion to sidereal years. The point shown on the graph for age is the radiocarbon age in sidereal years. In a comparison of about 200 radiocarbon ages of samples from marshes on the Oregon and Washington Coasts, Grant and others (1989) found that the two sigma range did not account for all of the uncertainty associated with these dates. Therefore, a range of at least two standard deviations is recommended in an analysis of radiocarbon dates of salt
marsh deposits (Sutherland, 1983).

**Elevation accuracy**

Elevations in Figure 45 are plotted at the midpoints of the vertical intervals of the sample, listed as average elevations in Table 3.

Vertical uncertainty in relating dated samples to former sea-levels can result from:

1) compaction of the sediment lying between the peat sample and a non-compressible substrate,
2) the range in elevation of the indicator species in the sample, and
3) changes in tidal range, altering marsh vegetation zones and their relationships to Mean Sea Level.

Table 4 lists the maximum vertical error possible for each sample due to compaction by overlying sediment. Estimates are based on the thickness of the overlying sediment and are derived from empirical soil engineering studies as reported by Weller (1959) and revised by Bloom (1964). These are maximum values because they are determined on the basis of clay buried by clay, and these marshes are largely overlain by peat, which is lighter than clay. The amounts listed for samples from Donovan Creek and Linger Longer Road are slightly exaggerated because sandy clay compacts less than does clay without a sand component. In Figure 45 the error bars for compaction extend upward
Table 4. Maximum error for each sample due to compaction.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Sediment under sample</th>
<th>Sediment thickness over sample</th>
<th>% Compaction</th>
<th>Maximum error</th>
</tr>
</thead>
<tbody>
<tr>
<td>SJT-1</td>
<td>10 cm. clay</td>
<td>40 cm.</td>
<td>20</td>
<td>2 cm.</td>
</tr>
<tr>
<td>SJW-2</td>
<td>20 cm. clay</td>
<td>40 cm.</td>
<td>20</td>
<td>2 cm.</td>
</tr>
<tr>
<td>QDC-3</td>
<td>10 cm. sandy clay</td>
<td>325 cm.</td>
<td>45</td>
<td>4.5 cm.</td>
</tr>
<tr>
<td>QDC-4</td>
<td>40 cm. sandy clay</td>
<td>120 cm.</td>
<td>36</td>
<td>14.5 cm.</td>
</tr>
<tr>
<td>QFP-5</td>
<td>40 cm. clay</td>
<td>300 cm.</td>
<td>43</td>
<td>17.2 cm.</td>
</tr>
<tr>
<td>QLL-6</td>
<td>35 cm. sandy clay</td>
<td>115 cm.</td>
<td>32</td>
<td>11.2 cm.</td>
</tr>
<tr>
<td>PB-7</td>
<td>35 cm. clay</td>
<td>345 cm.</td>
<td>48</td>
<td>16.8 cm.</td>
</tr>
<tr>
<td>PB-8</td>
<td>85 cm. clay</td>
<td>185 cm.</td>
<td>39</td>
<td>33.2 cm.</td>
</tr>
</tbody>
</table>
The elevation range for the indicator species represents the largest vertical variation possible in this study. The largest vertical range determined for *D. spicata* is 0.55 m, and this amount is shown in both directions for samples SJT-1, SJW-2, QDC-3 and PB-7 where *D. spicata* is the only plant fossil used to indicate the relationship of former sea-level. The maximum vertical range in which both *D. spicata* and *Triglochin maritimum* occur is also 0.55 m, and that is the amount shown in both directions for samples QFP-5, QLL-6 and PB-8 in which both species are present.

Sample QDC-4 included only *T. maritimum*, and, because this plant can occur as a pioneer on the low tidal flat as well as in the high marsh, the vertical species range for this sample extends 1.49 m in both directions. The difference in elevation between the data point shown and other samples less than 1000 years old may reflect the higher tidal range of the marsh at Donovan Creek. Because the peat samples are interpreted to be from the upper part of the low marsh or from the relatively narrow range of the high marsh, their elevations represent relationships to former sea-levels that are close to each other. They can therefore be compared with each other to determine relative sea-level changes.

Fletcher (1988) discussed the accuracy in comparing
present-day tidal ranges with paleotidal ranges from times of lower eustatic sea-levels. Paleotidal ranges have been found to increase with rising sea-level. Although the deeper water of rising eustatic sea-level initially reduces the damping effects of friction, rising shorelines that move inland are influenced by shallower depths and often by narrower basin widths, so that tidal range is increased.

As sea-level rises, the effect of resonance is also increased. Stride (1982) pointed out that although friction across the continental shelf is a factor, changes in tidal range are dominated by resonance of the standing wave of the tide across the continental shelf. Because most present-day continental shelf widths are less than 1/4 the wavelength of the tide, tidal amplitudes increase as distance from the shelf edge increases and approaches the 1/4 tidal wavelength. This distance results in greater resonance and a larger tidal range.

Fletcher (1988) concluded that this increased tidal range has two opposing effects on an analysis of paleotidal variations in samples. Although the change in tidal ranges from the present is greater for older samples, the narrower paleotidal ranges of older samples had narrower marsh vegetation zones, so that the difference in paleotidal range is balanced to some degree by a smaller vertical range for indicator plants. Fletcher cited a study by Decker that quantified this uncertainty in 7000 to 1000 year-old
deposits on Delaware Bay. According to Fletcher, Decker found that a 12% error in ability to relate samples to former sea-levels by vegetation zones was nearly balanced by a 14% error in tidal range variation.

The variation in paleotidal range for samples from the Puget Basin marshes in this study is probably less than those studied by Decker. Many of the dated samples from Delaware Bay were collected from vibracores in offshore areas now removed from the intertidal zone. The marshes in this study are still in the intertidal zone, indicating paleotidal ranges relatively similar to modern tidal amplitudes. Because the eustatic sea-level rise during the past 5000 years has been less than that from 7000-1000 years ago, changes in paleotidal range for this study may vary less than those determined at Delaware Bay.

In terms of this study, this understanding of paleotidal ranges implies that present-day tidal ranges and ranges of indicator plants are probably at a maximum for the late Holocene. Therefore the estimated ranges of indicator plants shown in Figure 45 are probably maximum values.

Fletcher’s and Decker’s work also implies that samples from San Juan Island’s narrow tidal range probably grew in tidal ranges more similar to those at the site today than those of similar ages from Quilcene Bay, where tidal ranges are larger. On San Juan Island the incremental increases in tidal amplitude have been smaller and marsh vegetation zones
have been narrower, so that the relationship of samples to former sea-levels is more accurate.

Samples from Quilcene Bay dated at 4000-3000 BP may have experienced greater changes in paleotidal range than more recent samples have. However, because the tidal range was narrower at that time, the greater accuracy to the past in the range of indicator species may balance variations in tidal range when compared to samples dated at less than 1000 years old. It should also be noted that erosion and associated changes in sedimentation patterns that accompany relative sea-level rise may have caused variations in tidal range as well, especially during periods of rapid regional eustatic sea-level rise.

Rates of relative sea-level rise

Calculations of average rates of relative sea-level rise (Table 5) are based on the average elevations of the samples and their radiocarbon ages after conversion to sidereal years. Average rates assume a constant rate of change, and do not compensate for changes in tidal range. A calculation of average rates between samples is based on the assumption that they are from similar levels in the tidal range. In calculations for time intervals of less than 1000 years, elevation differences of 0.5 m may alter the average rate by one order of magnitude.

For samples with ages that overlap at two standard
Table 5. Average rates of relative sea-level rise between two samples or between a sample and the elevation of the present surface. Ranges of average rates are determined using average elevations and 2-sigma ranges of calibrated radiocarbon ages (sidereal years before AD 1950).

<table>
<thead>
<tr>
<th>Site</th>
<th>Determined for samples</th>
<th>Range of calibrated $^{14}$C ages</th>
<th>Range of average rates</th>
</tr>
</thead>
<tbody>
<tr>
<td>Padilla Bay</td>
<td>PB-7 to PB-8</td>
<td>(4869-3873) - (4458-4419)</td>
<td>1.6 - 41 mm/yr</td>
</tr>
<tr>
<td></td>
<td>PB-7 to present</td>
<td>(4869-AD1950) - (4458-AD1950)</td>
<td>0.7 - 0.8 mm/yr</td>
</tr>
<tr>
<td></td>
<td>PB-8 to present</td>
<td>(4419-AD1950) - (3873-AD1950)</td>
<td>0.4 - 0.5 mm/yr</td>
</tr>
<tr>
<td>Quilcene Bay</td>
<td>QDC-3 to QFP-5</td>
<td>(3636-2886) - overlap at 2-sigma</td>
<td>0.6 - 420 mm/yr</td>
</tr>
<tr>
<td></td>
<td>QDC-3 to QDC-4</td>
<td>(3636-720) - (3145-960)</td>
<td>0.7 - 0.9 mm/yr</td>
</tr>
<tr>
<td></td>
<td>QFP-5 to QDC-4</td>
<td>(3359-720) - (2886-960)</td>
<td>0.6 - 0.8 mm/yr</td>
</tr>
<tr>
<td></td>
<td>QDC-3 to QLL-6</td>
<td>(3636-530) - (3145-687)</td>
<td>0.6 - 0.7 mm/yr</td>
</tr>
<tr>
<td></td>
<td>QFP-5 to QLL-6</td>
<td>(3359-530) - (2886-687)</td>
<td>0.5 - 0.6 mm/yr</td>
</tr>
<tr>
<td></td>
<td>QDC-3 to present</td>
<td>(3636-AD1950) - (3145-AD1950)</td>
<td>0.9 - 1.0 mm/yr</td>
</tr>
<tr>
<td></td>
<td>QFP-5 to present</td>
<td>(3359-AD1950) - (2886-AD1950)</td>
<td>0.9 - 1.0 mm/yr</td>
</tr>
<tr>
<td></td>
<td>QDC-4 to present</td>
<td>(960-AD1950) - (720-AD1950)</td>
<td>1.1 - 1.5 mm/yr</td>
</tr>
<tr>
<td></td>
<td>QLL-6 to present</td>
<td>(687-AD1950) - (530-AD1950)</td>
<td>1.8 - 2.4 mm/yr</td>
</tr>
<tr>
<td>San Juan</td>
<td>SJT-1 to SJW-2</td>
<td>(970-470) - overlap at 2-sigma</td>
<td>0.3 - 130 mm/yr</td>
</tr>
<tr>
<td>Island</td>
<td>SJT-1 to present</td>
<td>(970-AD1950) - (670-AD1950)</td>
<td>0.4 - 0.6 mm/yr</td>
</tr>
<tr>
<td></td>
<td>SJW-2 to present</td>
<td>(670-AD1950) - (470-AD1950)</td>
<td>0.6 - 0.9 mm/yr</td>
</tr>
</tbody>
</table>
deviations, the maximum average rate between them is unrealistically large. Recalculated average rates using the midpoint radiocarbon ages of these samples after conversion to sidereal years are:

PB-7 to PB-8 2.4 mm/yr
QDC-3 to QFP-5 1.8 mm/yr
SJT-1 to SJW-2 0.5 mm/yr

At Padilla Bay, the range of average rates of sea-level rise decreases from a minimum rate of 1.6 mm/yr before about 4000 years ago to a maximum rate of 0.8 mm/yr from 4000 BP to AD 1950.

The range of average rates from PB-7 and PB-8 to AD 1950 (0.4-0.8 mm/yr) may reflect the artificially low elevations of an altered marsh. This problem may apply in general to marshes with altered tidal circulation or artificial levees, because the elevation of the present marsh surface does not represent an accurate relationship to current sea-level.

The local tidal range, the high tide measurement and the ranges of indicator plant species observed at Padilla Bay suggest that the natural elevation for a low marsh surface at this site would fall in-between those of San Juan Island and Quilcene Bay. A range of average rates of relative sea-level rise that are recalculated for a 1-2 m greater elevation of the present surface at ±2 sigma may more accurately represent the long-term average rate:
PB-7 to AD 1950  0.9-1.2 mm/yr
PB-8 to AD 1950  0.6-1.0 mm/yr

Assuming that the average rates in Table 5 are reliable, then the average rate of relative sea-level rise at Quilcene Bay has increased in the past 500-1000 years compared to rates for the past 3000 years. The range of average rates from about 3000 years ago to 500-1000 years ago (QDC-3 and QFP-5 to QDC-4 and QLL-6) is 0.5 to 0.9 mm/yr. The average rate is 0.9 to 1.0 mm/yr from both QDC-3 and QFP-5 to AD 1950, a correlation between the Donovan Creek and the Fishermans Point marshes for the period from about 3000 BP to AD 1950. The higher range of average rates for the interval that includes the past 500-1000 years suggests that sea-level rose more rapidly in that time period than it did from 3000 to about 1000 years ago.

Average rates calculated for samples from San Juan Island also appear to suggest a more rapid rate for the more recent past. The range of average rates for the past 900 years is 0.4 to 0.6 mm/yr at Third Lagoon. An average rate of relative sea-level rise for Westcott Bay is 0.6 to 0.9 mm/yr for about 600 years ago to AD 1950. The average rates may be slightly less for Third Lagoon than for Westcott Bay because of erosion of the present surface at Third Lagoon. A small vertical difference here results in a proportionally large change in rate because of the shallow (38-40 cm) depth of the samples.
For San Juan Island, the results suggest that the average rate of relative sea-level rise in the last 800 years has not exceeded 0.9 mm/yr. This rate is lower than those calculated for the same period at Quilcene Bay, and may reflect marsh responses to smaller incremental increases in tidal range in the narrower tidal range of the San Juan Islands. The average rates determined for all samples for the past 1000 years do not differ significantly within the range of vertical error.

Discussion

Introduction

In this section, regional relative sea-level data for the northern Puget Basin and southwestern British Columbia are compared, followed by a discussion of the implications of the findings of this study for the hypothesized accelerated sea-level rise due to global warming. Finally, the tectonic implications of these findings are discussed in relationship to regional relative sea-level history.

Eustatic implications

Late Holocene relative sea-level changes - The absence of evidence of significant vertical crustal movement at the study sites suggests that the relative sea-level rise observed in this study may be dominantly eustatic rather
Figure 46. A comparison of the relative sea-level data from this study with that from Shine and southwestern British Columbia for the late-Holocene.
than tectonic in origin. A comparison of data from this study with results from Shine, Washington (Eronen and others, 1987) and southwestern British Columbia (Clague and others, 1982; Clague, 1989) is shown in Figure 46.

The samples from Shine were interpreted from microfossil analysis to be from higher high-tide levels, placing them at the same level or slightly higher in the intertidal zone than samples from this study. The core site was not surveyed; elevations are based on the authors' estimate. Compaction of sediment may have lowered sample elevations, and at least 1 m of vertical uncertainty can be assumed in relating samples to former sea-level. The core from which the samples came showed continuous upward growth of organic material for about 6000 years.

Clague's relative sea-level curve is based on a compilation of radiocarbon dates from many archaeological and geological studies, as well as from his own work on the relative sea-level history of the British Columbia coast. The curve represents the position of Mean Sea Level, a level lower in the intertidal zone than that of the samples from Shine and from this study. The envelope represents the uncertainty in relating dated samples to Mean Sea Level. Clague and Bobrosky (1990) suggest from these data that sea-level in southwestern British Columbia rose to within several meters of present sea-level by about 5000 BP and has not varied by more than 1 m since about 2000 BP.
The results of this study for the Puget Basin describe a relative sea-level rise of approximately 2-3 \( \pm \) 0.5 m between 5000 and 3000 years ago and a slower average rate of approximately 1 \( \pm \) 0.5 m between about 3000 and 500-1000 years ago. An increased average rate of relative sea-level rise is suggested at Quilcene Bay in the past 1000 years, the only site providing both older and younger sample ages.

The results of studies in southwestern British Columbia and at Shine support the hypothesis that the relative sea-level rise at the sites examined in this study has been primarily eustatic for the past 5000 years, with perhaps a small contribution from slow, steady regional subsidence of the fore-arc basin. The regional trend in relative sea-level rise is apparent in Figure 46, even when allowing for possible variations in tidal range.

**Historical relative sea-level changes** - A comparison of the average rates of relative sea-level rise for the past 1000 years with average historical rates of relative sea-level rise neither supports nor contradicts the hypothesis of an accelerated sea-level rise in this century. Hicks and Hickman (1988) determined the following average historical rates of sea-level rise derived from tide gauge analyses for Seattle (where the tidal range is comparable to that at Quilcene Bay) and Friday Harbor:
A comparison of an average long-term rate with a relatively short-term record is of limited value, but it is interesting to note that late Holocene ranges of average rates are generally slightly lower on San Juan Island, although historical rates for the island fall well within the late Holocene range of vertical error.

The average global eustatic sea-level rise (estimated at 10 cm for the past century) (Titus, 1988) may be responsible for geomorphic evidence of recent relative sea-level rise at San Juan Island. The strongest evidence produced by this study in support of the hypothesis of a recent acceleration in the rate of sea-level rise is the geomorphic evidence of changing marsh configurations on San Juan Island where they had apparently been unchanged for 500 to 1000 years previously. Further study is suggested to rule out salt-water intrusion as the agent of this change.

Geomorphic changes observed on San Juan Island may reflect the greater sensitivity of marsh vegetation zones to changes in relative sea-level in an area of narrow tidal ranges. However, further study of these and other marshes in the Puget Basin is required over longer time periods.
before these geomorphic observations can be identified as an early signal of accelerated sea-level rise.

If Fletcher’s (1988) conclusions regarding paleotidal range increases can be applied to the accelerated sea-level rise hypothesized by some scientists for the next century (Barth and Titus, 1984), then monitoring programs can anticipate two trends: 1) the earliest salt marsh vegetation changes can be expected in areas of narrow tidal range, such as San Juan Island or the Straits of Juan de Fuca, and 2) the greatest increases in tidal range can be anticipated where tidal amplitudes are presently largest, in southern and eastern Puget Sound.

Regional relative sea-level rise for this century is of too small a magnitude for an increased tidal amplitude to emerge from the range of natural variability in tide gauge analyses. There may be a critical level of sea-level increase that must be reached before the factors that induce the increased tidal range, such as distance from the continental shelf edge, begin to operate.

However, if tidal ranges increase during a relative sea-level rise, then the highest tides will become higher at a more rapid rate than sea-level is rising. The highest tides at any one location will be even higher than those otherwise predicted for elevated sea-levels. Where tidal amplitudes are presently larger, tidal ranges can be expected to show a proportionally greater increase. Since
the highest tides are responsible for the most erosion and flooding of the backshore, this line of reasoning suggests that coastal areas with larger tidal amplitudes in inner and southern Puget Sound are more vulnerable to accelerated sea-level rise than are areas of narrower tidal ranges along the outer coast and the Straits of Juan de Fuca.

Tectonic implications

The results of this study indicate that the relative sea-level changes observed are not related to sudden tectonic events involving significant amounts of vertical crustal movement. Correlation of these results with the relative sea-level curve of Clague (1989) and the samples from Shine (Eronen and others, 1987) in Figure 46 suggests that if the crust has moved vertically at the study sites since the sample dates, then total differential displacement has probably not exceeded 1 m.

The hypothesis of tectonically-induced differential tilting to the south and east across the region proposed by Holdahl and others (1989) for the historical period is not supported by the results of this study. Samples for Quilcene Bay dated at less than 1000 years old have higher average rates of relative sea-level rise for this period than do those from San Juan Island in the north, which might appear to support Holdahl’s hypothesis of a subsiding southern Puget Basin. The difference in rates is not
considered significant because all of these rates fall within the range of vertical uncertainty due to indicator plant species ranges and to regional variations in tidal range.

The lack of evidence for vertical crustal movement greater than 1 m seems to suggest a less active tectonic regime here than in the southern Puget Basin, where evidence has been found on Bainbridge Island suggesting 7 m of vertical displacement in the last 1700 years (Bucknam and Barnhard, 1989). Gravity data (Bonini and others, 1974) suggests that the east-west trending, deep basinal structures under the Puget Basin lie to the south of San Juan Island and Padilla Bay and to the east of Quilcene Bay. Although the southern Puget Basin has been more active tectonically during the historical period, further study in the northern Puget Basin may uncover sites with significant pre-historic vertical displacement.

Reconstruction of an accurate relative sea-level history for the northern Puget Basin requires correlation of the results of this and previous studies with further studies. It may be that the regional structure is so complex that no single regional relative sea-level curve can be determined for the late Holocene. Alternatively, late Holocene tectonic activity in the Puget Basin may be limited to certain areas of intense activity, so that differential movement in these areas shows a distinct departure from the
general trend of relative sea-level changes shown in Figure 46.

If the northern part of the region is tectonically stable for this period, then the results of this study, combined with data from eastern Vancouver Island and the Fraser Lowland, may serve as the basis of a regional relative sea-level curve that can be more clearly defined by further study. A regional relative sea-level curve for the late Holocene can be useful not only in determining areas of differential vertical crustal movement but also in discerning an early signal of sea-level rise induced by global warming.
CONCLUSIONS

The results of this study suggest that:

1) Salt marshes in the Puget Basin contain evidence of relative sea-level changes. Limitations to precise measurement of these changes are imposed by multiple variables that operate on a coast with a complicated and irregular configuration and differing tidal regimes.

2) Data from these marshes indicate a relative sea-level rise of approximately $2-3 \pm 0.5$ m between 5000 and 3000 years ago, and approximately $1 \pm 0.5$ m between 3000 and 1000 years ago. Regional relative sea-level has probably not risen more than about 1 m in the past 1000 years. These conclusions are supported by data from previous studies.

3) Differential vertical crustal displacements of more than 1 m have probably not occurred at Padilla Bay in the past 4500 years, at Quilcene Bay in the past 3000 years, and on San Juan Island in the past 800 years.

4) Differential crustal tilting across the region during the late Holocene is not supported by the results of this study within allowable error.

5) If the hypothesized acceleration of eustatic sea-level rise occurs as a result of global warming, then inner and southern Puget Sound shorelines that have larger tidal ranges may be more vulnerable to erosion and flooding due to a proportionally greater rate of increase in the elevations.
of the higher high tides.

6) Elevations of present marsh surfaces can be used to monitor future changes in sea-level. Changes in the elevation of marsh surfaces can be evaluated quantitatively with respect to changes in tidal range and other geomorphic factors.

A more complete data set for the Puget Basin is required to evaluate the results of this study. Future research on Puget Basin salt marshes is recommended:

1) to document the late Holocene history of regional relative sea-level change;

2) to locate areas that have experienced significant vertical crustal movement during the late Holocene so as to better understand regional geologic structures and identify areas of earthquake hazard;

3) to predict the response of variable tidal ranges to differing rates of relative sea-level rise so as to better identify areas within the Puget Basin most vulnerable to the impact of future sea-level rise;

4) to contribute further baseline data on salt marsh elevations and geomorphology for monitoring purposes in the event of an accelerated sea-level rise; and

5) to contribute to an inventory of Puget Basin tidal wetlands and to an improved understanding of the processes influencing them in order to establish conservation priorities.
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APPENDIX I

Salt Marshes Examined and Not Selected for Study
APPENDIX I

Salt Marshes Examined and Not Selected for Study

Twenty-two sites were examined and researched and found unsuitable for this study. These salt marshes are listed in Table 1.

Test cores made at four additional marshes indicate that the following sites are also unsuitable for this study:

1. Dosewallips delta, Jefferson County

Sedimentation from steep uplands that have been logged has resulted in a rapidly prograding marsh. Cores revealed beach cobbles and sand at less than 1 m. The deeper marsh deposits are now out of contact with tidal influence.

2. Kala Point, Jefferson County

Cores revealed substrates of sand and intertidal clay at depths of less than 0.5 m. Part of the berm is composed of the shells of an Indian midden that may have altered tidal circulation. A nearby mill at Irondale and the opening of the Port Townsend Canal in 1916 may also have affected the marsh configuration.

3. Garrison Bay, San Juan Island

Part of the small marsh has been altered by residential development. The remaining marsh fringes a moderately steep upland that has been logged at least once with possible
<table>
<thead>
<tr>
<th>No.</th>
<th>Marsh</th>
<th>County</th>
<th>Reason unsuitable for study</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Perego's Lagoon</td>
<td>Island</td>
<td>Berm breached in 1982; altered tidal circulation.</td>
</tr>
<tr>
<td>2</td>
<td>Harrington Lagoon</td>
<td>Island</td>
<td>Altered by residential development.</td>
</tr>
<tr>
<td>3</td>
<td>Discovery Bay</td>
<td>Jefferson</td>
<td>Affected by logging, agriculture, and residential development.</td>
</tr>
<tr>
<td>4</td>
<td>Lower Hadlock</td>
<td>Jefferson</td>
<td>Altered by mill (1870-1913) and port facilities for alcohol plant until 1930s.</td>
</tr>
<tr>
<td>5</td>
<td>Mystery Bay</td>
<td>Jefferson</td>
<td>Altered by Scow Bay closure and residential development.</td>
</tr>
<tr>
<td>7</td>
<td>Oak Bay</td>
<td>Jefferson</td>
<td>Altered by Port Townsend Canal.</td>
</tr>
<tr>
<td>8</td>
<td>Scow Bay</td>
<td>Jefferson</td>
<td>Altered by road construction.</td>
</tr>
<tr>
<td>9</td>
<td>Fort Flagler</td>
<td>Jefferson</td>
<td>Filled and altered for park.</td>
</tr>
<tr>
<td>10</td>
<td>Point Ludlow</td>
<td>Jefferson</td>
<td>Altered by residential development.</td>
</tr>
<tr>
<td>11</td>
<td>Mats Mats Bay</td>
<td>Jefferson</td>
<td>Altered by residential development.</td>
</tr>
<tr>
<td>12</td>
<td>South Point</td>
<td>Jefferson</td>
<td>Altered by residential development.</td>
</tr>
<tr>
<td>13</td>
<td>Right Smart Cove</td>
<td>Jefferson</td>
<td>Altered by residential development.</td>
</tr>
<tr>
<td>14</td>
<td>Whitney Point</td>
<td>Jefferson</td>
<td>Altered to construct state shellfish laboratory.</td>
</tr>
<tr>
<td>15</td>
<td>Fisherman’s Cove</td>
<td>Jefferson</td>
<td>Used for log storage; little of marsh remains.</td>
</tr>
<tr>
<td>16</td>
<td>Squamish Bay</td>
<td>Jefferson</td>
<td>Altered by residential development; poor tidal circulation.</td>
</tr>
<tr>
<td>17</td>
<td>Thorndyke Bay</td>
<td>Jefferson</td>
<td>Used for log storage; non-native biota introduced.</td>
</tr>
<tr>
<td>18</td>
<td>Foulweather Bluff</td>
<td>Kitsap</td>
<td>Tidal circulation poor; marsh is brackish.</td>
</tr>
<tr>
<td>19</td>
<td>Mitchell Bay</td>
<td>San Juan</td>
<td>Marsh is small; altered by road construction.</td>
</tr>
<tr>
<td>20</td>
<td>Davison Head</td>
<td>San Juan</td>
<td>Altered by road construction and residential development.</td>
</tr>
<tr>
<td>21</td>
<td>White Point Isthmus</td>
<td>San Juan</td>
<td>Altered by road construction and artificial berms.</td>
</tr>
<tr>
<td>22</td>
<td>Jakle's Lagoon</td>
<td>San Juan</td>
<td>Marsh is small; tidal circulation has been altered.</td>
</tr>
</tbody>
</table>
effects on the marsh. A sewage outfall pipe drains into the bay via the marsh. Cores indicate a thickness of less than 0.5 m.

4. Telegraph Slough, Padilla Bay

Cores revealed a shallow marsh of less than 0.5 m over tidal flat mud. The marsh has probably developed seaward of levees since nearby marshes were reclaimed and the Swinomish Channel was dredged.

A more detailed examination of the above marshes and those listed in Table 1 might reveal parts of some marshes with stratigraphic records adequate for a study of relative sea-level changes.

Marshes examined and suggested for possible future study include:

1. Race Lagoon, Whidbey Island

   Only part of the marsh has been altered by residential development. This marsh might provide data in a part of the Puget Basin where few marshes are preserved.

2. Tarboo Bay, Jefferson County

   Marshes behind a series of spits have not been altered by human activity, but Kunze (1984) reported evidence that the spits have shifted northward due to the long fetch on Hood Canal and Dabob Bay. The marsh behind Long Spit is likely to have the longest stratigraphic record.

3. Bywater Bay, Jefferson County

   The marsh here is small and exposed to a long fetch
along Hood Canal and Admiralty Inlet. Less exposed parts of the marsh may have an age suitable for study.
APPENDIX II

Core Data
APPENDIX II

Core Data

Representative cores are described for each cross-section and for each core from which samples were taken for radiocarbon dating.

Explanation of abbreviations:

Core #:
- TL - Third Lagoon
- WB - Westcott Bay
- FP - Fishermans Point
- DC - Donovan Creek
- LL - Linger Longer
- PB - Padilla Bay
- # - number of core on cross-section

Depth (cm): depth from surface in centimeters

Description: sediment (color, texture, structure, type), biota identified, debris, nature of contact below.

- color: brn - brown, ox - oxidized, dk - dark, blk - black, med - medium, bl - blue
- texture: cmpt - compact, fri - friable
- structure: biot - bioturbated
- biota: tri - Triglochin maritimum, dist - Distichlis spicata, pw - Salicornia virginica, unveg - unvegetated, spveg - sparsely veg, veg/gp - vegetation in growth position, horiz - lying horizontally
- contact: trans - transitional, hard - unable to core deeper, uncert - uncertain nature, base - base level, parting - parting contact
<table>
<thead>
<tr>
<th>Core #</th>
<th>Depth (cm)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>TL-2</td>
<td>1-16</td>
<td>brn silty peat, dist &amp; pw at surface, parting</td>
</tr>
<tr>
<td></td>
<td>16-43</td>
<td>ox peat, veg/gp, trans</td>
</tr>
<tr>
<td></td>
<td>43-48</td>
<td>gray clay, veg/gp, trans</td>
</tr>
<tr>
<td></td>
<td>48-54</td>
<td>ox cmpt peat, trans</td>
</tr>
<tr>
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<td>54-56</td>
<td>biot unveg gray clay, gravel at base</td>
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<td>brn silty peat, dist &amp; pw at surface, parting</td>
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<td>15-42</td>
<td>ox peat, veg/gp, dist SAMPLE (38-40 cm), trans</td>
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<td>42-49</td>
<td>gray clay, veg/gp, trans</td>
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<td>gray clay, spveg/gp, gravel at base</td>
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<td>WB-1</td>
<td>1-20</td>
<td>brn cmpt peat, brackish at surface, trans</td>
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<td>brn silty peat, hard soil at base</td>
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<td>brn peat, pw at surface, dist SAMPLE (38-40 cm), trans</td>
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<td>gray clay, spveg/gp, trans</td>
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<td>gray clay, unveg, grades to brn clay, sand and gravel at base</td>
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<td>WB-6</td>
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<td>36-51</td>
<td>gray clay, spveg/gp, trans</td>
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<td>gray clay, unveg, hard</td>
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<td>brn clay, veg/gp, trans</td>
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<td>36-124</td>
<td>ox peat, dist &amp; tri, wood, trans</td>
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<td>124-125</td>
<td>brn clay, veg/gp, parting</td>
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<td>125-190</td>
<td>brn silty peat, veg/gp &amp; hor, dist &amp; tri, trans</td>
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<td>gray clay, veg/gp, dist &amp; tri, 2 cm lens of ox peat at 245 cm, charcoal &amp; wood, trans</td>
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<td>300-355</td>
<td>gray clay, veg/gp, bl-blk gravel at base</td>
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<td>1-20</td>
<td>brn peat, high marsh veg at surface, trans</td>
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<td>20-21</td>
<td>brn clay, veg/gp, trans</td>
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<td>21-120</td>
<td>ox peat w/ occas silty laminae, dist &amp; tri, some hor veg, trans</td>
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<td>122-220</td>
<td>brn peat w/occas silty laminae, veg/gp, trans</td>
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<td>gray clay, spveg, parting</td>
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<td>235-290</td>
<td>ox peat, siltier below 255 cm, dist &amp; tri in gp, wood &amp; twigs, trans</td>
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<td>brn silty peat, dist &amp; tri SAMPLE, trans</td>
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<td>dk brn cmpt peat, dist &amp; tri, charcoal &amp; wood, trans</td>
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<td>sand &amp; gravel w/minor clay, parting</td>
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<td>100-140</td>
<td>brn to gray clay, veg/gp to unveg, charcoal and twigs, trans</td>
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<td>140-150</td>
<td>brn silty peat, charcoal &amp; wood, trans</td>
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<td>brn cmpt peat w/ silty laminae, dist &amp; tri, wood &amp; charcoal, trans</td>
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<td>gray clay, veg/gp to spveg, wood &amp; charcoal, bl-blk gravel at base</td>
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<td>1-4</td>
<td>dk brn peat, high marsh at surface, trans</td>
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<td>gray clay, veg/gp, trans</td>
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<td>8-109</td>
<td>ox peat w/ gray clay lamina at 60 cm, mod cmpt at base, dist &amp; tri, twigs, seeds, wood, trans</td>
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<td>111-130</td>
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<td>130-133</td>
<td>gray-brn clay, veg/gp, trans</td>
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<td>133-160</td>
<td>ox peat w/ sm laminae of clay/sand, dist &amp; tri, veg/gp, horiz grass, many twigs, trans</td>
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<td>160-240</td>
<td>brn to gray clay, veg/gp, wood, trans</td>
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<td>brn peat, dist &amp; tri, wood, trans</td>
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<td>gray clay, veg/gp, trans</td>
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<td>7-56</td>
<td>dk brn to ox peat, dist, twigs &amp; wood, trans</td>
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<td>56-58</td>
<td>brn clay, veg/gp, trans</td>
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<td>58-112</td>
<td>ox peat, dist &amp; tri, trans</td>
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<td>112-115</td>
<td>gray clay, spveg, charcoal, parting</td>
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<td>115-144</td>
<td>brn cmpt peat, veg/gp, charcoal, trans</td>
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<td>144-150</td>
<td>brn cmpt clay, veg/gp, charcoal, wood &amp; twigs, trans</td>
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<td>gray-brn cmpt clay, spveg/gp, hor grass, twigs, hard</td>
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<td>ox cmpt peat, high marsh at surface, tri, veg/gp, trans</td>
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<td>brn-gray silty peat, veg/gp, tri SAMPLE (115-120 cm), trans</td>
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<td>sand w/ dry silt near surface, gravel at base</td>
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<td>40-42</td>
<td>sand, trans</td>
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<td>ox cmpt clay, spveg/gp, trans</td>
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<td>sand, parting</td>
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<td>67-160</td>
<td>gray clay, spveg/gp to unveg, trans</td>
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<td>160-162</td>
<td>sand, parting</td>
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<td>162-310</td>
<td>gray clay, spveg/gp to unveg, trans</td>
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<td>310-330</td>
<td>dk brn cmpt peat, tri, veg/gp, trans</td>
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<td>gray clay, spveg to unveg, hard</td>
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<td>brn cmpt peat, high marsh at surface, trans</td>
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<td>35-37</td>
<td>brn clay, veg/gp, trans</td>
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<td>med brn peat, dist, horiz grasses, trans</td>
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<td>med brn peat, dist &amp; tri, veg/gp, trans</td>
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<td>bl-gray clay, spveg to unveg, charcoal, trans</td>
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<td>sandy silty peat, dist &amp; tri SAMPLE (325-330 cm), wood, trans</td>
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<td>dk brn cmpt peat w/faint peat lenses, spveg/gp, trans</td>
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<td>Depth (cm)</td>
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<td>140-150</td>
<td>bl-gray clay, veg/gp, trans</td>
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<td>brn to ox peat, high marsh at surface, parting at 23 cm, dist &amp; tri, charcoal, trans</td>
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<td>brn silty peat, dist &amp; tri SAMPLE (110-115 cm), trans</td>
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<td>dk gray clay, trans</td>
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<td>sandy clay, spveg/gp, twigs &amp; stems, trans</td>
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<td>ox cmpt peat, dist &amp; tri, parting</td>
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<td>dark gray sandy clay, spveg/gp, gravel at base</td>
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<td>1-60</td>
<td>brn peat, pw and dist at surface, trans</td>
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<td>60-70</td>
<td>brn fri organic, roots, trans</td>
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<td>brn fri organic, roots, parting</td>
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<td>brn silty peat, veg/gp, uncert</td>
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<td>sand &amp; gravel, uncert</td>
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<td>brn silty peat, dist &amp; tri SAMPLE (180-185 cm), trans</td>
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<td>gray clay, veg/gp to unveg, gravel at base</td>
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<td>dk brn fri organic, parting</td>
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<td>lenses of peat, clay and silty peat, veg/gp, dist SAMPLE (345-350 cm), trans</td>
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<td>ox peat, eroding peat at surface, trans</td>
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<td>pale gray ash, veg/gp, horiz grasses &amp; stems, wood, parting</td>
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<td>pale gray ash, spveg/gp, horiz fine grasses &amp; stems, trans</td>
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<td>bl-gray clay to dk bl micaceous sandy clay, spveg/gp to unveg, limit of corer</td>
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<td>bl-gray sandy clay, gravel at base</td>
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<td>brn peat high marsh at surface, dist, trans</td>
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<td>pale gray ash, spveg/gp, horiz grass</td>
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<td>&amp; stems, parting gray clay, veg/gp, trans</td>
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<td>gray micaceous fine sand, end of corer</td>
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