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Historical evolution of the Columbia River littoral cell

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ABSTRACT

This paper details the historical coastal evolution of the Columbia River littoral cell in the Pacific Northwest of the United States. Geological data from A.D. 1700 and records leading up to the late 1800s provide insights to the natural system dynamics prior to significant human intervention, most notably jetty construction between 1885 and 1917. All reliable surveys, charts, and aerial photos are used to quantify decadal-scale changes at the three estuary entrances and four sub-cells of the littoral cell. Shoreline, bathymetric, and topographic change over three historical intervals—1870s–1920s, 1920s–1950s, and 1950s–1990s—are integrated to provide an understanding of sediment-sharing relationships among the littoral cell components. Regional morphological change data are developed for alongshore segments of approximately 5 km, enabling comparisons of shoreline change to upper-shoreface and barrier volume change within common compartments. The construction of entrance jetties at the Columbia River (1885–1917) and Grays Harbor (1898–1916) has profoundly affected the evolution of the littoral cell, and has accentuated the morphological coupling between the inlets, ebb-tidal deltas, shorefaces, and barriers. The jetties induced erosion of the inlets and offshore migration of ebb-tidal deltas. The change in boundary conditions at the entrances enabled waves to rework the flanks of ebb-tidal deltas and supply enormous quantities of sand to the adjacent coasts. Over several decades the initial sand pulses have been dispersed alongshore up to tens of kilometers from the estuary entrances. Winter waves and coastal currents produce net northward sediment transport across the shoreface while summer conditions tend to induce onshore sediment transport and accumulation of the upper shoreface and barriers at relatively high rates. Historical shoreline progradation rates since jetty construction are approximately double the late prehistoric rates between 1700 and the 1870s. Erosion rates of the mid- to lower shoreface to the south of the jettied estuary entrances have typically been greater than the accumulation rates of the upper shoreface and barrier, suggesting that the lower shoreface has been an important source of littoral sediments over decadal and longer time scales. Until recent decades, sediment supply from the ebb-tidal delta flanks and lower shoreface has largely masked the decline in Columbia River sediment supply resulting from flow regulation and dredging disposal practices. With the contemporary onset and expansion of coastal erosion adjacent to the jettied estuary entrances, proper management of dredged sediment is imperative to mitigate the effects of a declining sediment budget.

1. Introduction

1.1. Study motivation

Shoreline mapping is a basic societal need because it supports many functions including legal boundaries, land-use planning and regulation, property insurance, navigational charts, and vulnerability assessments. As essential as the shoreline is to a well-functioning society, shoreline delineation and change analysis can be relatively complex endeavors. Even when a precise definition of 'shoreline' is used, it is nevertheless a sometimes amorphous feature that can make interpretation difficult. The shoreline fluctuates over all time scales in response to changes in: relative sea level; littoral sediment transport gradients; cross-shore gradients over the shoreface; and the balance of the sediment budget from sources (e.g., fluvial, estuarine, marine) and sinks (e.g., backbarrier, submarine canyons). This natural variability, when combined with mapping error and inconsistent interpretations, present many challenges in accurately assessing coastal change (Crowell et al., 1991; Moore, 2000; Ruggiero et al., 2003a).

Despite the importance of shoreline delineation, the shoreline is often an inadequate representation of the coast, which has three-dimensional sub-aerial and sub-aqueous morphology of varying

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composition. In general, the time scale of morphological change increases with distance offshore from the shoreline (Niedoroda and Swift, 1991; Stive and de Vriend, 1995; Nicholls et al., 1998; Cowell et al., 1999). However, changes in the offshore portion of the planform (i.e., the lower shoreface and continental shelf) have a disproportionately large influence on the upper shoreface, due to cross-shore length scales and mass continuity for sediment exchanges between the two zones (Roy et al., 1994; Cowell et al., 1999). The morphological coupling of the offshore and nearshore zones implies the need for a systems framework which integrates this interaction in order to predict large-scale coastal behaviour (decades or longer).

Cowell et al. (2003a) propose using a composite morphology within a systems framework, collectively referred to as the 'coastal tract,' as a way to assimilate the hierarchical nature of coastal systems. The coastal tract is the first-order system (i.e., system of interest) that accounts for morphological coupling and internal dynamics that extend from the lower shoreface across the upper shoreface and barrier to the backbarrier. The lower-order system (i.e., the larger environment) sets boundary conditions for the first-order system, while higher-order components transfer residual effects to the first-order system. In practical terms, the coastal tract specifies coherent morphodynamic systems and provides a way to integrate coastal-change models across multiple temporal and spatial scales.

Using this coastal tract framework, this study focuses on the Columbia River littoral cell (CRLC; Fig. 1) as the sediment-sharing system of interest. The study examines historical evolution spanning decades to centuries and an active zone of morphologic change that extends landward from approximately 40-m water depth across the lower shoreface and barrier to the estuary. This zone encompasses the morphological system composed of Columbia River sand (less than 10% silt); seaward of this zone mud deposits tend to increase significantly (Nittrouer and Sternberg, 1981; Twichell et al., 2010-this issue). Gross et al. (1969) infer a predominant onshore component of sand transport in less than 40-m water depth along southwest Washington. Smith and Hopkins (1972) and Harlett and Kulm (1973) hypothesize that coarse sediments are essentially trapped in the nearshore as a result of strong wave-dominated onshore transport along the bottom, while fine sediments are winnowed and transported offshore as suspended load toward the mid-shelf. This hypothesis is consistent with observations and modeling of the northern California continental shelf, which suggest that cross-shore gradation in sediment size may result from net erosion and offshore transport of coarse silt and fine sand in water depths shallower than 50 m (Harris and Wiberg, 2002).

This study combines historical shoreline change analysis with a sediment-budget approach to quantify the historical evolution of the CRLC over three intervals (1870s–1920s, 1920s–1950s, and 1950s–1999). In addition, shoreline changes prior to significant human influence (1700–1870s) are also calculated. Data from these intervals are supplemented with additional shoreline data at sub-decadal scale, where needed and available. The additional shoreline data allow for a higher-resolution assessment of sub-interval changes and help to distinguish fluctuations from mean trends. The trends and patterns of change evidenced in historical shoreline and bathymetric data are used to infer net sediment sources, sinks, and transport pathways.

The overarching goal of the study is to develop a better understanding of, and an ability to predict, large-scale coastal behaviour to more effectively and sustainably manage the coast. The CRLC is a superb natural laboratory given its sedimentary system of prograded barriers, depositional estuaries, abundant sediment supply, high energy regime, and large morphological changes as a result of high sediment transport rates, active-margin dynamics, and human interventions. The primary objective is to derive reliable change rates from morphological compartments within the first-order sediment-sharing system (i.e., the CRLC) to better understand the interactions among the estuary, inlet, ebb-tidal delta, barrier, upper shoreface, and lower shoreface. The secondary objective is to compare changes over the period prior to significant human influence (1700–1870s) with those of the modern era when human interventions began to alter natural processes. This study addresses several key questions:

- How has the littoral cell evolved over space and time?
- What relationships among the estuaries, inlets, ebb-tidal deltas, shoreface, and shoreline can be inferred from morphology change patterns?
- What sediment sources and pathways may explain historical shoreline progradation?
- What processes may explain the onset of an erosion trend along shorelines that historically prograded?
- What are the spatial and temporal effects of jetty construction at two inlets?
- How important is management of dredged material to coastal evolution?

1.2. Regional setting

Stretching between Tillamook Head, Oregon and Point Grenville, Washington in the Pacific Northwest of the United States, the CRLC comprises the geographic extent in which modern Columbia River sediment is deposited on the beaches; the beaches to the south and north receive insignificant quantities of that sediment (Clemens and Komar, 1988; Venkatarathnam and McManus, 1973). The 165-km littoral cell comprises four barrier sub-cells separated by the Columbia River and two large estuaries to the north: Willapa Bay and Grays Harbor (Fig. 1). The modern barriers and strand plains (i.e., prograded barrier beaches) of the CRLC built up sequentially following the filling of shelf and estuary accommodation space and the onset of a relatively slow rate of eustatic sea-level rise approximately 6000 years ago (Peterson et al., 2010-this issue-a). Approximately 4500 years ago, Long Beach and Clatsop Plains began to prograde, whereas Grayland Plains began to prograde about 2800 years ago, while the oldest portions of North Beach have sustained net progradation only for the last 2500 years (Peterson et al., 2010-this issue-b).

The CRLC is situated along an active tectonic margin of the Cascadia subduction zone that produces large earthquakes (magnitude ≥8) at approximately 500-yr recurrence intervals (Atwater and Hemphill-Haley, 1997). These episodic events cause coseismic coastal subsidence of 0.5 to 2.5 m (Atwater, 1996) and shoreline retreat on the order of a few hundred meters (Doyle, 1996; Peterson et al., 2000). Scarp formations in the subsurface of the CRLC barriers and strand plains (detected with Ground Penetrating Radar) provide evidence of this coastal subsidence (Meyers et al., 1996; Jol et al., 1996; Peterson et al., 2010-this issue-b). Meyers et al. (1996) and Woxell (1998) correlate the most seaward and recent paleoscarp to the A.D. 1700 Cascadia earthquake subsidence event on January 26, 1700 (Satake et al., 1996; Atwater et al., 2005).

Despite these multicentury-scale coseismic subsidence events, the CRLC barriers and strand plains have experienced net progradation (~0.5 m/yr) over the past few thousand years partly due to interseismic rebound, a large supply of fine sand delivered by the Columbia River (Woxell, 1998; Peterson et al., 1999), and a relatively intense wave climate capable of transporting the available sediment (Tillotson and Komar, 1997; Allan and Komar, 2000, 2006).

The wave climate along the U.S. Pacific Northwest coast is severe (Tillotson and Komar, 1997), with winter storms commonly generating deep-water significant wave heights greater than 10 m (approximately one event of this magnitude per year). The largest storms in the region have produced significant wave heights in the range of 14 to 15 m (Allan and Komar, 2002). High and long-period waves (averaging approximately 3 m in height and 12–13 s in period), high mean-water levels, and a west–southwest direction of wave approach characterize the winter months (November through February), while smaller waves (1.2 m and 8 s), lower mean-water