Intertidal sand body migration along a megatidal coast, Kachemak Bay, Alaska

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Using a digital video-based Argus Beach Monitoring System (ABMS) on the north shore of Kachemak Bay in south central Alaska, we document the timing and magnitude of alongshore migration of intertidal sand bed forms over a cobble substrate during a 22-month observation period. Two separate sediment packages (sand bodies) of 1–2 m amplitude and ~200 m wavelength, consisting of well-sorted sand, were observed to travel along shore at annually averaged rates of 278 m/yr (0.76 m/d) and 250 m/yr (0.68 m/d), respectively. Strong seasonality in migration rates was shown by the contrast of rapid winter and slow summer transport. Though set in a megatidal environment, data indicate that sand body migration is driven by eastward propagating wind waves as opposed to net westward directed tidal currents. Greatest weekly averaged rates of movement, exceeding 6 m/d, coincided with wave heights exceeding 2 m suggesting a correlation of wave height and sand body migration. Because Kachemak Bay is partially enclosed, waves responsible for sediment entrainment and transport are locally generated by winds that blow across lower Cook Inlet from the southwest, the direction of greatest fetch. Our estimates of sand body migration translate to a littoral transport rate between 4,400–6,300 m³/yr. Assuming an enclosed littoral cell, minimal riverine sediment contributions, and a sea cliff sedimentary fraction of 0.05, we estimate long-term local sea cliff retreat rates of 9–14 cm/yr. Applying a numerical model of wave energy dissipation to the temporally variable beach morphology suggests that sand bodies are responsible for enhancing wave energy dissipation by ~13% offering protection from sea cliff retreat.


1. Introduction

The spatial distribution and dynamics of beach sediments are crucial to understanding coastal processes and evolution. Along cliffed coasts, beach sediments offer sea cliffs a protective buffer from wave attack and along many barrier island coasts the beach sediments themselves comprise the only tangible real estate above sea level. Given the inherent mobility of beach sediment and the dynamic nature of nearshore forcing conditions, beach morphology is continuously evolving. Observed morphological changes can only be understood and ultimately predicted once the processes responsible are identified, quantified, and understood.

Much of the short- to medium-term morphological change of sedimentary coasts is linked to gradients in longshore sediment transport as well as to the character of the beach sediment itself. Numerous studies have documented the rates of gross and net longshore transport along well-sorted, sand-rich beaches [Komar and Inman, 1970; Duane and James, 1980; Dean et al., 1982, 1987; Haas and Hanes, 2004], resulting in a strong correlation between sediment transport rate and the longshore component of breaking wave power, a relationship first proposed by Komar and Inman [1970]. The dynamics of mixed sediment beaches, however, have received relatively little attention in the scientific literature [Mason and Coates, 2001] in spite of the estimate that bimodal beaches represent approximately 80% of the world’s nonrocky coastlines [Holland et al., 2004]. Understanding the processes responsible for the sorting of heterogeneous sediment, the shaping of intertidal bathymetry, and the mechanisms of longshore sediment transport along bimodal beaches will significantly aid in the management of coastal development, preservation, and restoration.

Herein, we present an example of a bimodal beach located in a megatidal (spring tide range greater than 8 m), sea-dominated setting in south central Alaska (Figure 1a). The mixture of beach sediment originates from two separate sources: (1) recent deglaciation has left behind a broad submarine bench composed of a cobble boulder nearshore substrate and (2) rapidly retreating sea cliffs, composed of loosely consolidated sandstones and mudstones, contribute...
a finer sedimentary component to the littoral system. This fine component is further sorted by nearshore processes; silt and clay is winnowed offshore, and the remaining fine to medium sand reorganizes into discrete polygonal bodies (hereafter called sand bodies) that travel along shore as coherent sedimentary masses (Figure 2).

In this paper, we focus our attention on the movement of these discrete intertidal sand bodies, utilizing the semi-diurnal exposure of beach provided by the megatidal environment. Using an Argus Beach Monitoring System, we document the spatial and temporal patterns of sand body movement over the cobble/boulder substrate. In conjunction with the Argus Beach Monitoring System (ABMS) data, we employ a tidal-contouring method to track beach profile changes resulting from alongshore sand body movement. Further, we use local wind and wave data to identify the environmental conditions responsible for driving this unique style of littoral transport, which leads to a hypothesis for the geomorphic evolution of the local littoral system.

2. Setting

Kachemak Bay is located in the north central Gulf of Alaska at latitude 59°35′N on the western side of the Kenai Peninsula. Figure 1(a) Silhouette map of Alaska identifying the Kenai Peninsula in south central portion of the state. (b) Map of Kenai Peninsula showing major geologic and physiographic features of the area. (c) Map of Kachemak Bay and Homer region. Stars identify locations of data collection stations used in this study. (d) Air photo of Homer spit with coastal features identified.
Figure 2. Solitary sand body visible at low tide along the intertidal beach near Munson Point, Homer, Alaska. Direction of sand body movement is eastward toward the left of photo.

Peninsula adjacent to lower Cook Inlet (Figure 1b). The bay and surrounding region display evidence of rock uplift by tectonics and sculpting by geomorphic processes of glaciation, fluvial incision, and coastal retreat. The north shore of the bay provides a unique natural laboratory for observing intertidal bed form migration and associated nearshore morphological changes because of a highly bimodal beach sediment grain size distribution, strong seasonality in environmental conditions, a gently sloped intertidal bathymetry, and a large spring tidal range that exposes nearly 0.5 km of beach twice daily.

2.1. Location and Geologic/Geomorphic Setting

[7] Tertiary alluvial sandy siltstones with interbedded coal seams on the north shore of the bay differ lithologically from the Triassic to Cretaceous metasedimentary and metamorphic rock units on the south shore [Bradley et al., 1997; Flores et al., 1997]. This lithologic contrast reflects the tectonic contact along the Border Ranges Fault (BRF) that forms the boundary between the inboard Wrangellia composite terrane and the outboard Chugach terrane (Figure 1). The BRF trends subparallel to the long axis of Kachemak Bay [Bradley et al., 1997].

[8] The cross-bay lithologic contrast is reflected in the dramatically different geomorphic character of the north and south shores of the bay. The north shore has a gently sloping intertidal zone backed by a marine terrace of variable elevation, ranging from 5 to 25 m above sea level. The south shore has abundant fjords, flooded embayments, and bedrock slopes that project steeply beneath sea level from the jagged, glaciated peaks of the Kenai Mountains.

[9] The long axis of Kachemak Bay aligns with the presumed axis of the Kachemak Bay lobe of the Kenai Lowlands ice expanse of the Naptowne glaciation [Karlstrom, 1964; Reger and Pinney, 1997]. Associated recessional moraines exposed along the sea cliffs in the city of Homer provide glacially derived, strongly bimodal sediment to the littoral system along the north shore of the bay.

[10] Shoreline change rates in the Homer area, estimated from historical air photos during the interval from 1951 to 2003, average roughly 1 m/yr, but are as high as 1.7 m/yr in the region of this study [Baird and Pegau, 2004]. These rates presumably accelerated after the 1964 Good Friday earthquake (Mw = 9.2) that caused local subsidence of as much as 2 m [Eckel, 1970]. This lowered the base of coastal bluffs, making them more prone to wave attack, undercutting, and episodic slumping. Sea cliff retreat and riverine inputs (though probably minor, given the relatively small coastal drainage basins) provide sediment to the intertidal zone, most of which is glacial in origin.

[11] Southeastward movement of nearshore sediment is responsible for constructing the most striking geomorphic feature in the area: the 6-km long Homer spit (Figure 1) which is the site of a large harbor and numerous local businesses making the spit an economically valuable piece of land of significant interest to the local community. Reger and Pinney [1997] hypothesize that the Homer spit formed as a submarine end-moraine complex by the partial grounding of a tidewater glacier, but no published studies have verified an age for the construction of this feature. The spit is located at the eastern (downdrift) end of the Homer littoral cell adjacent to a submarine trough present along the southern side of Kachemak Bay. This relatively deep trough may be a local sink for sediment moving along shore, limiting further growth/extension of the Homer spit.

2.2. Munson Point Study Site

[12] The morphological data presented in this study was centered on the beach at Munson Point within the city of Homer, Alaska (Figure 1c). The landform comprising the point is the lateral edge of a recessional moraine which now forms a ~5 m high sea cliff above the top of the bimodal beach below. Landward (north) of Munson Point is Beluga Slough, a tidally influenced wetland whose inlet (updrift of Munson Point) witnesses ebb tidal current velocities estimated to be greater than 1 m/s. The edge of Munson Point is approximately 1.5 km from the landward base of the Homer Spit, a site referred to as Mariner’s Lagoon (Figure 1d).

2.3. Oceanographic Climate

[13] Kachemak Bay has an area of 1500 km² and a shoreline length of over 540 km. Seven glaciers discharge freshwater with high seasonal variability. Deep trenches and holes extending to depths of 200 m characterize the bathymetry. The benthic nearshore habitat of the north shore consists of a mixture of boulders, cobbles, and sand, and kelp forests dominate subtidal depths to 20 m. At Homer, the mean maximum semidiurnal tidal range is 5.6 m with extreme tides exceeding a 9 m tidal range. Mean maximum tidal currents range from 0.5 ms⁻¹ on the ebb to 1.5 ms⁻¹ on the flood and mean maximum tidal excursions vary from 2 km during neap tides to 9 km during spring tides [Whitney, 1994]. Megatidal conditions coupled with the gentle slope of the intertidal zone on the north shore (~0.015) of Kachemak Bay expose a nearly 500 m wide beach at low tide in front of the city of Homer. This wide exposure allows for frequent documentation of changes to the morphology and character of the intertidal zone.

[14] The geostrophic (tidal) and baroclinic (density) currents driving the general circulation and water mass properties of Kachemak Bay were described by Burbank [1977], Muench et al. [1978], and Okkomen [2005]. The presence of
density currents driven by the strong freshwater glacial discharge alters the phase and duration of tidal currents [Okkonen, 2005]. Where mean density-driven flow is westward (along the northern shore), the onset of westward tidal flow (flood tide) occurs earlier and has longer duration than the onset and duration of eastward tidal flow. This results in a net westward excursion as shown by Lagrangian drifter experiments [Schoch and Chenelot, 2004]. It can be inferred from the asymmetry of the tidal flow that, over the fortnightly tidal cycle, tidal current velocities will be stronger during the spring tide and weaker during the neap tide. Furthermore, it can be inferred from the seasonal cycle of freshwater inputs to Cook Inlet (high inputs in summer and low inputs in winter) that density-driven currents will be weaker during winter than in summer. The general circulation pattern and drifter data are shown in Figure 3 and described by Schoch and Chenelot [2004].

In the nearshore, at depths less than 20 m, the currents are predominantly wave driven. Most waves that approach the north shore of Kachemak Bay are produced locally within Lower Cook Inlet by winds that blow through gaps in the topography on the western shore. The “Iliamna jet” is the name given to the wind stream that blows consistently from the west through the topographic gap between Mt. Iliamna and Mt. Saint Augustine (see Figure 1), two volcanic edifices within the Aleutian volcanic chain [Monaldo, 2000; Olsson et al., 2003]. The Iliamna jet is of particular importance because its overall mean direction is subparallel with the direction of greatest fetch across lower Cook Inlet. The result is a strong dependence of wave height upon local wind direction. Since the waves are locally derived and the greatest fetch is on the order of 100 km (from a direction of ~250°), wave periods average ~2.5 s and rarely exceed 8 s [CERC, 1984]. Strong seasonal variations in wave height reflect the contrast between the quiescent summer meteorological conditions and the stormy winter conditions that originate from the pressure differential between an atmospheric low over the Gulf of Alaska (i.e., the Aleutian low) and a high-pressure system over the interior.

2.4. Nearshore Sediment Character and Distribution

Intertidal beach sediments along the north shore of Kachemak Bay are strongly bimodal in their grain size distribution. The coarse component is glacial till, consisting of fist-sized cobbles (5–10 cm) to car-sized boulders (1–3 m), and comprises the rocky substrate that can only be mobilized by very large, infrequent, high-bed shear stress events. The fine component is fine-to-medium, well-sorted sand that experiences rapid episodic transport during discrete events.

The fine sediment moves over the coarse rocky substrate as distinct, solitary sand bed forms (sand bodies, Figure 2), whose wavelengths range from ~20 m to ~200 m, and have a maximum thickness at their leading edge crest of 1–2 m. The size of these sand bodies might qualify them as large to very large subaqueous sand waves [Ashley, 1990]. However, because of their disarticulated character and semidiurnal exposure in this megatidal setting, we opt to use the general term “sand bodies” throughout this paper. At the Munson Point study site, the sand bodies range in surface area from 5,000–15,000 m², and are roughly oval in plan view with straight leading edges (Figure 2). Assuming an average sediment thickness of 0.37 m (on the basis of field measurements), this translates to a volume of approximately 1,850–5,550 m³ of littoral sediment per sand body. From low-tide estimates, sand bodies cover approximately 20% of the intertidal zone at the Munson Point site, but the distribution is not uniform: The majority of sand body cover is in the midtidal zone (between 1.5 m and 3.5 m above mean lower low water), and sand is present in intermittent patches at the vicinity of the mean high-water shoreline. The leading edge of each sand body is typically steep and the trailing edge is gently sloped, defining a surface that is nearly parallel with the gently sloped intertidal beach upon which the sand body travels. Bulk sediment is inflated, or loosely packed (bulk density ~1,550 kg/m³, determined from a dried, weighed sample of known volume), and the top surface of each sand body is often decorated with current and oscillation ripples, suggesting active sediment transport.

Notably, at Bishop’s Beach (updrift) and along Homer Spit (downdrift), the sand bodies assume significantly different planform shapes from those at the Munson Point site (Figure 4). Both updrift and downdrift the sand bodies resemble the more commonly observed multiple elongate intertidal bars (“ridge and runnel morphology”) such as those described by Anthony et al. [2004] on coasts with a surplus of sand.

3. Methods

3.1. ABMS

Argus Beach Monitoring Systems (ABMS) observe and quantitatively document the coastal environment
[Lippmann and Holman, 1989; Holman et al., 1991, 1993; Plant and Holman, 1997]. These systems typically employ a group of digital video cameras mounted with overlapping fields of view at a remote fixed location, taking consistently timed images of the nearshore zone and storing them to an organized database. These overlapping digital images are merged and orthorectified, yielding map views of the study region that can be rendered at any frequency desired by the investigator (hourly photos during daylight hours in this study). Field surveys of stable features result in geometric solutions that provide a relationship between image coordinates (pixels) and real world locations [Holland et al., 1997].

While instantaneous snapshots are taken and stored, perhaps more valuable are the time-averaged exposures (3 min averages in this study). These serve to “average-out” the moving objects (wildlife, vehicles, people, etc.) that pass through the field of view that are not relevant to the goals of the project [Aarninkhof, 2003]. Most importantly the time-averaged exposures smooth through individual swash motions on the beach to delineate a time-averaged shoreline position. From the temporal sequence of map view images (snapshots or time-averaged exposures), the investigator can make accurate measurements of the timing and magnitude of shoreline change, calculate rates of sediment transport, and infer position of offshore sediment from the wave breaking pattern, to name just a few examples [Lippmann and Holman, 1989, 1990].

At the Munson Point–Homer site (http://zuma.nwra.com/homer/), we installed an ABMS system consisting of eight cameras spanning a field of view of approximately 220° in February of 2003. To acquire geometry solutions for the entire field of view, we surveyed the locations of cameras and ground control points (often large boulders within the field of view) with traditional methods. This allowed for the conversion from an oblique merged panoramic view to a merged planform view, such as the example shown in Figure 4, where each pixel in the planform view has true geographic coordinates. Using the spatial coordinates in each merged image, we tracked the movement of two separate, consecutive, sand bodies (first body: 1 March 2003 to 24 March 2004, second body: 24 January 2004 to 19 December 2004) and report the results below. In addition, we utilized the megatidal setting to contour the beach thereby obtaining initial videoderived estimates of morphological changes. By knowing the timing of tidal elevations at the Seldovia tide gage (discussed below), we were able to identify a known contour for each merged time-stamped ABMS image. The results of this “tidal-contouring” technique, presented below, were used to estimate cross-shore beach profiles and thicknesses of the sand bodies.

### 3.2. Field Surveys

During the summers of 2003 and 2004, we conducted kinematic differential GPS surveys of the beach within the region of our ABMS field of view. During both years, low-tide surveys were conducted on foot with GPS-mounted backpacks and during the 2003 campaign high-

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**Figure 4.** Sample-merged Argus Beach Monitoring System (ABMS) images from eight individual cameras, showing the transformation from photos to nearshore geographic data. (a) Panoramic view of intertidal beach. (b) Planform view after geometry solution has been applied. Sand bodies are outlined.
tide surveys were conducted from a small skiff outfitted with GPS and an echo sounder at the same transect locations, providing significant overlap of surveys to verify beach morphology. Data from these surveys were used to construct high-resolution digital elevation maps of the intertidal beach which we use to ground truth the data from the ABMS.

During the summer 2004 field campaign we also collected over 150 high-resolution digital still photographs of the sand component of the system using the “Beachball Camera,” an Olympus 5-megapixel digital camera fitted with macro lenses housed in a waterproof plastic case [Rubin, 2004]. Grain size analysis from digital images requires a calibration process in which sediments from the field are sieved into quarter-phi size class intervals and digitally photographed. A Matlab algorithm is then applied to generate an autocorrelation curve for each known size class resulting in grain size distribution data for each digital image requiring significantly less time and expense than traditional sieve analysis [Rubin, 2004].

3.3. Environmental Conditions

Local meteorological and oceanographic conditions were obtained from three sources. Wind speed and direction were recorded by the meteorological station (ID: FILA2) at Flat Island Light, Alaska, owned and operated by the National Data Buoy Center (NDBC); the data were obtained online (http://www.ndbc.noaa.gov). Wave height, period, and local tide level were recorded by a Sea-Bird Electronics SBE 26plus SEAGAUGE wave and tide gage deployed on the seabed, approximately 4 m below mean lower low water (MLLW) less than 1 km seaward of the approximate

Figure 5. Series of three oblique photographs (from ABMS camera 2) of intertidal beach along the north shore of Kachemak Bay showing sand migration. Leading edge of sand body is outlined and position of large beach boulder is marked as a reference location. (a) 27 February 2003: Leading edge is substantially west of beach boulder. (b) 30 July 2003: Leading edge is approaching beach boulder. (c) 29 November 2003: Sand body has overtaken boulder.
location of MLLW contour on the north shore of Kachemak Bay. Continuous tide levels were also recorded by a NOAA tide gage located at Seldovia, on the south shore of Kachemak Bay approximately 25 km south–southwest of Homer, also available online at the NOAA website (http://tidesandcurrents.noaa.gov). Locations of the wind station, wave/tide recorder, and the Seldovia tide gage are provided in Figure 1.

4. Results

4.1. Patterns of Sand Body Migration

[25] Eastward sand body migration toward the landward end of the Homer spit is evident from the oblique snapshots taken with the individual ABMS cameras (Figure 5). These views show a sand body overriding an intertidal boulder, conveniently located as a stationary object. Given the generous exposure of beach afforded by the large tidal range, we are able to view and record the sand body positions directly while subaerially exposed. This differs from typical ABMS applications that infer subaqueous nearshore morphology through breaking wave patterns [Lippmann and Holman, 1989; Holman et al., 1991, 1993; Holland et al., 1997; Plant and Holman, 1997]. While these oblique views provide no quantitative information about sediment movement, the merged planform images, obtained by stitching together time-averaged digital images from the individual ABMS cameras and applying the geometry solution from the field surveys, provide spatial constraints on the sand body positions through time (Figure 6). The correspondence of daylight hours and tide levels low enough to expose the sand bodies limits the frequency at which we can map locations of sand bodies. Also, adverse weather conditions can fog up camera lenses for several days at a time, making daily sand body position documentation difficult. Given these logistical obstacles, our attempted daily mapping interval, has some short gaps (several days at most) indicating breaks between clearly viewable images. Two example sand bodies, visible in Figure 4, show the variability of shape and size during the observation period. These sand bodies do not appear to migrate in concert; the leading sand body moves approximately twice the distance traveled by the trailing sand body between September of 2003 and March of 2004. Later in the study, however, the second sand body exhibits similar temporal pattern of travel as the first, when occupying a more downdrift position.

[26] Monthly mapped positions of the leading sand body, shown on Figure 6, reveal that summer months see very little migration, whereas movement during the winter is quite rapid. This is further quantified by examining the first sand body leading edge position through time (Figure 7). Three parallel alongshore transects (AT-1, AT-2, and AT-3) were selected as guides to track sand body leading edge position from the ABMS images. The three alongshore transects represent approximate high-, mid-, and low-tide shorelines, respectively. Straight lines (on Figure 7) connecting contemporaneous leading edge position points on the three alongshore transects show the approximate position of the leading edge every 5 days. These lines are tightly spaced during spring and summer of 2003, indicating slow sand movement, and are widely spaced during late fall and winter of 2003–2004, indicating rapid sand body migration. Total migration rates for each sand body, computed over the duration of the study, are 278 m/yr (0.76 m/d) and 250 m/yr (0.68 m/d), respectively.

[27] Cumulative movements of both sand bodies along three shore parallel lines are plotted together in Figure 8. The cumulative movement analyses display two distinctly different slopes for each sand body suggesting two distinctly different intervals of sand body migration rates. These seasonal rates are reported as the slopes of lines shown on Figure 8. The period of most rapid transport is clearly the late fall and winter, when sustained sand body migration rates average 1.41 m/d and 2.15 m/d, for the two sand bodies respectively. The quiescent period of slow transport during the late spring–summer–early fall witnesses sustained sand body migration rates of 0.11 m/d and 0.07 m/d, for the two sand bodies respectively.

4.2. Sand Body Characteristics

[28] The average mean grain size estimated from the approximately 150 digital sediment samples collected on several sand bodies present during the 2004 survey was 0.29 mm (medium sand). While mean grain sizes ranged from 0.15 mm (fine sand) to 0.83 mm (coarse sand), 95% of the sample means were within the fine-to-medium sand range. An example grain size analysis derived from sieving resulted in a normally distributed sample with a mean of 0.34 mm, well within the range of grain sizes derived from the digital images. In general, the sand fraction along the
north shore of Kachemak Bay is well sorted, however, there is a trend of coarsening toward the trailing edge of individual sand bodies with finer sands at the leading edge (not shown).

From the ABMS images, we obtain detailed information about spatial and temporal distribution of sand bodies in the intertidal zone during the study, but individual ABMS images alone do not directly provide information about the topography/bathymetry of these sand bodies. To acquire information about sand body geometry in the cross-shore (height, thickness, slope, etc.), we employ the height of tide during the time the images are taken as a known elevation datum, following the methods described by Plant and Holman [1997] and Aarninkhof [2003]. The tide gage record provides a time series of water elevations, providing elevations for various shoreline positions visible on the ABMS images, naturally contouring the intertidal bathymetry. Typical problems associated with using this method such as wave setup and runup are minimized in Kachemak Bay as it is relatively low energy in the summer. By noting horizontal positions of the intersection between water and land at known tidal levels, we can construct cross-shore intertidal beach profiles that display the shapes of the sand bodies in cross-sectional view. Subtracting a cross-shore profile at a given location in the absence of a sand body from a cross-sectional profile at the same location when a sand body is present provides information on the thickness, and hence volume, of sand in the package.

Detailed beach surveys within the ABMS field of view provided ground truthing for the tidal-contouring technique. We compare a cross-shore beach profile derived from the ABMS images in June of 2003 with a profile extracted from gridded differential GPS survey data collected during that same month, in Figure 9. The agreement between the two profiles is good above 1 m (MLLW) with a mean difference of 0.07 m and an RMS difference between the two techniques of 0.13 m. The scatter of the ABMS inferred elevations below 1 m (MLLW) is caused by the distance and angle from the camera resulting in poor resolution of the image pixels at long cross-shore distances.

Three cross-shore transects (CT-1, CT-2, and CT-3), spaced 200 m apart were used at quarterly intervals throughout 2003 (March, June, September, and December) to evaluate seasonal changes in beach morphology due to sand body migration. The sand body moves eastward through the temporal sequence of photos, leaving a record of bathymetric evolution, which we document through the “tidal-contouring” technique. Intertidal bathymetric data, derived by picking the shoreline positions at various known tidal levels through approximately 5 days worth of tidal cycles, are shown on 10. Although some scatter exists for any quarterly profile, differences in intertidal beach morphology between June and December are clearly visible, particularly on line CT-2. Comparing September and December data near the cross-shore position of 125 m on line CT-2 reveals nearly 1 m of vertical difference in the profiles. This reflects the presence of a sand body in September and its subsequent removal by eastward migration as of December. As the oblique ABMS photographs show an exposed beach in that vicinity in December, the vertical difference in profiles is in fact the thickness of the sand body.
4.3. Changes in Beach Morphology

GPS-derived gridded topographic intertidal surfaces collected in summer 2003 and summer 2004 illustrate the morphological change associated with sand body migration. Figures 11a and 11b presents beach contour maps constructed from summer surveys in 2003 and 2004, respectively, superimposed on corresponding ABMS images taken on the same day. The location of sand body 1 (SB1) and sand body 2 (SB2) are indicated on both images, identified both visually as well as by sand body influence on contour positions. The difference between the intertidal surfaces (Figure 11c) reveals a morphological system dominated by the migration of sand bodies. In the region of observation, the rocky substrate remains relatively constant in elevation. In contrast, sand body migration explains beach surface elevation gains and losses of up to 0.8 m in the regions where sand bodies travel. The topographic difference plot also indicates that the sand bodies are predominately migrating along shore, eastward toward the Homer Spit. However, in 2004 SB1 reached a higher elevation on the beach profile than in 2003 indicating that cross-shore processes also play a role in sand body migration.

4.4. Environmental Conditions

We assembled data documenting the local environmental conditions that coincide with our ABMS time-averaged video images, in order to identify the factors responsible for shaping the coastal landscape along the north shore of Kachemak Bay. Variations in wind speed and direction affect the strength of the local wave field. Waves carry geomorphic energy to the coast, and tide levels dictate the location of wave energy delivery within the beach–sea cliff system. It is therefore critical to document and understand the timing and magnitude of changes in environmental conditions to resolve the factors responsible for sediment transport and ultimately morphological change along the coast.

Time series of available data on wind, wave, and tidal conditions are displayed on Figure 12. Wind speeds, measured 34 m above MSL, during summer months are low, rarely exceeding 10 m/s, but increase during the fall and winter months with episodes reaching nearly 20 m/s. Wind direction is highly variable throughout the year, but has a tendency toward southerly winds during the summer. Wave heights stay small, below 1.5 m during the summer and early fall (0.5 m summer mean wave height), whereas storms in the early spring, late fall, and winter result in wave heights in excess of 2.5 m (0.8 m winter mean wave height). The spring tidal range is over 8 m, whereas the neap tidal range can be as low as 2.5 m. Figure 12d illustrates the time series of water levels with respect to mean lower low tide. Sand bodies can be mobilized by nearshore processes during intervals of submergence when water level is above approximately 3.25 m (elevation of the crest of sand body 1, transect CT-2, Figure 10), and are stationary during intervals of subaerial exposure when water level is below the elevation of the seaward extent of the sand body. Not shown are the wave period measurements, which average 2.4 s (standard deviation = 0.4 s) over the duration of the study.

Large wave heights correspond to strong winds, but only during periods when the winds blow from the west (Figure 13), the orientation with near-maximum fetch. During a 4-day interval in late December 2003 (interval 1 on Figure 13), strong winds (10–20 m/s) blew consistently from the west (260°–290°), forcing large wave heights with a daily moving average of approximately 2.5 m. Conversely, during a strong wind period in January 2004 (interval 2 on Figure 13), when 10–20 m/s winds blew from the northeast, wave heights did not respond because of the limited fetch, barely exceeding 0.5 m.

4.5. Sand Body Forcing

To understand the relationship between the observed temporal pattern of sand body migration and wave forcing, we plot weekly averaged sand body migration rates (averaged for each of the three shore parallel guide transects: AT-1, AT-2, and AT-3) against the time series of weekly averaged wave heights measured in Kachemak Bay (Figure 14). During intervals of high wave conditions in winter, sand body migration rates are greatest, often exceeding 3 m/d during several winter weeks of 2003 and 2004, reaching a maximum of >6 m/d in late December 2003. This forcing response relationship is further investigated on the correlation plot in Figure 15, where negative sand body migration rates have been removed from the data set, as have data collected during weeks where there were insufficient wave height readings were collected. The correlation is found to be significant at the 95% confidence interval. Figure 15 illustrates the general correlation between wave height and sand body migration rate, but the scatter of data suggests that there may be an uninvestigated variable influencing the migration rate. If the simple linear dependence of rate on wave height is assumed, the relationship takes the following form:

$$ \bar{R} = 7.1 \bar{H} - 0.3 $$

where $\bar{R}$ is weekly averaged sand body migration rate (in meters/day) and $\bar{H}$ is weekly averaged wave height.
(in meters). The function given in equation (1) is not forced through the origin of the correlation plot and the negative y-intercept value may be due to transport by currents originating from other processes in a direction opposite to wave-induced transport.

[37] This simple relationship, equation (1), may underestimate the role played by incident wave direction as previous studies have suggested a link between the longshore component of wave power and bed load sediment flux [Komar and Inman, 1970]. Because the correlation between migration rate and wave height (Figure 15) is not particularly strong, we explore the relationship between predicted longshore sediment transport and observed sand body migration. To do so, we use the CERC equation presented in the

Figure 10. Cross-shore profiles generated by tidal-contouring method applied to Argus data for mid-March, mid-June, mid-September, and mid-December profiles along transects CT-1, CT-2, and CT-3. Note the vertical extent of the sand body visible on line CT-2.

\[ Q_L = K \left( \frac{\rho_f \sqrt{g}}{16\gamma^{1/5} \left( \rho_s - \rho_f \right) (1 - n)} \right) H_b^{1/2} \sin(2\alpha_b) \]  

in which \( K \) is an empirical proportionality constant (set to 0.6), \( g \) is the gravitational acceleration constant, \( \gamma \) is the breaker index (ratio of breaking wave height to breaking wave water depth, set to 0.78), \( \rho_s \) and \( \rho_f \) are sediment and fluid densities respectively, \( n \) is the in-place sediment porosity (set to 0.4), \( H_b \) is the breaking wave height, and \( \alpha_b \) is the breaking wave angle. \( H_b \) and \( \alpha_b \) are computed through a simple linear shoaling calculation. \( Q_L \) is corrected by multiplying by the fraction of beach covered by sand bodies (0.2) and the fraction of time within the tidal cycle that sand bodies are submerged (0.44). Cumulative potential longshore sediment transport predicted by equation (2), and

Figure 11. High-resolution, differential GPS-derived, elevation contours of Munson Point–Homer beach draped over ABMS-merged photos derived from June 2003 survey (a) and July 2004 survey (b). (c) Beach volume change map. Note significant changes in beach volume associated with sand body migration.
corrected for fraction of cover and duration of submergence, for Kachemak Bay conditions starting in May 2003 is plotted with migration data for the first sand body observed in the study (Figure 16). Periods of high predicted potential sediment transport coincide with times of rapid sand body migration, most notably in November and late December of 2003. The total potential sediment transport calculated for the period shown in Figure 16 (slightly less than one year) is approximately 15,000 m$^3$, a value that will be used for comparison in an independent calculation of littoral sediment transport discussed below.

5. Discussion

[38] The unique combination of tidal range, gentle bathymetric slope, bimodal size distribution of beach sediments, and strong seasonality makes the north shore of Kachemak Bay an ideal site to document the behavior of mobile nearshore bed forms within the intertidal zone. Data collected during this study affords the opportunity to speculate on (1) the local littoral sediment budget, (2) the physical processes dominating littoral sediment transport in the area, and (3) the effect of migrating sand bodies on wave energy attenuation and sea cliff protection.

5.1. Littoral Sediment Budget

[39] The pattern of longshore sediment transport, as documented by the ABMS images in this study, is eastward toward the Homer spit. North of Anchor Point (Figure 1), coastal sedimentary landforms suggest northward longshore sediment transport up Cook Inlet. In agreement with the landform evidence, the Iliamna Jet, an atmospheric phenomenon (low-level jet) responsible for strong winds from the west across Cook Inlet, has been imaged with synthetic aperture radar [Monaldo, 2000; Olsson et al., 2003]. When the Iliamna Jet is active, strong winds from the most advantageous fetch direction (~250°) create waves whose rays approach Anchor Point, and split to drive longshore currents north of the point northward up Cook Inlet, and longshore currents east of the point eastward toward Homer. This implies that the sands passing by the ABMS site at Munson Point, are derived from sources along the 30 km stretch of coastline between Anchor Point and Homer, either from riverine inputs or bluff failures. We speculate that there is very little riverine contribution of sediment to the littoral system from the small catchment area, low-discharge, coastal streams (Travers Creek, Troublesome Creek, and

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**Figure 12.** Environmental conditions in Kachemak Bay from 1 January 2003 to 31 December 2005. (a) Wind speeds measured at Flat Island meteorological station (see Figure 1). (b) Wind direction measured at Flat Island. (c) Wave heights measured by nearshore wave gage in 4 m water depth offshore of Munson Point field site. (d) Tidal record as measured at Seldovia, across Kachemak Bay from Homer. Datum equals mean lower low water. Dark lines on wind speed and wave height time series are daily moving averages.

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**Figure 13.** Three month window of environmental conditions during the winter of 2003–2004 in Kachemak Bay. (a) Wind speeds measured at Flat Island meteorological station (see Figure 1). (b) Wind direction measured at Flat Island. (c) Wave heights measured by nearshore wave gage in 4 m water depth offshore of Munson Point field site (see Figure 1). Dark lines on wind speed and wave height time series are daily moving averages. Note correspondence of large waves with strong westerly winds during interval 1 and small waves in spite of strong winds during interval 2.
Diamond Creek on Figure 1). Several coastal slump blocks have produced piles of colluvial material at the base of oversteepened sea cliffs between Homer and Anchor Point. It is therefore likely that cliff backwearing and depletion of slumped material are the dominant sources of intertidal sand to the littoral zone. At the end of the Homer spit, a deep submarine trough (C24 100 m deep) trends roughly SW–NE and appears to limit the growth of the Homer spit. Sediment carried to the end of the spit likely cascades down into the submarine trough (a fraction probably leaks to northeast side of the spit), defining a local sediment sink.

The above assessment of littoral sediment sources, transport paths, and sink describes a well-defined littoral cell, as originally defined by Inman and Chamberlain [1960], which we will refer to as the Homer Littoral Cell (HLC). From the ABMS data, we can estimate the littoral transport rate for the HLC and, on the basis of some simple assumptions, make a rough calculation of regionally averaged sea cliff retreat rates.

Volumetric sediment discharge in the littoral zone can be estimated by

\[ Q_v = whsc \]

where \( w \) is the width of beach through which littoral sand is mobile, \( h \) is the average thickness of a sand body, \( \bar{s} \) is the annually averaged sand body migration rate, and \( c \) is the fraction of intertidal beach covered with sand bodies. Wave base (a proxy for maximum depth of sediment entrainment) is strongly tied to wave period, and given the short wave periods that dominate this study site, it is safe to assume very little sand is in transport below mean lower low water depth, though it is possible that large, wind-driven waves occurring during low tide might transport sediment outside of the observable intertidal zone. To simplify calculations, we set \( w \) to be equal to the distance from the base of the sea cliff to the mean lower low water shoreline, approximately 350 m at Munson Point. Average sand body thickness, \( h \), can be estimated from Figure 10 to be 0.37 m. We use the annually averaged sand body migration rates determined in this study: 250 m/yr and 278 m/yr, for the first and second sand bodies, respectively. Fraction of intertidal sand cover, measured along the beach between Munson Point and Mariner Lagoon, is found to be approximately 0.2. Evaluating equation (3) with these numbers yields a range of littoral transport rates of 4,400–6,300 m\(^3\)/yr, approximately half the littoral transport rate estimated for Miami, Florida [Dean and O’Brien, 1987]. These values are of the same order of magnitude, though less than, the values calculated for potential longshore sediment transport using the CERC formulation, equation (2), above, possibly because the littoral system is weathering limited as opposed to transport limited.

We can use the estimate of littoral transport rate from equation (3), with the assumption of negligible riverine sand contribution, to calculate a range of average sea cliff retreat rates, \( \bar{R} \) within the littoral cell through the following relation

\[ \bar{R} = \frac{Q_v}{Lbf} \]
where $L$ is the length of cliffed coast within the HLC, $b$ is the average cliff height and $f$ is the fraction of unconsolidated cliff material that is immediately incorporated into longshore littoral transport upon cliff failure. Using the range of littoral transport rates derived from equation (3), $L = 30$ km, $b = 10$ m, and an estimate of $f = 0.05$ (derived from the observed thickness of unconsolidated sand cover draped over the Upper Miocene to Pliocene Beluga and Sterling Formations comprising the sea cliffs [Flores et al., 1997]) we obtain a range of temporally averaged sea cliff retreat rates from 9–14 cm/yr. Given the episodic nature of sea cliff failure, however, retreat may occur as large (several meters) block failures every few centuries during anomalously vigorous storm conditions. Our calculation of sea cliff retreat rates is roughly an order of magnitude lower than those determined by Baird and Pegau [2004] ($\approx 1$ m/yr). This may reflect that our rates represent the present-day yearly averaged values, whereas the rapid rates determined by Baird and Pegau’s [2004] historical air photo analysis represent a transient response to the rapid subsidence experienced by the 1964 Good Friday earthquake.

5.2. Littoral Processes: Wave Dominated Versus Current Dominated

[43] ABMS camera monitoring of beach morphology within the intertidal zone reveals that sand bodies move at variable rates, but in consistent directions. Rapid movement is well correlated to large waves, which are correlated to strong winds blowing from the west–southwest. The temporal correlation of episodic sand body migration and large wave events in the bay suggests that littoral sediment transport is wave event driven, but other plausible sediment transport forcing mechanisms may be operating within the system and should be examined, particularly tidal currents associated with bay circulation, and wind-generated longshore currents.

[44] Cold, nutrient-rich water flows into Kachemak Bay from the Gulf of Alaska along the southern shore, swirls into two surface gyres within the inner bay, and moves out of the bay along the north shore [Burbank, 1977]. This circulation pattern, documented with drifter data by Schoch and Chenelot [2004] and illustrated in Figure 3, would move sediment westward along the north shore of the bay toward Cook Inlet, but the sand body migration data herein show only eastward movement. Furthermore, a typical sand body occupies a vertical position on the beach profile between 1.75–3.25 m above mean lower low water, which indicates complete subaerial exposure during 30% of the time-integrated tidal elevation record and complete submergence during 44% of the time-integrated tidal elevation record (Figure 17). When the flooding or ebbing tidal current velocity is at its maximum (at the inflection points of the tidal record curve), a typical sand body is not submerged, hence sediment is not accessible by the swiftest tidal currents. To examine the potential for transport by tidal currents when the sand body is submerged, we combine the “law of the wall” logarithmic turbulent velocity profile

$$\frac{u_{z,0}}{u_0} = \frac{1}{k} \ln \left( \frac{z}{z_0} \right)$$


with the Shields criterion for sediment entrainment

$$\theta_c = \frac{\tau_b}{(\rho_s - \rho_f)gd}$$


**Figure 16.** (a) Time series of potential sediment transport at Munson Point site calculated using equation (2). (b) Time series of sand body position for first sand body observed in study.

**Figure 17.** (a) Cross-shore profile of intertidal beach at CT-2 showing vertical extent of sand body. (b) Frequency distribution of hourly tidal elevation observations during this study. Dashed lines show elevation of mean higher high water (MHHW), mean sea level (MSL), mean lower low water (MLLW), sand body crest, and sand body base.
and the relationship between basal shear stress and shear velocity

$$
\tau_b = \rho_f u^* \ \ \ \ (7)
$$

to obtain

$$
u_{s,c} \geq \frac{1}{\kappa} \ln \left( \frac{z}{z_o} \right) \left[ \theta_c \left( \rho_s - \rho_f \right) g d \right] \ \ \ \ (8)
$$

$u_{s,c}$ is the critical surface velocity required to entrain sediment of size $d$, within a flow depth of $z$, for a roughness height of $z_o$, where $\kappa$ is von Karman’s constant (0.4 for clear water), $\tau_b$ is the basal shear stress, $u_*$ is the shear velocity, $\theta_c$ is the Shields parameter, $\rho_s$ and $\rho_f$ are sediment and fluid densities respectively, and $g$ is the gravitational acceleration constant. If $u_{s,c}$ is greater than the right-hand side of equation (8), then entrainment by a “steady” tidal current is possible. An unknown quantity in equation (8) is the roughness height, $z_o$, on the bed. A particularly “smooth” bed might have a roughness height controlled by the particle size ($z_o = d/30$), suggesting that a relatively sluggish tidal current could provide sufficient basal shear stress to entrain the sediment. However, if bed forms are present, the roughness height is substantially increased thereby slowing the flow. The roughness height is similarly increased by wave-current interaction [Ribberink, 1998].

Figure 18 shows an analysis of potential entrainment by tidal currents for typical spring tidal cycle along the north shore of Kachemak Bay using a value of 0.03 for the Shields parameter and a sediment particle size of 0.29 mm. It is noted that this is the minimum value of the Shields parameter necessary for incipient grain motion, though we expect a higher value should be used to mobilize the entire bed form. Figure 18a shows water level elevations at the Munson Point site on 14 June 2003. Figure 18b shows the submergence depth, $z$ in equation (8), over the crest of the sand body depicted in Figure 17a, during all times of submergence during the chosen tidal cycle. Using ABMS images taken with a frequency of 15 min over a complete
tial cycle we derive a time series of cross-shore current velocity estimated by temporal measurements of horizontal tidal inundation and use these data as a proxy for surface tidal current velocity (Figure 18c), to evaluate times of potential entrainment by tidal currents in Figure 18d. To obtain cross-shore tidal current velocities, shoreline positions on each ABMS plan view photo were chosen along the same cross-shore transect. The linear difference in shoreline position divided by the difference in time stamps for each ABMS photo (15 min) yielded the velocity. Therefore this method accounts for cross-shore variation in beach slope. The proxy tidal current velocities on the beach are an order of magnitude slower than tidal current velocities in deeper water, as measured by drifter data experiments [Schoch and Chenelot, 2004]. Dashed lines on Figure 18d show calculations, using equation (8), of surface current velocities required to entrain sediment on the crest (highest elevation) of the sand body depicted in Figure 17a. The solid line shows the absolute value of the surface velocity of the ABMS-derived tidal current generated during the tidal cycle. Even at maximum estimated surface velocity, the tidal current on the beach at Munson Point does not generate a basal shear stress sufficient to entrain the medium-grained sand that comprises the sand bodies examined in this study. This is not surprising given that the setting is a shallow water open beach environment where the tidal current is only a small fraction of wave generated currents and swash velocities.

[46] The potential for sediment transport by wind-generated currents is difficult to isolate at the Munson Point site, because the dominant wind direction coincides with the dominant wave approach direction. On the basis of the work of Whitford and Thornton [1993], we estimate that wind forcing exceeds one-quarter of the wave forcing less than 10% of the time during our study. Though it may be true that much of the sand body migration occurs during the strong wind and wave events, longshore current augmentation by wind is unlikely to constitute the majority of longshore sediment transport forcing at the Munson Point site.

[47] Because of (1) the observed direction of sand body movement, (2) the temporal correlation of large wave events with sand body migration rate, (3) the unfavorable conditions for sand entrainment by the tidal currents, and (4) the speculated small fraction of longshore current augmentation by wind, we posit that sand transport (sand body migration) at the Munson Point site is dominated by waves and wave-generated longshore currents.

5.3. Wave Energy Dissipation

[48] We hypothesize that the presence of intertidal sand bodies decreases the amount of wave energy delivered to sea cliffs and the upper beach face. To quantify the dissipative effects, we use a well-known model of surf zone wave energy dissipation [Thornton and Guza, 1983] to simulate wave energy losses on the beaches at the Munson Point study site. A detailed numerical modeling study was conducted for a range of wave conditions [Adams and Ruggiero, 2005]. Below, we summarize the modeling work and interpret the results within the context of natural coastal bluff protection.

[49] Several existing models that predict decay of wave heights and wave energy dissipation due to frictional drag and wave breaking within shallow water have been successfully tested against field data [e.g., Thornton and Guza, 1983; Dally et al., 1985; Baldock et al., 1998; Battjes and Janssen, 1978]. The common starting point in the development of these models is the wave energy flux balance

$$\frac{\partial (EC_f)}{\partial x} = -\varepsilon_f - \varepsilon_b$$

(9)

where $E$ is wave energy density. $C_f$ is wave group speed defined by

$$C_f = \frac{C}{2} \left[ 1 + \frac{2kh}{\sinh (2kh)} \right]$$

(10)

$\varepsilon_f$ is the loss of wave energy flux due to friction and $\varepsilon_b$ is the loss of wave energy flux due to wave breaking. $k$ is the wave number, defined as $2\pi/L$ ($L$ is the wavelength), and $h$ is the local water depth. The majority of the energy is lost in wave breaking and the turbulence associated with the propagation of broken surf bores.

[50] We employ the model proposed by Thornton and Guza [1983], in which the dissipation functions, for friction and wave breaking, respectively, are defined as

$$\varepsilon_f = \rho \nu \frac{1}{16\sqrt{\pi}} \left( \frac{2\pi f H_{rms}}{\sinh kh} \right)^3$$

(11)

and

$$\varepsilon_b = \frac{3\sqrt{\pi}}{16} \frac{\rho g B^3 f^2 H_{rms}^2 h}{\gamma^2 h^3} \left[ 1 - \frac{1}{\left( 1 + (H_{rms}/\gamma h) \right)^{3/2}} \right]$$

(12)

where $\nu_f$ is a bed friction coefficient, $f$ is average frequency of incident wave field, $H_{rms}$ is the root-mean-square wave height within the surf zone, $B$ is a breaker coefficient (usually set to 1), and $\gamma$ is the depth-limited wave breaking coefficient.

[51] Sample model results for midtide and moderate wave conditions ($H_{rms} = 1$ m, $T = 4$ s) are illustrated in Figure 19 and illustrate the effects of different intertidal bathymetry (presence versus absence of sand body) on the spatial pattern of wave energy dissipation through friction and breaking. The intertidal profiles beneath the sand body are identical for June and December of 2003 (Figure 19) as are the input wave conditions for each of the two simulations. Therefore it can be inferred that any differences in the spatial pattern of wave energy dissipation are attributable to the sand body. Maximum dissipation due to wave breaking is roughly 50 times greater than maximum dissipation due to friction (Figure 19) justifying our focus, hereafter, solely upon dissipation associated with wave breaking processes. Dissipation by wave breaking corresponds to steepness of bathymetric profile; the sand body appears to offer a natural protection to the upper beach face and sea cliffs through two mechanisms: (1) by moving the locus of wave breaking seaward and (2) by increasing energy expenditure associated with the turbulence of wave breaking.

[52] To assess the dissipative effects of the sand wave over a range of tidal heights and wave conditions, we
evaluate the wave energy flux at cross-shore locations marking the seaward and landward extents of the sand body, respectively ($P_S$ and $P_L$, on Figure 19), approximately 115 m apart. The difference in wave energy flux between these two locations is equal to the quantity of wave energy lost over the intervening cross-shore reach. Figure 20 shows the results of 34 simulations of cross-shore wave energy dissipation over both the June and December profiles at varying tide levels (2 m to 6 m above mean lower low water) for extreme wave conditions in Kachemak Bay ($H_o = 2$ m, $T = 8$ s). Figures 20a and 20b show that wave height and wave energy flux, respectively, decrease shoreward and respond to the presence of the intertidal sand body. Figures 20c and 20d display the percent wave height decrease and percent wave energy flux loss between positions $P_S$ and $P_L$, respectively. In the presence of the sand body, wave heights and energy fluxes are lowered to greater proportions than in the absence of the sand body, irrespective of tide. The amount by which wave dissipation is enhanced (taken as the difference between the black and gray curves on Figures 20c and 20d) is plotted in Figures 20e and 20f. Maximum dissipation enhancements are 19% and 13% for wave height and wave energy flux, respectively, and occur at midtide to high tide for extreme Kachemak Bay wave conditions.

As noted in previous work [Adams and Ruggiero, 2005] there is a positive feedback operating within the intertidal zone; the existence of a sand body enhances wave energy dissipation over the sand body, which in turn increases the amount of turbulent energy used to entrain and mobilize the sediments, propagating the sand body along shore, provided that there is an oblique component to the incident wave field. As viewed from the perspective of the sea cliff, there is a potentially valuable negative feedback; large waves attack the sea cliff, driving retreat, however the eroded material now in front of the sea cliff lowers the assailing energy of future incoming waves, protecting the cliff from further retreat. Shutting off this sediment supply, by anthropogenically armoring the coastline, could eventually remove the dissipative effect provided by the sediment and may foster more dramatic cliff erosion events during anomalously large wave conditions.

Lastly, we note the interesting flow and sediment transport mechanics that arise in a mixed sediment environment subject to the combined influence of waves and tides. The focus of our study has been to document the movement of discrete sand packages over a gravel/cobble substrate, and the consequences of the sand body migration.

Figure 19. (a) Bathymetric input conditions and (b) sample output from numerical model of wave energy dissipation applied to Munson Point study site for moderate wave conditions ($H_o = 1$ m, $T = 4$ s). Gray and black lines represent June 2003 and December 2003 results, respectively.

Figure 20. Wave energy dissipation modeling results for extreme Kachemak Bay wave conditions ($H_o = 2$ m, $T = 8$ s). (a) Wave heights through the surf zone for all 34 simulations. Gray lines computed for June profile (sand body present). Black lines computed for December profile (sand body absent). (b) Simulated wave energy fluxes. (c) Percentage wave height loss between $P_S$ and $P_L$. (d) Percentage wave energy flux losses. (e) Dissipation enhancement for sand wave presence derived from wave height output (difference of gray and black curves in Figure 20c). (f) Dissipation enhancement for sand wave presence derived from wave energy flux output (difference of gray and black curves in Figure 20d).
We have not investigated in detail the self-organization of the sand bodies themselves, or the mechanics of sand body movement over gravel, but hypothesize that the two processes are coupled. Increased turbulence over the hydrodynamically rougher cobble substrate could lead to preferential advection of sand grains toward areas of higher sediment concentration, e.g., sand bodies, and away from the interstitial spaces between individual cobbles. The discrete sand bodies resulting from this hypothesized self-organization mechanism then feedback with the nearshore hydrodynamics by increasing wave breaking and wave-induced currents which in turn mobilize the sand bodies over the cobble substrate. Future work will attempt to characterize and quantify, over various timescales, the coupled sediment transport dynamics and morphological development of this prototypical mixed sediment system.

6. Conclusions

This study documents a quantitative link between oceanic forcing and nearshore morphologic change. On the basis of a 22 monthlong set of digital video imagery that show the migration of sand bodies over a cobble substrate and data documenting local environmental conditions, we conclude that seasonally variable migration of 20–200 m long, ~1 m thick, self-organized, intertidal sand bodies is driven by episodes of high wave activity, and not driven by tidal or baroclinic circulation currents within Kachemak Bay. During the winter, these sand bodies move at weekly averaged rates of ~2 m/d, and during large storms we document migration as fast as 6 m/d. Sufficient waves are only generated within lower Cook Inlet, adjacent to Kachemak Bay, during periods of strong sustained southwesterly winds, a condition that occurs as a direct result of the Iliamna Jet meteorological phenomenon. These measurements of sand body migration provide estimates of littoral sediment transport rates and sea cliff retreat rates within the Homer Littoral Cell. It is further quantified that sand bodies provide a natural buffer to sea cliff retreat through wave energy dissipation.

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