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Scale-dependent perspectives on the geomorphology and evolution of beach-dune systems

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Abstract:

Despite widespread recognition that landforms are complex Earth systems with process-response linkages that span temporal scales from seconds to millennia and spatial scales from sand grains to landscapes, research that integrates knowledge across these scales is fairly uncommon. As a result, understanding of geomorphic systems is often scale-constrained due to a host of methodological, logistical, and theoretical factors that limit the scope of how Earth scientists study landforms and broader landscapes.

This paper reviews recent advances in understanding of the geomorphology of beach-dune systems derived from over a decade of collaborative research from Prince Edward Island (PEI), Canada. A comprehensive summary of key findings is provided from short-term experiments embedded within a decade-long monitoring program and a multi-decadal reconstruction of coastal landscape change. Specific attention is paid to the challenges of scale integration and the contextual limitations research at specific spatial and/or temporal scales imposes.

A conceptual framework is presented that integrates across key scales of investigation in geomorphology and is grounded in classic ideas in Earth surface sciences on the effectiveness of formative events at different scales. The paper uses this framework to organize the review of this body of research in a 'scale aware' way and, thereby, identifies many new advances in knowledge on the form and function of subaerial beach-dune systems.

Finally, the paper offers a synopsis of how greater understanding of the complexities at different scales can be used to inform the development of predictive models, especially those at a temporal scale of decades to centuries, which are most relevant to coastal management issues. Models at this (landform) scale require an understanding of controls that exist at both 'landscape' and 'plot' scales. Landscape scale controls such as sea level change, regional climate, and the underlying geologic framework essentially provide bounding conditions for independent variables such as winds, waves, water levels, and littoral sediment supply. Similarly, an holistic understanding of the range of processes, feedbacks, and linkages at the finer plot scale is required to inform and verify the assumptions that underly the physical modelling of beach-dune interaction at the landform scale.

Keywords: beaches; foredunes; sand dunes; coastal geomorphology; aeolian geomorphology; sediment transport; airflow dynamics; computational fluid dynamics; coastal erosion; sand

Introduction and purpose

Despite widespread recognition that landforms are complex systems with process-response linkages that span temporal scales from seconds to millennia and spatial scales from sand grains to landscapes, research that integrates knowledge across these scales is fairly uncommon (Bauer and Sherman, 1999). Our understanding of Earth surface systems is often scale-constrained due to a host of methodological, logistical, and theoretical factors that limit, either explicitly or implicitly, the span of what can be (or is being) studied (Sherman, 1995). As such, it is not surprising that traditional geomorphic research was incapable of providing critical insights into the conceptual bridges between fundamental process-response dynamics (studied at micro or meso scales) and long-term processes and controls that govern landform evolution over centuries and longer. This remains a key challenge for many Earth scientists and it is particularly true for aeolian-coastal geomorphology research that focuses on the evolution and maintenance of beach-dune systems that straddle the highly dynamic terrestrial-marine interface (Short and Hesp, 1982).

This paper reviews recent advances in understanding of beach-dune systems derived from over a decade of extensive and collaborative research that began in 2002 at the Greenwich Dunes on Prince Edward Island (PEI), Canada. The paper provides a comprehensive summary of findings from several short-term experiments embedded within a decade-long monitoring program and a longer-term (multi-decadal) reconstruction of coastal landscape change. Furthermore, the review situates the results from this specific research collaboration in a broader (global) context with research from elsewhere to draw attention to the challenges associated with scale integration in geomorphology. In particular, emphasis is placed on the constraints that neighboring (smaller and larger) scale perspectives impose on the understanding of knowledge derived at the scale of the event-based, instrumented field (plot scale) experiment. For example, measurements of net accretion along the toe of a coastal foredune using traditional cross-shore topographic profiles at monthly or seasonal intervals cannot reveal information about whether there was a gradual accumulation of sand through time, or deposition from a single event. Similarly, such survey data do not provide any information about potential intervening events that may have caused foredune erosion via wave scarping, nor can these data be used to extrapolate to long-term scenarios of dune maintenance and evolution without knowledge of the erosion-deposition tendencies over seasons, years and decades. Thus, without appropriate context, such survey data are of limited utility, revealing only what happened at a particular place and time.

Section 2 of the paper presents a conceptual framework based on classic ideas in Earth surface sciences that guide thinking on scale awareness and the effectiveness of geomorphic events in landform and landscape evolution. This framework provides the structure for an extensive review of scale-dependent research on beach-dune systems that is 'scale aware' and identifies critical gaps in knowledge. The review is grounded in the extensive research from the PEI study site (described in Section 3) but also considers a wide range of contributions, both classic and contemporary, from around the world. Sections 4 through 6 each provide a brief synthesis of classic knowledge and the state of the science up to the early 2000s. This is followed by focused summaries of major advances at distinct spatial-temporal scales (plot, landform, landscape for Sections 4-6 respectively) over the last 15 or so years. Section 7 offers an overarching summary of key advances at each scale, issues of integration between them, and presents future research opportunities and challenges.

Given the significant range of spatial and temporal scales covered in this review, the domain of long-term Quaternary studies is not incorporated. It is acknowledged, however, that glacio-isostatic adjustments and altered rates of relative sea-level (RSL) rise, for example, exert key controls on the evolution of global coastlines that, in turn, may have implications for littoral cell sediment budgets and influence millennial-scale evolution of beach-dune systems.

1. Conceptual Foundations: Effectiveness and Scale of Geomorphic Events

Advances in Earth sciences are typically incremental, often building on the research of past generations of scientists. Modern process geomorphologists, are often motivated by earlier works on complex system behaviour (or 'process-response' dynamics) in geomorphic environments. Wolman and Miller (1960), for example, argued that the largest magnitude events in Earth surface systems are not necessarily those that perform the greatest amount of work. 'Catastrophic' storms may have immense capacity to alter pre-existing landscapes over short time spans, but events of moderate magnitude may account for greater cumulative work in a system because of their more frequent recurrence within the historical sequence of events that yield landscape evolution. In contrast, seemingly innocuous events individually may not cause major landscape disruption, but can be significant in landscape dynamics because of the sustained work they perform over decades. Wolman and Gerson (1978) further argued that the degree to which an event may leave an indelible imprint on the landscape (referred to as geomorphic 'effectiveness') is not a simple, linear function of event magnitude but, rather, depends on: (i) the historical sequencing of events and their timing; (ii) the antecedent conditions that predispose a landscape for rapid change; and (iii) the capacity for a landscape to recover from the change imparted by the most recent event. Thus, a large catastrophic event may alter the landscape significantly, but the system may rebuild to the pre-event state under every-day processes that cause landscape change. In contrast, even small landscape disturbances may persist for decades if there is little capacity in the system to recover to the prior state.

The effectiveness of a geomorphic event, which must include its magnitude-threshold-frequency characterization, is also closely linked to the idea of equilibrium behavior (Thorn and Welford, 1994) by virtue of implicitly embedding events into a historical sequence that yields landscape change. The notion of 'embeddedness' was described masterfully by Schumm and Lichty (1965) who asserted that the spatial-temporal scale at which a geomorphic system is examined has implications for how system equilibrium may be manifested or perceived. Although Schumm and Lichty (1965) referred to each of the scale domains as representative of time, their framework implicitly embodies spatial dimensions. Based on these seminal perspectives on geomorphic effectiveness and the conceptual foundation established by Schumm and Lichty (1965) for general geomorphic systems, this paper provides a 'scale aware' approach for reviewing recent advances on beach-dune dynamics using a dominantly spatial reference terminology. Specifically, three characteristic scales of interest are identified: (1) the experimental 'plot' scale, which operates on 'steady' time scales of seconds to days and length scales of metres; (2) the 'landform' scale, which functions on 'graded' time (months to years) and on length scales of hundreds of metres; and (3) the 'landscape' scale, which operates on 'cyclic' time (decades to centuries) across length scales of kilometres. Table 1 proposes a list of variables applicable to a beach-dune system across the three scale ranges, which coincides with our predisposition to investigate geomorphic processes and landform dynamics at scales relevant to the human management of coastal resources. It is acknowledged that a different definition of scale may be needed to create a similar classification for other coastal systems (e.g., rocky, muddy, ice-dominated, etc.). Furthermore, challenges that exist at key scale transitions (plot-landform, landform-landscape) are identified and recent efforts to bridge them are discussed.

At the landscape (largest) scale, the dominant research interests of a coastal geomorphologist might include characterizing and classifying the dynamic nature of the coast according to whether the shoreline is prograding, aggrading, or retrograding, how wide and steep the beach is, and whether the foredune is receding or advancing and growing or shrinking. These factors are closely tied to the geometry and morphology of the beach-dune system (Short and Hesp, 1982; Hesp, 1988; Sherman and Bauer, 1993; Bauer and Sherman, 1999; Houser and Ellis, 2013; Hesp and Walker, 2013), and they collectively define the "dependent" variables (i.e., those that geomorphologists are interested to understand and predict). The main "independent" variables (i.e., those that serve as controls or drivers of system change and allow geomorphologists to gain insight into the dependent variables) are the geological framework of the coast (i.e., tectonic/isostatic setting, structural controls, rock type/history, fracture patterns, submarine bathymetry), eustatic sea-level trends, regional climatology, and exposure to oceanographic forcing (e.g., wave climatology, coastal currents, tidal fluctuations). Time is a relevant variable, simply because it takes time for major landforms to respond and adjust toward equilibrium.

Many of the landscape scale variables listed toward the bottom of Table 1 are considered “indeterminate” (i.e., those that have large variance but little impact on system dynamics at the scale of interest) because there is often insufficient information to adequately parameterize them and predict their state. For example, the degree to which a beach may have surface salt crusts, snow cover, and flotsam during an individual transport event is of limited importance to understanding whether there was (or will be) shoreline progradation or landward translation of the beach-dune profile at the scale of decades to centuries.

At the plot (smallest) scale, the research focus is on prediction of sediment transport at discrete locations over short time spans with the intent of understanding erosion and deposition across the beach-dune system as it relates to foredune maintenance and evolution. Thus, sediment transport rate and the pattern of flux divergence (leading to erosion or deposition) are the primary dependent variables in the system, whereas the primary driver is the near-surface wind vector (speed, direction) consistent with standard formulations of aeolian sediment transport models (e.g., Bagnold, 1941; Kawamura, 1951). In addition, there are a large number of supply-limiting surface controls, such as moisture content, snow cover, salt crusts, textural gradations, roughness elements (e.g., woody debris, wrack, foot prints, bedforms, lag deposits), that dictate the spatial-temporal pattern of sand transport across a beach-dune system (Sherman, 1990; Ellis and Sherman, 2013). Table 1 also catalogues "parameters", which are defined as controlling variables that are largely time-invariant at the scale of inquiry, although at larger scales they may be treated as time dependent variables. For example, in most plot-scale studies of sediment transport across a beach, it is reasonable to assume that foredune geometry is unchanged over periods of hours to days, that vegetation cover is constant because of slow growth rates (unless buried by a large event), and the tendency for the shoreline to erode or accrete is relatively unaltered if nearshore forcing is constant. Long-term factors such as mean wave conditions, climate patterns, relative sea-level trend and the geological context are not relevant at the plot scale, whereas at the landscape scale, these are dominant independent variables. As shown by Schumm and Lichty (1965), the specific combination of independent and dependent variables defining the dynamics of geomorphic systems will change depending on the scale of investigation.

Between the landscape and plot scales is the landform (intermediate) scale, which spans a period of time long enough to include seasonal cycles of adjustment as well as multiple extreme events. Table 1 shows that there is no overlap between the list of variables that are dependent (i.e., predictable) or independent (i.e., imposed) at the landscape versus plot scales, suggesting that knowledge gained at these end-member frames of reference is largely incommensurate. Research at the landform scale provides rich opportunities to connect these disparate knowledge domains, although research at the landform scale requires a commitment to longitudinal experimental designs that span a decade or more, which is often logistically challenging to maintain.

In many respects, the landform scale is the most challenging and demanding to conduct research in as it retains the requirements of a short-term (plot-scale) assessment with the need to scale up to a medium-term (landform-scale) understanding of processes that span a much wider range of variables. Moreover, the contextual controls imposed by the broader landscape scale are also relevant for understanding landform scale adjustments. Foredune maintenance and evolution is best understood with observations that span periods of many years to decades, and this category of understanding offers greater utility for management strategies intended to mitigate damage from human alteration of the coast and/or within a framework of climate non-stationarity and global sea-level rise (e.g., Davidson-Arnott, 2005; McLean and Shen, 2006; Hesp, 2013). Indeed, this was one of the key motivations for our research at Greenwich Dunes and our research partnership with Parks Canada.

Table 1: A proposed classification of system variables for beach-dune interaction that integrates across key spatial-temporal scales of reference from the plot scale (seconds to days, metres), to landform scale (months to years, 100s of metres), and up to landscape scale (decades to centuries, kilometres). This conceptualization is limited in scope to sandy coastal systems and scales relevant for human management of coastal systems, rather than the long-term geological evolution of the coastline.

Beach-Dune Variable	**Status of Variable During Spatial-Temporal Frame of Reference**		
	LANDSCAPE	**LANDFORM**	**PLOT**
Time	Independent	Not Relevant	Not Relevant
Geological Context	Independent	Parameter	Not Relevant
Sea-Level Transgression	Independent	Parameter	Not Relevant
Climatology	Independent	Independent	Not Relevant
Coastal Oceanography	Independent	Independent	Not Relevant
Shoreline Progradation/Erosion	Dependent	(In)dependent	Parameter
Vegetation Cover (Biogeography)	Dependent	(In)dependent	Parameter
Foredune Size And Geometry	Dependent	Dependent	Parameter
Beach Width And Slope	Dependent	Dependent	Independent
Surface Moisture & Snow/Ice	Indeterminate	Dependent	Independent
Salt Crusts	Indeterminate	Dependent	Independent
Surface Debris	Indeterminate	Dependent	Independent
Human Influences	Indeterminate	Dependent	Independent
Wind Approach Angle	Indeterminate	Dependent	Independent
Wind Speed	Indeterminate	Dependent	Independent
Sediment Transport Rate	Indeterminate	Dependent	Dependent
Erosion/Deposition Patterns	Indeterminate	Dependent	Dependent

2. Study Area: Greenwich Dunes, Prince Edward Island, Canada

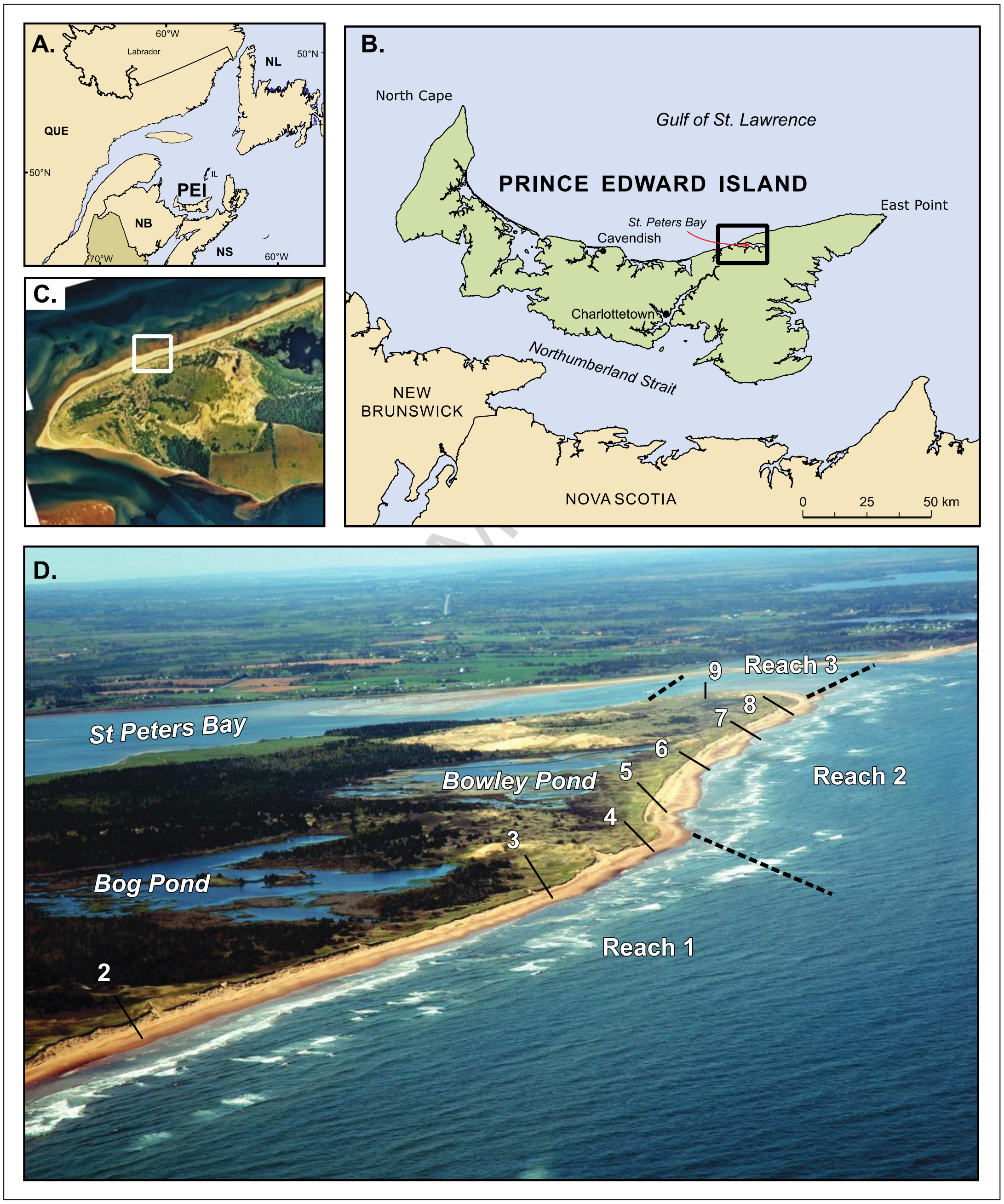
The Greenwich Peninsula is located on the northeastern shore of Prince Edward Island (PEI) in the Gulf of St. Lawrence in eastern Canada. The Greenwich Dunes complex was incorporated by Parks Canada Agency (PCA) into PEI National Park in 2000 to protect an area of established foredunes backed by wetlands, ponds, stabilized transgressive dunes, and a large parabolic dune complex (Figure 1). Much of the northeastern coast of PEI consists of horizontally-bedded, red sandstone, with some siltstone and mudstone, of Permian-Carboniferous age (van de Pol, 1983), that erodes readily. During the Holocene, the coast of PEI evolved under marine transgression and bedrock erosion. Shoreline retreat averaged about 0.5 m a^{-1} over the past 6,000 years (Forbes et al., 2004), which is similar to recession rates over the past half century along the NE coast (Webster, 2012). Sand supply associated with marine erosion and transgression is stored primarily as a thin wedge on the beach and inner shoreface, in barrier islands and mainland dune systems, and in flood and ebb tidal deltas associated with inlets of the barrier systems (Forbes et al, 2004; Coldwater Consulting, 2011). Today, low bedrock cliffs and headlands are typical with extensive sections of barrier islands and spits that enclose lagoons and shallow estuaries. The tidal range is micro-tidal (~1.0 m) with a mixed, semi-diurnal regime. Recent estimates of relative sea-level (RSL) rise for Charlottetown, PEI, give rates of land subsidence (due to glacial isostatic effects) of -1.45 mm a^{-1} and an estimated eustatic rise in sea level of +1.07 mm a^{-1}, producing an estimated RSL rise of about +0.25 m $century^{-1}$ (James et al., 2012). This value is roughly consistent with estimates of long-term RSL rise for the past 6,000 years of +0.3 m $century^{-1}$ (Forbes et al., 2004; Webster, 2012).

Prince Edward Island experiences a cool, temperate climate with a strong marine influence. Daily average temperatures range from a low of -8°C in February to a high of 19°C in August with maximum temperatures seldom exceeding 30°C. Average annual precipitation is about 1200 mm with less than 25% falling as snow, although winter snowfall amounts are highly variable. Prevailing winds at the site are from the SW, although strong northerly winds are common in March and April. Dominant winds from the NW, N and NE are driven by the passage of mid-latitude cyclones, which occur frequently in late fall through winter (October through

March) and exert significant control on precipitation and wind patterns in PEI (Manson et al., 2002; Forbes et al., 2004; Manson et al., 2015). Occasionally, hurricanes also track NE from the Caribbean, particularly in September and October. While the direct impact of hurricanes and post-tropical storms is generally moderate, storms tracking close to the area can interact with other mid-latitude systems to produce intense wind and wave conditions and storm surges that may persist for many hours. Extensive coastal erosion, flooding, and localized overwash of mainland and barrier dune systems is associated with extreme storms, as appears to have been the case for a major fall gale in 1923 (see section 5.2.3).

Foredunes at the site range in height from 8 to 12 m and have fairly uniform, straight seaward stoss slopes and a complex, undulating dune crest with intermittent depositional lobes and blowouts (Hesp and Walker, 2012). The foredune toe is occasionally scarped by waves during major storms but aeolian processes rapidly rebuild the slope by scarp in-filling. Incipient dunes up to 1 m high and 5-6 m wide also develop and can persist for 2-4 years between major storms. The dominant vegetation on the foredune is American Beach Grass (*Ammophila breviligulata*), whereas the annual Sea Rocket (*Cakile edentula*) is common on the backshore and occasionally Saltwort (*Salsola sp.*) is present. Beach Pea (*Lathyrus japonicas*) and Seaside Goldenrod (*Solidago sempiverens*) are common on lee slopes during the summer and fall months and shrubs such as Bayberry (*Myrica pensylvanica*) are found in more sheltered areas.

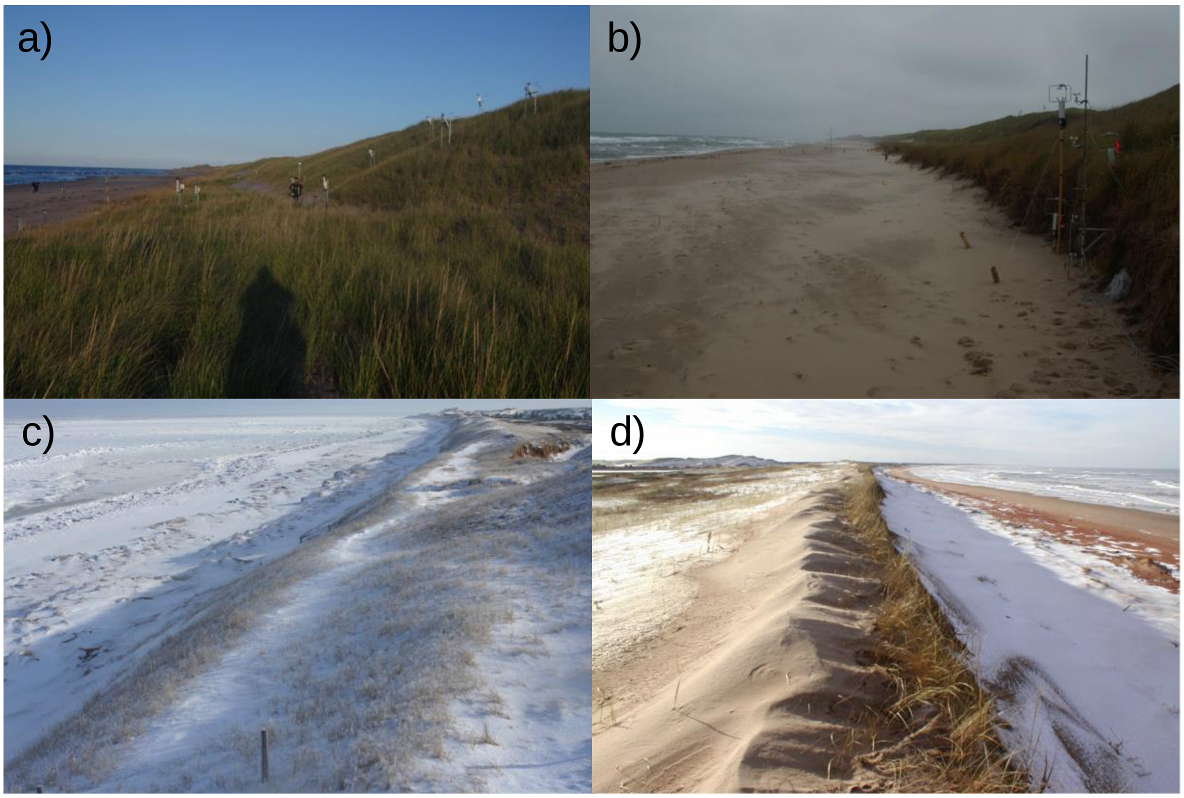
Figure 1: Location of study area showing: a) location of PEI in the Gulf of St. Lawrence and surrounding provinces; b) the Greenwich Dunes and St. Peter's Estuary area; c) vertical aerial photograph of Greenwich Dunes and the entrance to St. Peters Bay; d) oblique aerial photograph of the beach and dune system at Greenwich Dunes including the locations of characteristic study reaches (1-3) and cross-shore topographic profiles (see Ollerhead et al. 2013, and Figs. 21 and 25).



For the purposes of coastal erosion and dune dynamics monitoring for PCA, the study site was divided into three reaches (Figure 1d, each described in section 5.2.2). Representative profiles across the beach and foredune were established within each reach and surveyed annually between 2002 and 2012 (Ollerhead et al., 2013). Net littoral sediment transport is from E to W and there is evidence of about 100 m of westward progradation into the estuary over the past 80 years with several small foredune ridges formed over this time (Mathew et al., 2010). The littoral sediment budget is negative in Reach 1 with measured recession of the foredune of 0.5-1.0 m a^{-1} over the observation period. In contrast, the sediment budget in Reach 2 transitions from negative to positive, with Line 7 having a neutral budget (Ollerhead et al., 2013).

Plot scale field experiments were conducted in Reach 2 just west of Bowley Pond near Line 7 (Figs. 1d, 2a, b) to measure wind flow and sediment transport on the beach and foredune in May-June 2002, October 2004, October 2007, and April-May 2010. Beach width at this location was 30-40 m and sediments consisted of dominantly quartz sand with some feldspar with a mean grain size of 0.26 mm.

Figure 2: Site photographs of the beach and foredune system at Greenwich Dunes, PEI. The uppermost photos show the site for plot scale experiments in 2004 with a wide, vegetated incipient dune a) before the arrival of Tropical Storm Nicole and during the storm in b) with a visible eroded scarp at the foredune toe and active sand transport and deposition. Photo c) shows shorefast ice, snow cover, and vegetation dieback typical of the winter season. Photo d) illustrates distinct depositional lobes landward of the foredune crest that result from onshore sand transport and deposition in the winter months.



3. Plot scale

For most of the 20th century, aeolian geomorphologists have worked within a paradigm of steady, uniform flow for which (dry) sand flux is controlled mainly by the strength of the wind under what is referred to as ‘transport-limited' conditions. These conditions were replicated well within wind tunnel experiments that dealt primarily with horizontal, uniform beds of unimodal sediments, which also facilitated the development of theoretical models based on the fundamental physics of saltation. Plot scale field experiments were often sited to minimize topographic and surface complexity so as to conform to the paradigm. The development of ultrasonic 2D and 3D anemometers and fast-response sediment sensors has allowed aeolian geomorphologists to make high frequency measurements of turbulent wind flow and sand transport, which has enabled a shift away from the steady-state paradigm toward consideration of more natural conditions (e.g., Stout and Zobeck, 1997; Bauer et al., 1998; Sterk et al., 1998; Davidson-Arnott et al., 2005; Walker, 2005; Bauer et al., 2013). The plot scale experiments at PEI were specifically designed to explore the characteristics and effects of unsteady, non-uniform flow together with spatial and temporal variations in topography and surface characteristics.

This section provides a summary of research that was designed to characterize the complex flow dynamics over the beach and foredune and related patterns of sand transport. A summary and critique of traditional models of airflow dynamics and surface shear stress over low hills is provided as a starting point. More comprehensive reviews of secondary flow dynamics over dunes in general, and related semi-coherent flow structures over dunes, are provided by Walker and Hesp (2013) and Bauer et al. (2013), respectively.

3.1 Airflow dynamics over the beach-dune profile

3.1.1 Classic models of boundary layer flow over low hills

Theory on boundary layer flows over flat surfaces were extended to low symmetrical hills by climatologists (see Walker and Hesp, 2013), and seized upon by aeolian geomorphologists interested in predicting sand transport over dunes (e.g., Howard et al., 1978; Walmsley et al., 1982; Lancaster et al., 1996; Jensen and Zeman, 1985; Lancaster, 1985; Walmsley and Howard, 1985; Mulligan, 1988; Weng et al., 1991; Frank and Kocurek, 1996a; Wiggs et al., 1996b; McKenna Neuman et al., 1997, 2000; Walker and Nickling, 2002). The Jackson and Hunt (JH) model (Jackson and Hunt, 1975; Hunt et al., 1988) delineated 'inner' and 'outer' flow regions that resulted from topographically-forced streamline perturbations. Outer flow in the JH model is modified only by the pressure field, whereas within the inner region turbulent momentum transfers and surface shear effects are also considered and create two sub-layers: i) the thin, inner surface layer (ISL) where fluid shear is in equilibrium with surface roughness (i.e., the constant stress region) and ii) the overlying shear stress layer (SSL) where shear effects decrease with height until negligible. The JH model established a new theoretical framework for understanding boundary layer flow dynamics, successfully characterizing: i) flow stagnation and deceleration immediately upwind of hills and ii) flow acceleration or 'speed-up' on the windward (stoss) slope.

Rasmussen (1989) was among the first to apply a modified version of the JH model to a foredune. Due to varying roughness and slope transitions, he found that the depth of the ISL, from which surface shear stress is derived, was very thin and therefore traditional velocity-profiles measured using bulky instruments were of limited utility in estimating sand transport. Similarly, Hesp (1983) and Arens et al. (1995) found that flow accelerations up the windward slope deviated from those predicted by the JH model due to vegetation effects. They also noted that, as winds became more oblique, the effective slope (i.e., aspect ratio) of the dune decreased, reducing flow acceleration and the transport rate on the stoss slope. Arens et al. (1995) noted a decline in sand flux up the stoss slope at a rate that was dependent on incident wind speed. At slow speeds, the decline in sand flux was drastic, whereas at faster speeds sand traveled farther inland because of turbulent suspension. This effect was pronounced for steeper dunes and occurred despite changes in vegetation density.

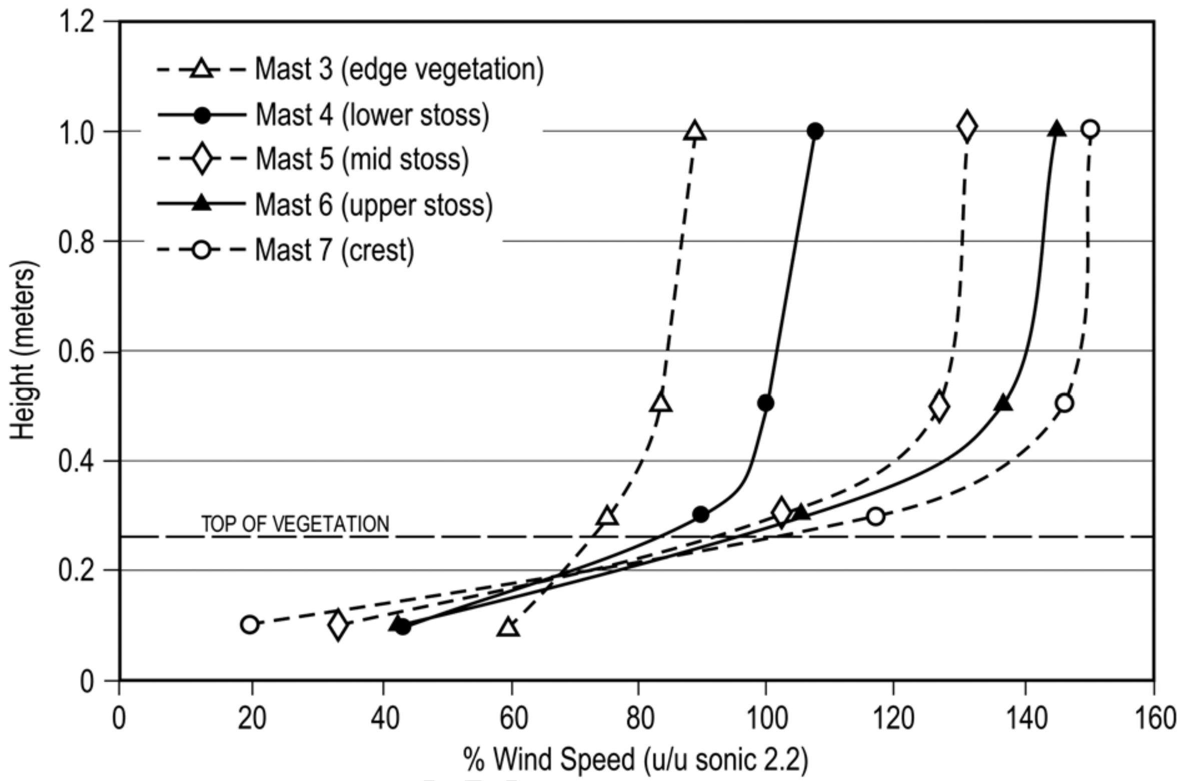
These early studies revealed that the ability to simulate flow dynamics over foredunes using climatological models was limited. Field experiments in the 1990s and 2000s also showed that typical foredune terrain leads to flow separation and flow reversal, unlike flow over a low hill (see Walker et al., 2006; Walker and Hesp, 2013). Empirical models of flow behavior in the lee of transverse desert dunes also emerged (e.g., Sweet and Kocurek, 1990; Frank and Kocurek, 1996; Wiggs et al., 1996; Walker and Nickling, 2002; 2003) and provided new conceptual foundations upon which flow dynamics over more complex, vegetated dunes could be understood.

3.1.2 Advances in flow dynamics over complex dune terrain

A vegetated foredune induces flow deceleration upwind of the dune toe, promoting deposition of sand at the bottom of the dune slope (e.g., Arens, 1996a; Davidson-Arnott and Law, 1996; Hesp, 1989; Sarre, 1989; Wal and McManus, 1993; Hesp, 2002). Beyond the foredune toe and up the stoss slope, the protrusion of the foredune into the boundary layer results in the compression of flow streamlines, increasing surface shear stress and wind speed toward a maximum at the crest. Accordingly, non-log-linear velocity profiles are common (Fig. 3). This type of topographic forcing on flow speed and shear stress distributions over aeolian dunes has been documented widely (see Walker and Hesp, 2013). The effect is most pronounced with wind perpendicular to the crest line and decreases steadily as the wind direction becomes more oblique (Hesp et al., 2005; 2015; Smyth and Hesp, 2015; Walker et al., 2006; 2009b).

In the PEI plot scale research, conventional velocity profiles were measured as part of some experiments (e.g., Hesp et al. 2009), although several drawbacks to this approach are recognized. Conventional anemometers (rotating cups, propeller-fuselage, sonic anemometry) are bulky compared to the shallow depth of the ISL. Thus, it is difficult to estimate shear stress in the thin constant flux region (where the Law of the Wall applies). Some researchers have measured velocity profiles that extend above the ISL (into the overlying SSL) as a proxy for estimating shear stress over desert dunes (e.g., Mulligan, 1988; Lancaster et al., 1996; Wiggs et al., 1996b). However, this often produces segmented and/or non-linear profiles (e.g., Bauer et al., 1990; 1996; Hesp et al., 2005; 2013; see Fig. 3). Additionally, a vegetation canopy over the foredune stoss slope imposes other limitations for applying boundary layer theory to estimate surface shear stress. Thus, careful sampling and assessment of flow conditions within the near surface zone is required if reliable sediment transport predictions are to be achieved (Bauer et al., 2004). This remains difficult with existing instrument designs (Walker, 2005).

Figure 3: Percentage wind speed profiles up the PEI foredune stoss slope from an experiment in 2002 (see Hesp et al. 2005; 2013). Speed observations at positions 3–7 are normalized against windspeed measured by a sonic anemometer at 2.2 m on a mast on the upper beach. Wind speed is topographically accelerated upslope above the vegetation, while within the vegetation, drag increases upslope and speeds decelerate.



The PEI experiments included measurements during flow conditions moderately above the threshold for sand transport in 2002 and substantially above threshold in 2004. In both studies, flow across the foredune was characterized by significant flow compression and acceleration. However, during the 2002 experiment a significant reduction in wind speed (deceleration) resulted over the foredune from enhanced drag exerted by the vegetation canopy as observed in other studies (e.g., Arens et al., 1995). During the gale event in 2004, there was a marked speed up above the vegetation, but also significant penetration of high-speed flow into the vegetation that, at times, produced sediment entrainment within the plant canopy (Hesp et al., 2005, 2009; 2013; Walker et al., 2009; see Fig. 3). The vertical (W) velocity component of the flow field was positive (upwards) across the stoss slope under slow wind conditions but shifted to negative (downwards) during gale conditions. In addition, a jet developed approximately 1 m above the vegetation canopy and extended from the upper stoss slope to the foredune crest during the gale event (Hesp et al., 2009; 2013). Formation of jet flow is common over distinct topographic breaks (e.g., Bowen and Lindley, 1977; Hsu, 1977, 1987; Tsoar et al., 1985; Arens, 1996a), but had not been observed on foredune stoss slopes (Hesp and Smyth, 2016a). These two phenomena, flow speed up within the plant canopy and jet flow development, are important for moving sediment to the lee of the dune during strong wind events (e.g., Arens, 1996a; Peterson et al., 2011; Hesp et al., 2009; 2013; Hesp and Smyth, 2016a).

Tall grassy vegetation exerts significant aerodynamic roughness that likely varies with wind speed as the plants flex downward and become more streamlined under extreme winds (e.g., Hesp et al., 2009). This dynamic behavior of the vegetation layer makes it difficult to parameterize surface roughness as an aerodynamic roughness length (z_0) or with a displacement height (d). This quandary is also a major limitation with current numerical modelling approaches (Smyth, 2016). As a result, time-averaged and spatially coarse velocity profiles over foredunes are likely inaccurate for characterizing the highly spatially and temporally variable surface shear stresses that drive sand transport.

3.1.3 New perspectives on turbulence and coherent flow structures

Much work has been done recently to describe time-averaged conditions and turbulent structures in flow over aeolian dunes (see Walker and Hesp, 2013) similar to earlier research in rivers (e.g., McLean and Smith, 1986; Nelson and Smith, 1989; Bennett and Best, 1995; Venditti and Bauer, 2005). Nevertheless, the relationship between turbulence intensity, Reynolds shear stress (RS = $-\rho\,\overline{u'\ w'}$ where u', w' are horizontal, vertical velocity fluctuations and ρ is fluid density), and sand transport across aeolian dunes remained essentially unexplored until the early 2000s following work on sand transport and turbulence over flat sand surfaces (e.g., Bauer et al., 1998; Sterk et al., 1998; Leenders et al., 2005; Baas, 2006).

Research over desert dunes and in wind tunnels demonstrated that RS at the toe of a dune often exceeds time-averaged, streamwise shear stress ($\tau = \rho\, u_*^2$, where u_* is shear velocity derived from velocity profiles) (e.g., Wiggs et al., 1996; Walker and Nickling, 2002; 2003; Parsons et al., 2004; Baddock et al., 2011; Weaver and Wiggs, 2011; Smyth and Hesp, 2015). Wiggs et al. (1996) argued that semi-coherent flow structures in the upwind boundary layer were conveyed toward the bed at the dune toe by concave streamline curvature in this region. These structures, which cause fluctuations in local RS, were thought to aid the maintenance of grain transport across the beach and through the flow deceleration region at the dune toe. Toward the dune crest, surface shear stress increases as a result of streamline compression and flow acceleration, assisted by streamline convexity that suppresses vertical motions and enhances horizontal fluctuations. These patterns of turbulence modification have been documented in flow over desert dunes (see Wiggs et al. 1996; Walker and Nickling, 2002;

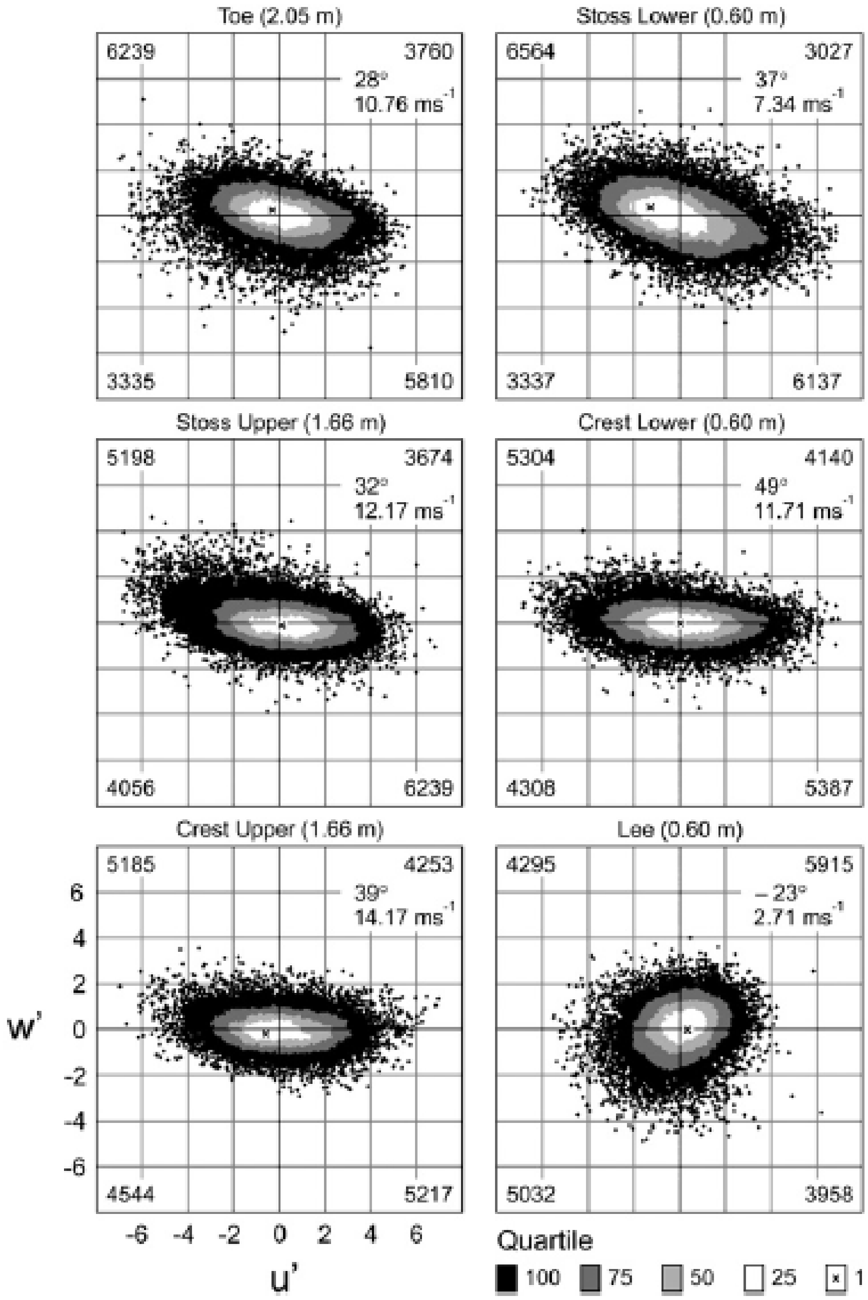
2003; Walker and Hesp, 2013) and over the foredunes at the PEI study site (Chapman et al. 2012; 2013).

Research in fluvial systems has shown that ejection and sweep events and larger macro-structures (e.g., kolks, boils) are often associated with enhanced sediment entrainment and transport via suspension (e.g., Jackson, 1976; Drake et al., 1988; Best, 1993; Robert et al., 1996; Roy et al., 1996, 2004; Best and Kostachuk, 2002; Kostaschuk et al., 2008, 2009; Shugar et al., 2010). However, few studies have focused on bed load transport, which is more comparable to the saltation-dominated mode of transport over aeolian dunes (cf., Drake et al., 1988; Valyrakis et al., 2010). Some of the experiments at PEI were dedicated to exploring the relationships between turbulent stresses (including semi-coherent structures) and sediment transport (Chapman et al. 2012; 2013) over foredunes using ultrasonic anemometry to acquire high-frequency (1-32 Hz) measurements of 3D velocity vectors (U, V, W) at two sampling heights across a transect extending from the upper beach to the lee of the dune crest. Sand transport intensity was measured using Laser Particle Counters (LPCs) positioned at 0.014 m and higher. Quadrant analysis was used to assess the distribution of quasi-instantaneous components of the RS signals over the foredune as a means to interpret potential links between fluid stress and resulting sand transport (Chapman et al., 2013).

Chapman et al. (2012) showed that the activity level in each of the four quadrants varied with height and position across the beach-dune profile (Fig. 4). Q2 activity ($u' <0$, $w' >0$), which is often associated with 'ejections', and Q4 activity ($u' >0$, $w' <0$), which is associated with 'sweeps', generally dominated the turbulence structure over Q1 ($u' >0$, $w' >0$) and Q3 ($u' <0$,

w' <0) activity, which conform to 'outward' and 'inward' interactions, respectively. Such Q2-Q4 skew is a characteristic signature of a turbulent boundary layer and was particularly evident across the beach, dune toe, and lower stoss slope of the foredune. In contrast, as the dune crest is approached, Q2 activity declines whereas Q1 becomes more dominant. The frequency of ejection and sweep activity is reduced toward the crest. In the lee of the crest, where flow separation occurs, the quadrant distributions were more symmetrical due to mixed, multi-directional flow. In terms of correlations between quadrant signatures and sand transport, Chapman et al., (2013) found that Q4 activity was most frequently associated with transport on the beach (52%), foredune toe (60%), and stoss locations (100%), whereas Q1 activity was dominant at the crest (25 to 86%), followed by Q4 (13 to 59%). Q3 activity appeared to be largely irrelevant in terms of correlation with observed sand transport at any location.

Figure 4: Quasi-instantaneous (32 Hz) quadrant plots derived from a 10-minute Run at 1700 h on 11 October 2004 during a gale force event. Average incident flow angle and resultant speed for each location are shown in the top right. Quadrant counts (in each corner) represent the total number of observations (modified from Chapman et al. 2012: Fig. 10).



Understanding the dominance of certain quadrants over others at varying positions across the beach-dune profile provides insight into why there is generally a poor correlation between sand transport and time-averaged RS, contrary to what might be expected across an extensive horizontal sand surface. Specifically, fluid fluctuations that yield activity signatures in Q2 and Q4 provide positive contributions to RS, whereas those in Q1 and Q3 are negative contributions. If either couplet dominates the distribution (as with diagonally-skewed ellipsoids shown in Fig. 4), there will be either positive or negative momentum transfer toward, or away from, the bed, respectively. However, when the activity signatures are balanced (i.e., a circular pattern), the positive and negative quantities balance each other in the time-averaged RS. Thus, it is possible to have intense activity in Q1 and Q4, as we find at the dune crest, which implies significant turbulent fluctuations in the streamwise (positive) direction, but poor correlation with vertical fluctuations. This situation yields a small value of RS, despite significant potential in the flow field to sustain sediment transport. As a result, the relationship between sand transport and turbulence across beach-dune profiles is complex and cannot be described well using RS alone (Chapman et al. 2013). Figure 5 presents a conceptual model that summarizes these relations.

Other research has examined the distribution of Reynolds normal stresses (i.e., u'^2, w'^2) and turbulent kinetic energy (TKE) in flow over desert dunes (e.g., Baddock et al., 2011; Weaver and Wiggs, 2011). Increasing evidence suggests that positive streamwise velocity fluctuations are associated with the bulk of aeolian transport (e.g. Bauer et al., 1998; Sterk et al., 1998; Schönfeldt and von Löwis, 2003; Leenders et al., 2005; Baddock et al., 2011; Weaver and Wiggs, 2011; Wiggs and Weaver, 2012). As such, the relationship between near-surface turbulence, especially RS, and sand transport is not as straightforward as in traditional equations that relate sand flux to surface stress directly and unambiguously.

3.1.4 Advances in understanding topographic steering of near surface flow and sand transport vectors

Interaction of regional wind flow with surface topography results in deviations in the magnitude and directionality of near-surface flow vectors - a phenomenon termed 'topographic steering'. The mechanics of topographic steering are driven largely by pressure differences that the flow field encounters along streamlines that traverse the dune toe (deceleration, positive pressure gradient) and stoss slope (acceleration, negative pressure gradient). More in-depth explanations of this mechanism are provided by Walker and Hesp (2013) and Hesp et al. (2015) and references therein.

Early work on topographic steering over beaches and foredunes (e.g., Svasek and Terwindt, 1974; Bradley, 1983; Mikklesen, 1989; Rasmussen, 1989; Arens et al., 1995; Hesp and Pringle, 2001) demonstrated that winds approaching a foredune at an oblique angle tend to be deflected toward crest-normal and that this effect is greatest when incident angles are between 30° and 60° to the crestline. Highly oblique winds less than 30° to the foredune crest (where 90° is directly on shore) are generally deflected parallel to the crestline. Recent research at the PEI site (Walker et al., 2006, 2009a, b; Bauer et al., 2012; Hesp et al., 2015) and elsewhere (e.g., Lynch et al., 2008; 2009; 2013; Jackson et al., 2011; Delgado-Fernandez et al., 2013, Smyth et al., 2011, 2012), suggests a common set of flow responses over morphologically simple foredunes. Bauer et al. (2012) presented a conceptual model (Fig. 7) of flow-form interaction over foredunes for a variety of flow approach angles from onshore (crest-perpendicular) through oblique, and offshore that also incorporates knowledge of resultant sediment transport vectors (Bauer et al., 2015).

From these collective empirical results, it is now clear that topographic steering plays a significant role in determining the near surface wind field and, consequently, the sediment transport pathways across the beach-dune profile during onshore, oblique, and offshore regional wind flows. To extend understanding beyond these empirical observations, a more detailed computational fluid dynamics (CFD) simulation of flow over the PEI foredune (Hesp et al., 2015) was conducted to simulate near-surface flow response in 10° increments from onshore (0°) to alongshore (90°) wind approach angles. The results are summarized below into: I) crest perpendicular winds – onshore and offshore; II) crest oblique winds – onshore and offshore; and III) shore parallel winds.

Figure 5: Conceptual model showing observed streamline behaviour, flow dynamics, Reynolds stress (RS) quadrant event activity, and sand transport responses over a foredune. (Chapman et al. 2013: Fig. 7).

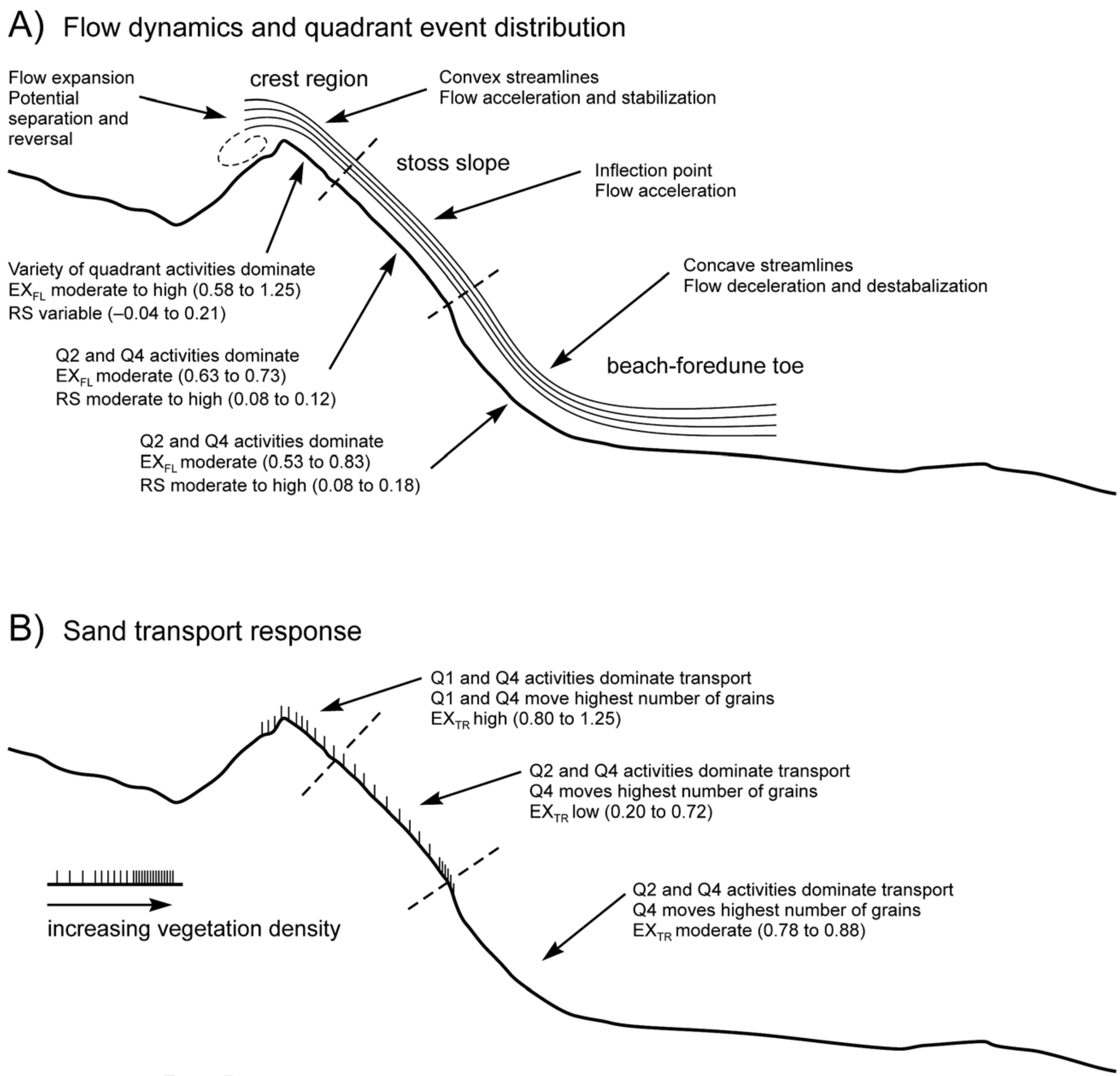
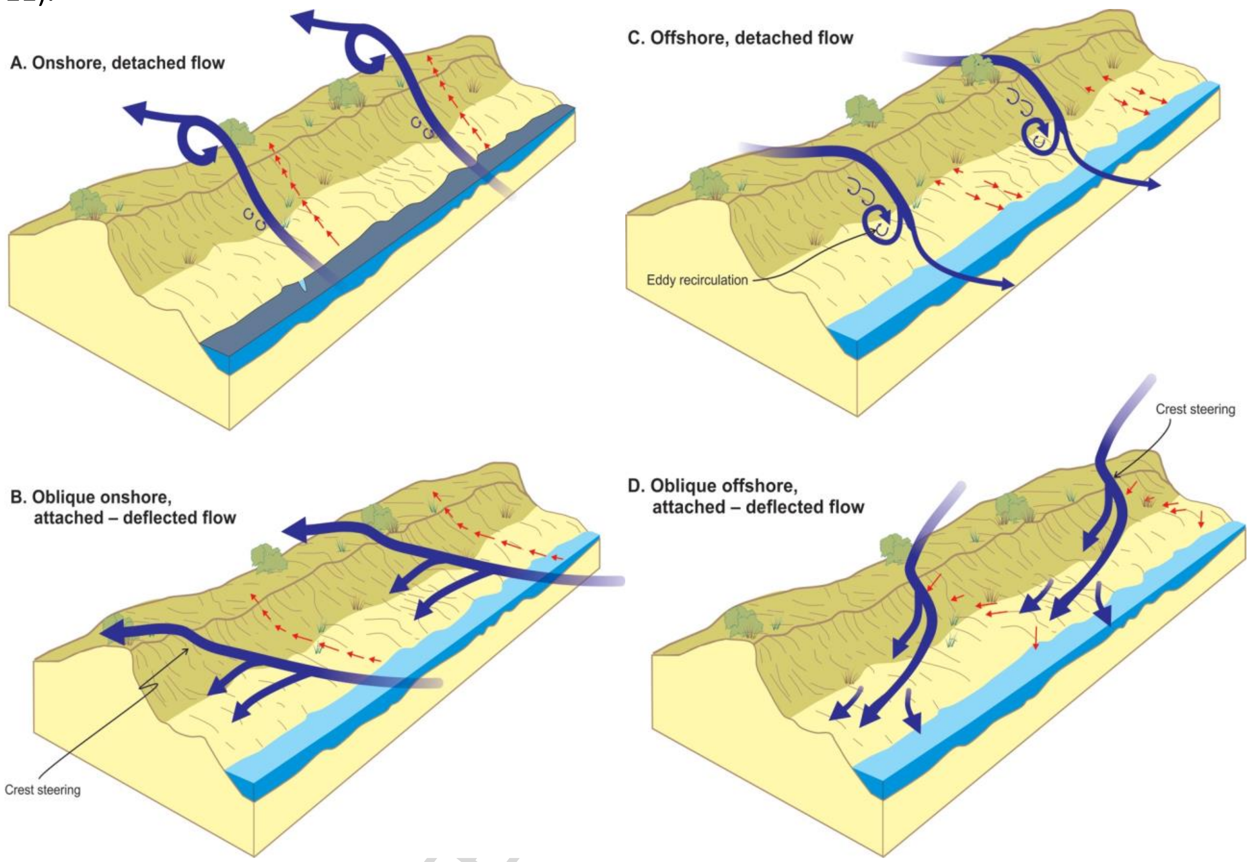


Figure 6: Conceptual model of flow–form interaction and topographic steering over a large foredune for variable wind approach directions. Large solid arrows correspond to near-surface wind flows and small arrows show likely sediment transport directions. (Bauer et al. 2012: Fig. 11).



I. Crest perpendicular winds

Crest perpendicular winds are accelerated up the stoss (upwind) slope of the dune and, if the foredune ridge is sufficiently high and steep, flow detachment occurs at the crest. A recirculation cell occupies the dune lee with a reattachment point located somewhere downwind depending on dune height and topographical complexity. Flow reversals at the bed are not uncommon (Fig. 6A and C) (e.g., Delgado-Fernandez et al., 2011; Jackson et al., 2011). During offshore winds, when the beach is in the 'lee' of the foredune ridge, anemometers located above the foredune crest and on tall beach towers record the regional (offshore) wind flow, while those close to the surface show drastically reduced flow speed and often reversed and highly variable wind directions, which are typical of lee side eddy circulation in general (Walker and Nickling, 2002; Jackson et al., 2011; Delgado-Fernandez et al., 2013; Bauer et al., 2012; 2015). The results of the PEI work on onshore and offshore flow conditions support detailed findings of others in Northern Ireland (Lynch et al., 2009; 2010; 2013; Jackson et al., 2011; Delgado-Fernandez et al., 2013) who documented distinct flow recirculation in the lee of a large foredune during offshore winds. During strong winds from either onshore or offshore directions, flow acceleration towards the crest can result in sand transport high enough above the bed to be incorporated within and above the lee-side flow separation eddy and deposited on the lower part of the downwind slope and beyond (Arens 1995; Peterson et al., 2011; Hesp et al., 2013). During offshore winds, some sand may be entrained near the crest and transported onto the upper seaward slope of the foredune, while on the beach, onshore transport may occur both seaward and landward from the point of flow reattachment, thus leading to a pronounced transport discontinuity (Bauer et al., 2012; 2015; Davidson-Arnott et al., 2012).

ii. Oblique winds

Winds approaching a foredune at an oblique angle are deflected toward crest-normal along the stoss slope (Fig. 6B and D) (Walker et al., 2006; 2009b; Hesp et al., 2015). The degree of deflection is dependent on incidence angle as well as height above the surface, with the most pronounced steering near the surface and nearer to the crest where flow acceleration effects are most prevalent (Arens et al., 1995; Mikkelsen, 1989; Walker et al. 2006; 2009b; Walker and Shugar, 2013; Hesp et al., 2015). Significant onshore steering of near-surface flow vectors can occur (as much as 37° from the incident wind as in Walker et al. 2009b), even during highly oblique winds.

Figure 7 shows CFD-generated flow streamlines in near-surface boundary layer flow (from 0.66 to 2 m) over the PEI foredune and depicts the resulting degree of streamline deflection for three incident wind approach directions (20°, 40° and 80°)(Hesp et al. 2015). The lowest streamlines show the strongest response to topographic forcing and display the greatest degree of deflection, similar to that observed empirically at the PEI site by Walker et al. (2006; 2009b). Near-surface flow speed responses show that the greatest speed-up occurs for winds that are most directly onshore when the dune has the steepest aspect ratio and then decreases as the incident wind becomes increasingly oblique. For example, at 0.66 m above the bed the wind speed at the foredune crest for incident wind directions from 50° to 30° to the crest is on average 25% lower than for winds in the 30° to 0° range (Fig. 8). Beyond the crest, flow separation occurs for onshore to moderately oblique winds and is manifest as a fairly simple reversing roller vortex, as in Fig. 6A above and as captured in smoke visualization by Walker (2005: Fig. 6). Flow separation and expansion results in notable flow deceleration leeward of the crest, particularly closer to the surface (Fig. 8B). However, as the flow trends towards more alongshore (from 50° to 70°), the degree of lee-side flow deceleration declines. This generally reflects a change in the effective aspect ratio imposed by the dune, such that from onshore (0°) to oblique-alongshore (~60°), incident winds still encounter a relatively steep and asymmetric topography. Beyond this range, as the incident wind approaches crest-parallel, there is significantly less topographic forcing due to the decline in dune aspect ratio, and little to no flow separation in the lee, as evident in the markedly different surface velocity distribution.

Figure 7: Examples of topographic steering of lower boundary layer flow (0.66 to 2 m) streamlines generated by a field-validated CFD simulation (Hesp et al., 2015) for three incident wind approach directions: 20° (oblique-onshore, uppermost), 40° (oblique, middle) and 80° (oblique-alongshore, lowermost). The lowest streamlines show the strongest response to variations in surface morphological changes and display the greatest degree of deflection. (Hesp et al. 2015: Fig. 8, reproduced with permission).

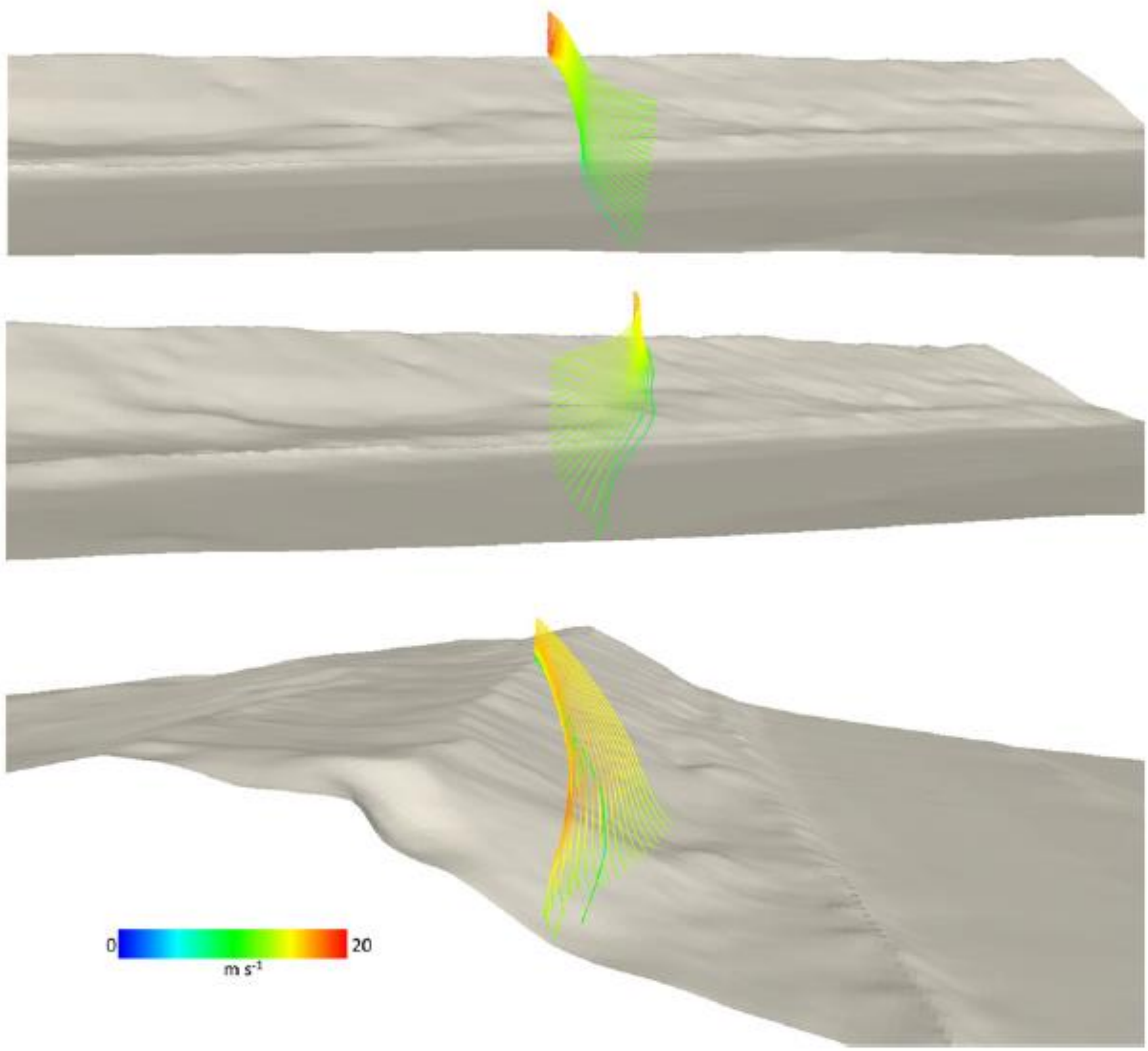
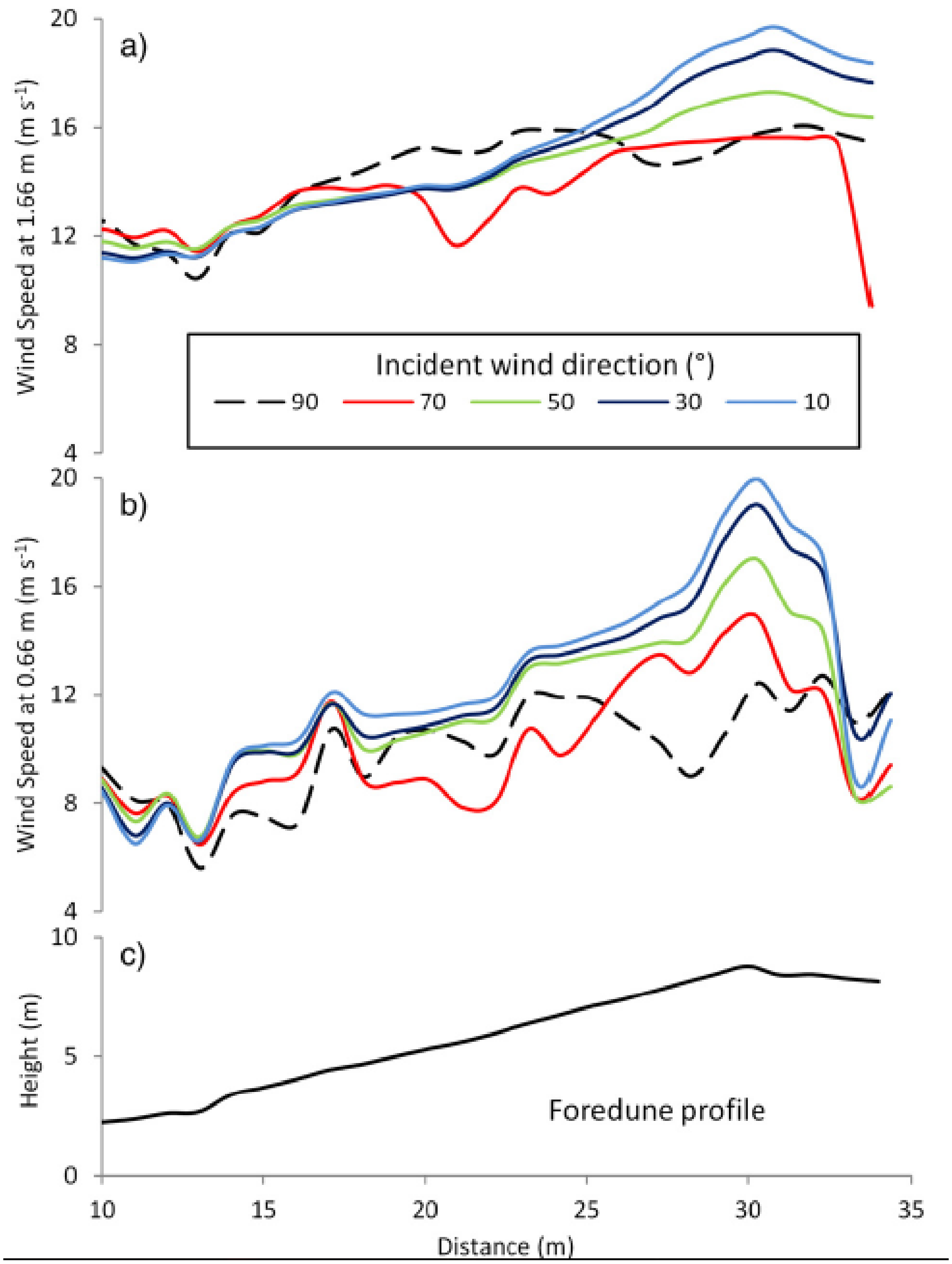


Figure 8: Near-surface wind speed responses generated by the CFD simulation of Hesp et al. (2015) showing speeds at 1 m intervals across the foredune at heights of 1.66 m(a) and 0.66 m (b) above the dune profile (c) for five incident wind directions. (Hesp et al. 2015: Fig. 7, reproduced with permission).



III. Alongshore winds

As the incident wind becomes more oblique (i.e., alongshore), the reduction in mean wind speed at the dune toe and the increase in wind speed toward the crest become less pronounced due to the declining effects of flow stagnation and streamline compression over the effectively less steep dune form (Arens et al. 1995; Parsons et al., 2004). Overall, there is also less spatial variability in near-surface flow speed over the dune (Fig. 8), although this does depend on the variability in surface morphology as well as vegetation cover and distribution.

The reduced flow acceleration effect over the dune for highly oblique flows can often result in increased sand transport potential along the beach (vs. into the foredune). However, the greater drag on wind flow over the vegetated surface of the lower stoss slope can also produce rapid wind speed decreases and some topographic steering towards the foredune toe, which may enhance sand transport from the upper beach onto the lower stoss slope. Transport potential over the stoss slope also decreases as a consequence of vegetation-induced drag, thereby creating a large disparity between sand transport on the stoss slope versus that on the beach.

If the foredune is scarped as a result of storm wave erosion, flow deflection patterns may be significantly different, than that for a non-scarped dune. Winds above the scarp may be deflected onshore towards the crest (Hesp et al., 2013) while flow seaward of the scarp is deflected along the beach during oblique and alongshore winds (Hesp and Smyth, 2016a), which may aid in the development of a dune ramp that rebuilds the eroded region (see Ollerhead et al., 2013).

The geomorphic implications of these flow deflection phenomena are important for several reasons. First, oblique winds can transport sediment onto a foredune or away from it depending on the angle of incident wind or the presence of a dune scarp, thereby affecting sediment supply to the dune. Second, deflected surface winds can influence net transport pathways and sedimentation patterns on the foredune, as has also been documented over transverse desert dunes. Third, transport conditions on the beach may be decoupled from those on the foredune at certain approach angles. Fourth, fetch distances and sand transport pathways into, and over, foredunes may be greater or less than predicted depending on the nature and magnitude of flow deflection. Finally, sedimentary strata may be deposited more crest transverse than the regional wind regime would indicate, thereby confounding paleo-environmental interpretations of relict dunes. Thus, assessments of landscape-scale dune evolution using regional wind statistics from nearby weather stations or relict dune morphology must also consider the confounding effects of topographic steering on near-surface flow patterns and the overall foredune sediment budget (Hesp and Hyde, 1996; Walker et al., 2006). In some settings (e.g., offshore oriented wind regimes), this may exert significant control on the total sand supply to, and/or the distribution of sand within, the foredune system (Hesp, 2002; and Davidson Arnott and Law, 1996; Walker et al. 2006; 2009a; 2009b; Lynch et al., 2009; 2010; 2013; Jackson et al., 2011; Bauer et al. 2012; Delgado-Fernandez et al., 2013), as discussed in Section 5.

At the plot scale, the nature and degree of topographic forcing on near-surface flow vectors is now conceptually understood and supported by rich empirical datasets and recent

CFD simulations (e.g., Parsons et al., 2004; Beyers et al., 2010; Jackson et al., 2011; Hesp et al., 2015). Implementation of this understanding into predictive models remains a challenge.

3.1.5 Innovative Computational Fluid Dynamics (CFD) modeling of flow over foredunes

The development of robust CFD modeling has significantly advanced our understanding of flow dynamics over dunes. Due to the logistical limitations of deploying field instrumentation to measure wind flow over complex terrain (Walker, 2005), CFD simulations are being used increasingly as a proxy and/or in conjunction with field measurements to accurately model complex flow behavior over aeolian landforms (e.g., Parsons et al., 2004; Omidyeganeh et al., 2013; Pelletier et al. 2015; Hesp et al., 2015; Hesp and Smyth, 2016a; Smyth, 2016).

CFD is a numerical method of solving fluid flow by converting the Navier-Stokes (N-S) equations to algebraic equations and solving them iteratively within a gridded computational domain of a study area. Unlike the Jackson and Hunt (1975) model, which solved the N-S equations linearly, CFD is capable of solving complex turbulent flow using a range of methods. The two most common approaches are Reynolds-Averaged Navier-Stokes (RANS) and Large Eddy Simulation (LES). RANS separates velocity and pressure into mean and fluctuating components, which are substituted into the original N-S equations producing a steady state solution of the mean flow dynamics. Unsteady or transient RANS (URANS and TRANS respectively) can also be calculated by retaining the unsteady terms, instead of averaging, making the dependent variables not only a function of space but also of time. LES produces a transient solution of flow dynamics by modelling smaller scale vortices, which are close to homogenous, and simulating larger-scale turbulence, which largely depends on geometry and boundary conditions. The locations in the mesh where the N-S equations are simulated (i.e., the N-S equations are solved) depends on the spatial resolution of the mesh and a spatial filter. Where the cells are larger (smaller) than the filter, the flow is calculated exactly (modelled using approximations).

Direct Numerical Simulation (DNS) of the N-S equations without any turbulence modelling is also possible. However, the computational power required to solve all scales of turbulence spatially and temporally makes the computational cost prohibitively expensive for use at high Reynolds numbers over aeolian landforms. To date, most studies of wind flow over aeolian landforms have been performed using RANS turbulence modelling, with the exception of Jackson et al. (2011) who compared RANS, LES and a hybrid RANS-LES model with measured data. In addition, Omidyeganeh et al. (2013) conducted an LES study of flow over a barchan dune at a relatively high Reynolds number, more akin to flow conditions found in fluvial environments. Building on this work, Pelletier et al. (2015) quantified turbulent shear stresses that produce grain flows on the slip faces of aeolian barchan dunes. Smyth (2016) provides a comprehensive review of recent progress in the use of CFD in aeolian research.

Despite recent advances, several limitations remain in CFD modelling of flow over aeolian landforms (Smyth, 2016). Most notable for research on coastal dunes is the ability to accurately model surface roughness imposed by vegetation. Vegetation drastically reduces wind velocity and shearing force exerted near the surface, which causes sediment to be deposited, which may over time result in increasing dune mass. In the majority of CFD codes, vegetation is simply parameterized as a fixed, surface roughness length. This parameter limits the vertical resolution of the computational domain, as the cell closest to the surface (where the roughness element resides) must equal twice the aerodynamic roughness length. The problem is compounded by the recommendations of Franke et al. (2004), who advise that at least two cells must exist between the surface and the area of interest inside the computational domain. This remains a key challenge in aeolian geomorphology as sediment transport is driven by flow dynamics very close to the surface within the ISL (see section 4.1.1), yet roughness lengths can extend to tens of centimetres within and through the ISL.

3.2 Instantaneous sediment transport across the beach-dune profile

4.2.1. Classic ideas on equilibrium 'saturated' sand transport

A great deal of effort has been devoted to understanding the detailed physics of aeolian saltation, usually under ideal conditions such as dry, unimodal sand on a flat, extensive surface without vegetation or moisture controls. Many aspects of saltation (e.g., grain-fluid momentum transfer, impact cratering, boundary layer adjustments) have also been simulated using complex analytical and numerical models (e.g., Bagnold, 1941; Anderson and Haff, 1991; Durán and Herrmann, 2006; Kok and Renno, 2009) but, in general, there is a presumption that the transport rate is in steady-state equilibrium with the wind. This has been referred to as the 'saturated' flux condition (Sauermann et al., 2001). In parallel, a large number of empirical studies have tested the performance of the basic predictive relations under natural field situations, with often disappointing performance. In early experiments, sand transport was measured with integrating traps over periods of 10-20 minutes and compared to values of u_* derived from the wind profile. Measured flux rates in the field were often much less than the maximum theoretical rate predicted for saturated sand transport (e.g., Sarre, 1988; Bauer et al., 1990; Sherman et al., 1998; Sherman et al., 2011). Sherman and Hotta (1990) summarized how the basic transport equations have been modified to accommodate the influence (usually singly) of supply-limiting factors such as surface moisture, binding salts, topographic slope, and sediment texture, which tend to reduce the maximum transport rate below that from standard models (see review in Ellis and Sherman, 2013).

In the 1990s, a number of fast response sensors for high frequency measurement of sand transport were developed and tested in the field, including: acoustic impact sensors (Spaan and van den Abeele, 1991; Arens, 1996; Ellis et al., 2009); piezoelectric impact sensors (Stockton and Gillette, 1990; Stout and Zobeck, 1997; Baas, 2004); and electronic balance traps (Jackson, 1996; Bauer and Namikas, 1998; McKenna Neuman et al., 2000). These sensors have permitted field measurements of "instantaneous" sediment transport in combination with high frequency measurements of wind flow. As a consequence, greater insight has been gained into the links between wind turbulence and the resulting characteristics of aeolian transport, including transport intermittency (e.g., Davidson-Arnott and Bauer, 2009; Davidson-Arnott et al., 2009; Davidson-Arnott et al., 2012) and the event-based nature of saltation (e.g., "flurry" characterization per Bauer and Davidson-Arnott, 2014). Advances in the ability to measure surface moisture content have also enabled improved understanding of non-saturated flux related to supply-limited conditions (e.g., Yang and Davidson-Arnott, 2005; Davidson-Arnott et al., 2008; Bauer et al., 2009; Darke et al., 2009; Delgado-Fernandez et al., 2009).

4.2.2. *Improved understanding of the fetch effect on beaches and sand delivery to foredunes*

Increasing evidence collected from field studies from the 1970s to the 1990s identified a persistent mismatch between measured and predicted transport rates on beaches (e.g., Svasek and Terwindt, 1974; Sarre, 1988; Bauer et al., 1990; Davidson-Arnott and Law, 1990; Nordstrom and Jackson, 1992; 1993). This compelled aeolian geomorphologists working on coasts to contemplate the ways in which the beach-dune environment is different from desert surfaces and wind-tunnel simulations. A primary factor involves the complexities of flow-transport interactions from open water to sandy beach to foredune (Sherman and Bauer, 1993; Hesp and Smyth, 2016b) that generates complex boundary layer adjustments, as well as specific constraints on sediment transport imposed by the 'fetch' effect (Gillette et al., 1996; Bauer and Davidson-Arnott, 2003; Delgado-Fernandez, 2010). Wind tunnel studies with dry, uniform sand showed that the distance downwind from a sediment source boundary required for the saltation cascade to achieve a constant transport rate (i.e., 'saturated' transport) was only a few metres (e.g., Nickling, 1988; Shao and Raupach, 1992; Dong et al., 2004), although this may depend somewhat on working section length, height and flow speed (Dong et al., 2004). However, it has long been recognized for agricultural fields that, where some form of supply-limiting factor exists, this distance can be significantly longer (Chepil and Milne, 1939). Coastal geomorphologists began exploring how important the fetch effect was for reconciling differences between measured and predicted transport rates across beaches, especially on the foreshore and lower beach (e.g., Svasek and Terwindt, 1974; Davidson-Arnott and Law, 1990;

Bauer et al., 1990). Just as there is a time lag or period of adjustment between the response of the saltation layer to a change in wind speed (e.g., Butterfield, 1999), there is a corresponding spatial distance over which such process-response adjustments occur (e.g., Shao and Raupach, 1992). The downwind distance that is required to achieve equilibrium transport via the saltation cascade is referred to as the 'critical fetch distance' (F_c). If one measures sediment transport downwind of F_c, then it is reasonable to expect that an equilibrium model (e.g., Bagnold, 1941) could be applicable. Within the fetch-limited zone ($F < F_c$), however, measured transport will always be less than that predicted by equilibrium-type models.

Figure 9 depicts the conceptual model of Bauer and Davidson-Arnott (2003), wherein the influence of fetch on sand supply to a foredune is characterized geometrically as a function of beach geometry (w/L) and incident flow angle (α). The model identifies the region landward (downwind) of F_c where sediment transport rate reaches a maximum (equilibrium flux) state, which, in turn, governs total transport into the foredune. Figure 10 shows various simulations that depict the magnitude of normalized specific sediment transport (relative to maximum rate per unit width) for a 1:1 (w:L) beach form for three different wind angles (α = 0°, 20°, 45°) and three fetch ratios (F_c/w = 0.2, 1.0, 1.3). Essentially, the simulations reveal that F_c exerts an important control on the amount of sand delivery to the foredune, but the proportion of sediment delivered to the dune, relative to the amount eroded from the beach, is influenced dominantly by angle of wind approach, not fetch. When angle of wind approach becomes more oblique, the downwind portion of the beach closest to the dunes experiences enhanced sand transport rates (ultimately reaching the equilibrium potential rate), however, the total amount of sediment supplied to the foredune actually decreases relative to that during shore-normal conditions as most of the sediment is lost to the downwind margin of the beach.

Figure 9: Conceptual model of Bauer and Davidson-Arnott (2003: Fig. 4) characterizing the fetch effect on a rectangular beach of length, L, and width, w. The beach is defined as the zone of dry sand between the limit of wave swash and the dune toe (limit of dune vegetation or significant break in slope). Critical fetch length, F_c, is the distance for aeolian sand transport to reach its maximum value (equilibrium flux rate). The shaded zone is the region where maximum values exist, as determined by wind speed, incident wind approach angle, α, and sediment size. F_m is maximum fetch resulting from the relationship between beach width relative to shore normal. Distance *l* represents a unit of alongshore length at the dune toe mapped out by two parallel streamlines of the wind field separated by perpendicular distance, *b*, such that $b = l (\cos \alpha)$. T represents a total transport line, or alongshore length of a line parallel to the dune toe that will receive sand transported from the beach for a given wind angle, α.

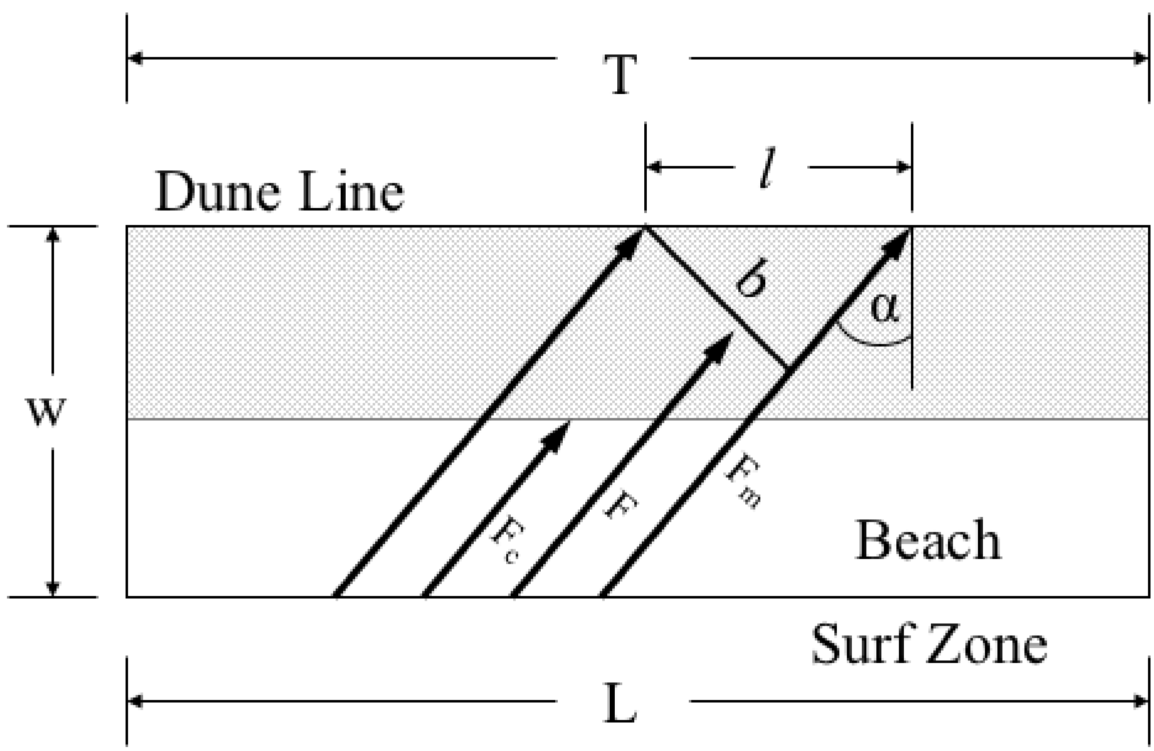
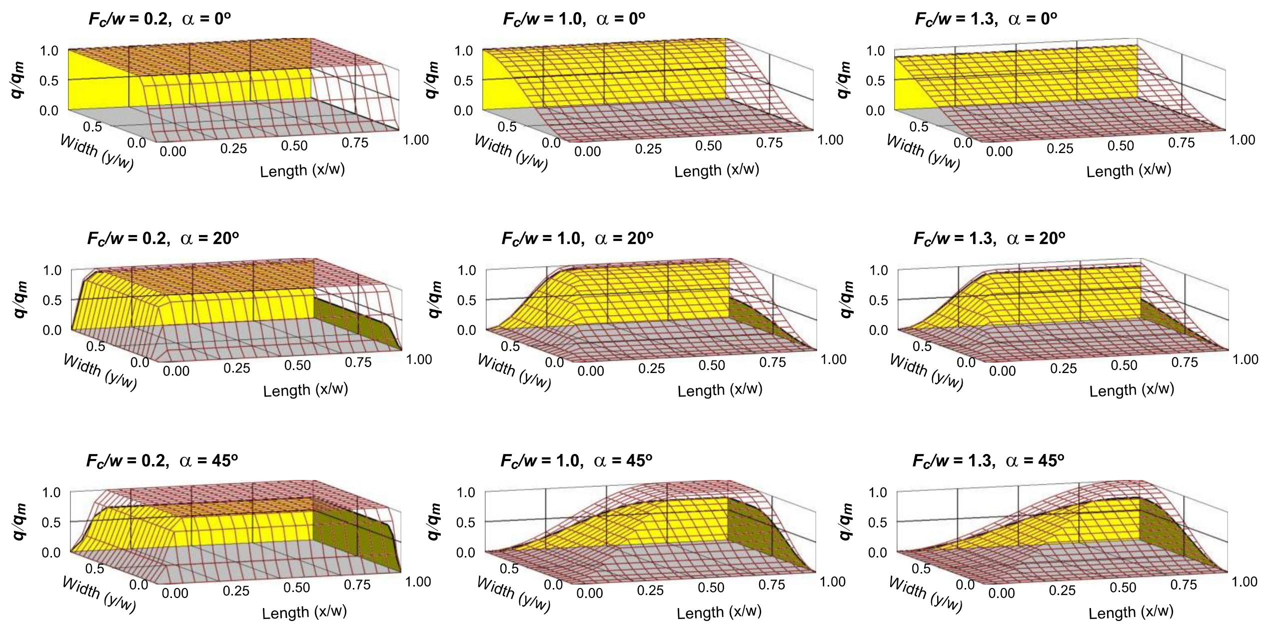


Figure 10: Distribution of normalized specific sediment transport (relative to maximum rate per unit width) over a beach with geometry w/L = 1.0 for a combination of three wind angles (α = 0°, 20°, 45°) and three fetch ratios (F_c/w=0.2, 1.0, 1.3) simulated by Bauer and Davidson-Arnott (2003: Fig. 9) based on the conceptual model presented in Fig. 9. The mesh grid shows magnitude of normalized specific transport parallel to the wind vectors and shaded portions of the axis planes indicate magnitude of transport across dune line or downwind margin of beach (as controlled by the cos α effect). Zones of net transport and erosion on the beach correspond to level and sloping regions of the mesh grid, respectively, where steeply sloping regions indicate intense erosion.



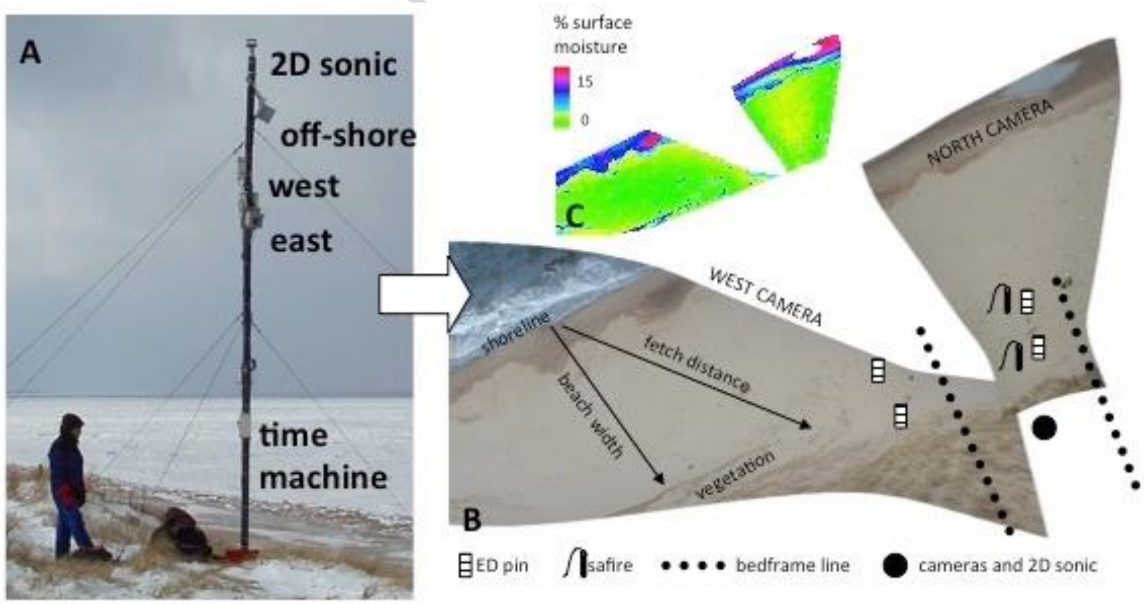
4.2.3. *Advances in understanding of fetch and moisture interactions on beaches*

In addition to the fetch effect, there are a host of other confounding natural factors that limit our ability to predict accurately the amount of sediment transport from beaches into foredunes. It is well known, for example, that increased surface moisture reduces the maximum rate of sand transport across a beach (e.g., Namikas and Sherman, 1995, McKenna Neuman and Langston, 2006; Davidson-Arnott et al., 2008; Edwards et al., 2012). It is also known that sand transport does not shut down completely during intense rain events, provided there is sufficient wind speed (e.g., Jackson and Nordstrom, 1998; McKenna Neuman and Scott, 1998; Hesp et al., 2009; Rotnicka, 2013). Wet portions of a beach, (e.g., foreshore, groundwater emergence zones) are subject to greater transport intermittency and spatial variability (e.g., Davidson-Arnott et al., 2005, 2008; Davidson-Arnott and Bauer, 2009), with the result that the critical fetch distance, F_c, will increase with increasing surface moisture (Davidson-Arnott and Dawson, 2001). All other factors equal, foredunes fronted by typically dry (wet) beaches will experience enhanced (reduced) sediment delivery and dune growth. Similar increases in F_c can be expected for other supply-limiting conditions such as the presence of pebbles or flotsam (e.g., de Vries et al., 2014).

Until recently, surface moisture measurement in the field was estimated gravimetrically by taking field samples to the laboratory – a tedious and time-consuming process. The Theta Probe impedance sensor was initially tested on beach surfaces by Atherton et al. (2001) and Wiggs et al. (2004), and this approach permitted rapid determination of average moisture at a sampling point over a depth of 0.1 m. Yang and Davidson-Arnott (2005) demonstrated that the probe length could be reduced to 0.02 m without significant loss in accuracy, thus permitting measurements that were much more representative of the forces related to moisture content very near the surface that constrain grain entrainment. Further evaluation of these instruments was done by Edwards and Namikas (2009) and Edwards et al. (2012). Rapid changes in surface elevation due to erosion and deposition in beach and foredune environments makes it difficult to deploy impedance sensors over long periods by simply embedding the instrument in the sand. Thus, repetitive sampling is required, with the prospect of unduly affecting the surface conditions.

An alternate near-field remote sensing approach based on surface brightness signatures from digital photographs has also been used to measure surface moisture (e.g., McKenna Neuman and Langston, 2006; Darke and McKenna Neuman, 2008). This method was applied to oblique photographs taken from cameras mounted on a tower on the dune crest at the PEI site (Fig. 11), thus providing coverage of an area on the order of 100 m^2 (Darke et al., 2009) and for a period of several months (Delgado-Fernandez et al., 2009; see section 5.1.2). Ortho-rectification and incorporation of these photos into a GIS facilitated the mapping of a number of other variables, in addition to moisture, on a regular basis (Delgado Fernandez and Davidson-Arnott, 2011; Delgado-Fernandez et al., 2012).

Figure 11: Components of the PEI long-term monitoring station located at the crest of the foredune (A). A 2D sonic anemometer was located at the top of a 5-m mast, with three digital SLR cameras below to take oblique, overlapping colour photographs of the beach and foredune toe region for ortho-rectification (B), and moisture mapping (C). Modified from Delgado Fernandez and Davidson-Arnott (2011: Fig. 2) and Delgado-Fernandez et al. (2009).



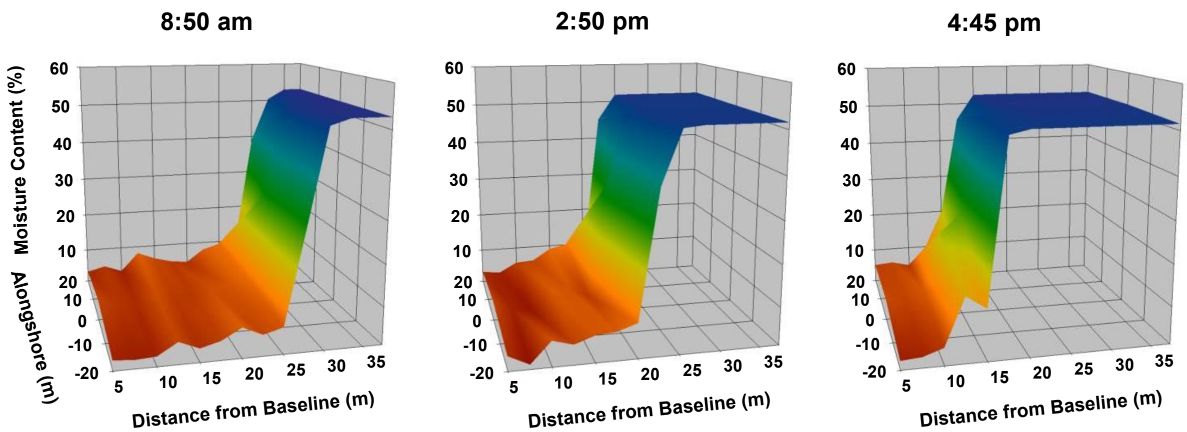
The control on aeolian transport imposed by surface moisture is ordinarily thought of as either a spatial phenomenon (i.e., zones or patches of wet or dry sand) or a temporal phenomenon (i.e., increasing moisture during storms and subsequent drying via evaporation). However, surface moisture exerts a supply-limiting control on aeolian sediment transport on beaches that varies in both space and time coincidentally. Indeed, the moisture state of a beach surface will often interact with the fetch effect to yield very complex process-response feedback loops (e.g., Nordstrom and Jackson, 1992; 1993; Bauer et al., 2009).

Consider the scenario of a wide beach that has experienced uniform surface drying via solar radiation for several hours and sand at the surface retains a moisture content of about 4%. A short-lived, onshore wind event begins that has the capacity to entrain sediment from the dry surface layer. Sediment stripped from the upper foreshore is transported to the foredune toe and deposited. As there is no supply of dry sediment to the foreshore from upwind, progressive erosion of the surface layer exposes wetter sediments beneath that are increasingly more difficult to entrain. As a consequence, the critical fetch distance, F_c, required for sand transport to reach its maximum (equilibrium) flux rate is effectively lengthened and the equilibrium transport zone fronting the foredune becomes narrower (see Fig. 10). Eventually, the berm and lower beach are stripped of dry sediment, exposing more closely packed, moist sediments, which further extends the F_c. In contrast, the upper beach and foredune toe are zones of deposition, comprised of newly delivered sediment that is dry and unconsolidated. The outcome of this scenario is that the system progresses from an initial beach surface with uniform moisture conditions to one with distinct zones of erosion, transportation, or deposition, in response to aeolian transport alone and without any changes other meteorological or tidal conditions that control the moisture state of the beach. Delgado-Fernandez (2010) and Delgado-Fernandez et al. (2012) describe field measurements consistent with these trends.

A different scenario is presented in Figure 12, which shows surface moisture conditions across the beach at the PEI site over an 8-hr interval (see Bauer et al., 2009). In the morning (0850 h), the upper beach was relatively wide and dry, with moisture contents ranging from 2-4% at the top of the beach, 4-6 % in the mid-beach, and saturated conditions in the foreshore zone from wave run-up. Moisture contents over the beach decreased slightly as the sun rose then increased with increasing wind speed and spray from wave breaking. As described below, because of drying of the surface layer by wind, low transport activity occurred occasionally even from areas that had 6% moisture or greater as a result of a relatively wide fetch zone.

During the day, wind speed increased progressively and, by 1000 h, sand transport was active across the entire beach except on the lower foreshore. Wind direction also shifted from essentially alongshore in the early morning to obliquely onshore by late morning. Despite a relatively narrow beach (< 20 m wide at 1200 h), sand transport across the upper beach remained active because of the oblique angle of wind approach, creating an effective fetch length > 80 m. By 1450 h, a combination of enhanced wave set-up, run-up, and rising tide caused the lower half the beach to become saturated and, by 1645 h, almost the entire beach except a 5-m strip in front of the foredune was either totally or periodically inundated. So, even though the wind field was competent to transport sediment, aeolian activity was inhibited by excess surface moisture across most of the beach. These types of complex interactions on beach-dune systems that involve changes in fetch distance that result from the interaction of wind angle, wind speed, wave set-up, tidal excursions, and rainfall, and, in turn, they can have significant implications for modelling sand supply to the foredune over a period of months to years (see section 5.2.3)

Figure 12: Surface moisture contents across the beach at the PEI study site from the foredune toe (baseline origin) to lower foreshore over an 8-hr interval during which wind speeds increased above transport threshold by 1000h and wind direction shifted from alongshore to obliquely onshore by late morning.



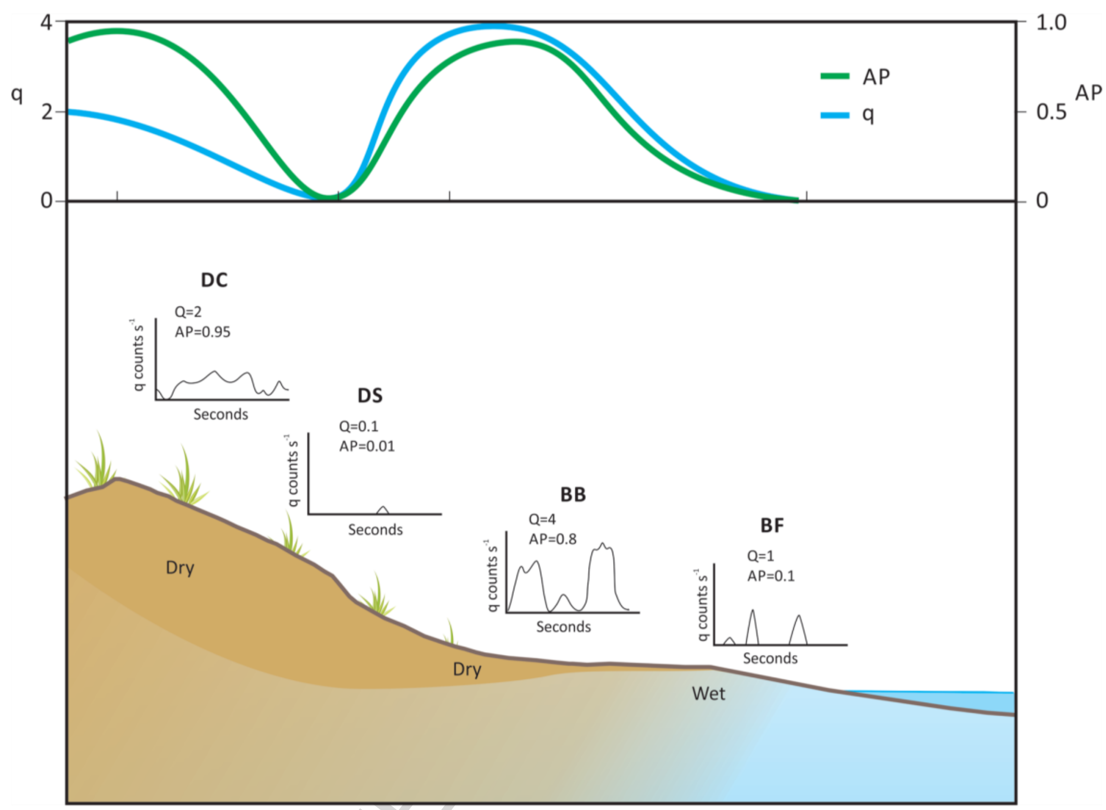
4.2.4. Exploring wind unsteadiness, transport response, and intermittency

Natural winds tend to be unsteady rather than constant, adding another level of complexity to sediment transport processes at the plot scale. Rather than a constant state of maximum flux, there is a semi-continuous state of disequilibrium between the time-varying nature of the wind and the phase-lagged response of the saltation system (Butterfield, 1991, Spies et al., 2000). This disequilibrium is most pronounced when wind speed fluctuates above and below the entrainment threshold, leading to discontinuous and constantly varying rates of sand transport. Stout and Zobeck (1997) proposed an 'intermittency' parameter that characterizes the degree of transport continuity as a function of the number of data points in a measurement period during which active transport occurs, expressed as a fraction of the total number of data points in the period. A time series with continuous transport, during which wind speed is consistently above the entrainment threshold, will have an intermittency value of 1, whereas a value of 0 indicates no transport. Davidson-Arnott et al. (2012) recommended adoption of an 'activity parameter' (AP) rather than 'intermittency parameter' (IP) per Stout and Zobeck (1997) because a large value for AP (or IP) indicates a very active transport system with minimal intermittency (taken literally). It should be noted, that while the AP (or IP) is a fairly simple concept, differences in the sampling effectiveness of different sensors make comparison of AP values derived between sensors and studies difficult (Baas, 2005; Davidson-Arnott et al. 2009; Barchyn and Hugenholtz, 2010).

Field experiments at the Greenwich Dunes in PEI and elsewhere have shown that the sand transport rate increases downwind from the limit of swash run-up toward the upper beach (e.g., Nordstrom and Jackson, 1992; Bauer and Davidson-Arnott, 2003; Bauer et al., 2009; Delgado-Fernandez, 2010; De Vries et al., 2014). A pattern of exponential increase in sand transport rate with downwind distance is often evident in time-averaged measurements along transects over beaches using depth-integrating traps (e.g., Davidson-Arnott and Law, 1990; Davidson-Arnott et al., 2008). Consistent with the fetch effect, this suggests that there is an increase in transport toward the saturated flux condition somewhere on the upper beach. However, observations also indicate that there can be considerable variation across the beach-dune profile, with some locations showing semi-continuous transport while others show significant transport intermittency. This suggests that the increase in the time-averaged transport rate at different positions across the beach-dune profile reflects both an increase in the instantaneous transport rate (larger flux peaks) and an increase in the proportion of time that transport occurs (Davidson–Arnott and Bauer, 2009).

There is also a positively reinforcing interaction between the fetch effect and the spatial pattern of surface moisture that leads to an increase in overall sand transport rate downwind toward the foredune. Figure 13 is a conceptual schematic of expected transport variation and AP values across a beach-dune profile for an obliquely onshore wind, based on observations at the PEI site. Sand transport on the beach foreshore (BF) is very intermittent, producing small AP values and small total transport (q_s). Toward the beach backshore (BB), the effects of the saltation cascade and decreasing moisture content produce an increase in activity and total transport. Behind the beach is a near vertical scarp that, coupled with the presence of vegetation, prevents most sand transported across the beach from reaching the foredune. As a consequence, both AP and q_s are very small at the dune toe and lower stoss slope (DS). Near the foredune crest (DC), sand is entrained from the upper stoss slope reflecting both small values for surface moisture and significant wind speed up towards the crest due to flow compression. This results in nearly continuous transport (large AP) at the crest but smaller values of sand transport in comparison to the back-beach.

Figure 13: Conceptual model of expected transport variation (q_s) and activity (AP) across a beach-dune profile at beach foreshore (BF), back beach (BB), dune slope (DS) and dune crest (DC) locations for an obliquely onshore wind, based on our field experiments at the Greenwich Dunes.



The consequence of these spatial-temporal controls on aeolian sand transport across the beach-dune profile (including fetch length, moisture interactions, vegetation, and topographic effects) is that there is often very poor correspondence between quasi-instantaneous (i.e., 1 Hz) wind speed and sand transport at any given location (e.g., Davidson-Arnott et al., 2008; Davidson-Arnott and Bauer, 2009). Indeed, regressions between transport flux and the cube of wind speed often have very small R^2 values, suggesting poor explanatory power (see example below in Fig. 15). The relationship usually improves with longer averaging intervals, which is consistent with the observation of Namikas et al. (2003) regarding u_*. Stout and Zobeck (1997) sought to use the observed fluctuations in wind speed and sand transport to derive a 'time fraction equivalent' threshold wind speed. However, measurements at the PEI site by Davidson-Arnott et al. (2005; 2008) and Davidson-Arnott and Bauer (2009) as well as others (e.g., Wiggs et al., 2004a, b) have shown that sand transport can occur when quasi-instantaneous wind measurements are below the calculated threshold of motion, and vice versa. In part, this is explained by the phase-lagged response of saltation to changes in wind speed (Spies et al., 2002), but there are also spatial dimensions involving the delivery of saltating sediments to a sensor location from upwind sources that have differing surface controls and wind patterns.

4.2.5. New observations of vertical sediment flux variations and transport events (flurries)

Aeolian sand transport is a near-surface phenomenon in that saltation layers are of limited vertical extent. The bulk of transport occurs in a very thin layer immediately above the surface by grains moving in saltation (saltons) and as surface creep (reptons). The grain concentration in a given volume of air decreases with increasing distance from the surface in a non-linear manner, as does the transport rate. Usually an exponential-decay function is used to describe vertical profiles of sand transport (Ellis et al., 2011; Rotnicka, 2013; Bauer and Davidson-Arnott, 2014). Energetic saltons that rise higher into the air column tend to have larger particle speeds than low-energy saltons that are constrained to a near-surface layer and, therefore, the concentration profile and the flux profile are not ordinarily interchangeable unless information is available on the particle speed profile. Another source of confusion arises from the use of three different transport quantities (mass, volume, particle count) to represent the vertical profile. These are, in theory, interchangeable but in practice there can be insurmountable challenges and uncertainty around grain size distributions, particle shapes, and mineral densities. There remains considerable debate in the literature regarding whether the vertical profile of sediment flux is smoothly continuous or layered (e.g., Butterfield, 1999; Dong et al., 2006; 2011; Farrell et al., 2012) and how the profile should be parameterized (e.g., Martin et al., 2013; Bauer and Davidson-Arnott, 2014).

Insight into the nature of vertical mass flux profiles across beach-dune systems was facilitated with the adoption of the segmented sand trap in field studies (e.g., Williams, 1964; Rasmussen et al., 1989; Rasmussen and Mikkelson, 1998; Sherman et al., 1994; 2014; Namikas, 2003). The cumbersome nature of these first-generation traps was eliminated by the development of smaller and more sophisticated laser particle counters (LPCs) and acoustic sensors, which could be stacked vertically. Some key advantages of the LPCs used in the PEI experiments are that they are commercially available, relatively affordable, and manufactured to high technical standards. This implies that the results from one unit are precisely reproducible by another unit, eliminating the need for extensive cross-calibration (cf. Baas, 2004 in regards to Safire-style piezoelectric probes). As with any field instrument, LPCs have certain shortcomings (see Barchyn et al., 2014 and references therein), the most challenging of which is the conversion of particle counts to mass flux. It is advisable to co-locate a passive, segmented trap such as a multi-layered ‘hose trap' (Sherman et al., 2014) to verify results from the high-frequency sensors with direct mass flux measurements.

In the PEI research, vertical arrays of LPCs yielded novel insights into the nature of aeolian saltation at the plot scale. Figure 14 shows time series of: (A) wind speed; (B) wind angle; (C) particle flux; and (D) AP during an intense wind event on 4 May 2010 at the PEI field site (from Bauer and Davidson-Arnott, 2014). The wind speed and particle flux traces suggest a crude correspondence for which the most intense and variable speed segments align with the greatest flux events. However, a simple regression analysis using the 1 Hz data (Fig. 15) reveals that the R^2 is only 0.33 (P <0.0001), which is typical for raw, high-frequency data that have not been averaged. There are periods near the beginning of the time series when transport was not very active and a large number of intervals that had no transport whatsoever. Figure 14D shows APs for different layers in the vertical flux profile calculated over 15-minute intervals. The first (red) bar in each interval shows AP for the lowermost LPC (0.014 m), and each bar progressively declining to the right shows LPCs higher in the profile (up to 0.472 m). The lowermost LPC in the first interval had an AP of only 0.37 followed by 0.75 for the second interval, and 0.99 for the third interval. All subsequent intervals had APs in excess of 0.91 for the lowermost LPC (in most cases it was 0.99 or 1.00), indicating a very energetic transport system over a prolonged period. Of particular interest is the substantial difference in the nature of the vertical flux profiles before 1545 h relative to those afterward. In the earlier intervals, there was a rapid reduction in AP above the lowermost LPC with values typically < 0.2, which means that the majority of particle flux was contained in a near-surface layer of approximately 0.05 m height. In contrast, after 1545 h when the total transport rate increased, the AP of the mid-level LPCs was typically > 0.2 and often as large as 0.7, which indicates that there were significantly more energetic saltons higher in the profile. The uppermost LPC (at 0.472 m) had APs between 0.05 and 0.15, whereas in the earlier period there were very few saltons recorded at this height. A detailed assessment of these flux profiles was undertaken by Bauer and Davidson-Arnott (2014), wherein it was demonstrated that the geometry of the vertical flux profiles (shape, slope) depended on the event-like nature of the sand transport time series. Specifically, during intervals when transport was highly intermittent (small AP), there were fewer significant transport events (referred to as sediment 'flurries') interspersed between longer periods of quiescence. In addition, these flurries tended to have shorter life-spans (several seconds), which means that the saltation system rarely achieved the equilibrium (saturated) transport state. Thus, there are intricate linkages between wind unsteadiness, transport intermittency, and the geometry of the vertical flux profile and these linkages are further complicated by topographic position and vegetation characteristics over the beach-dune profile.

Figure 14: Time series of: (A) wind speed; (B) wind approach angle; (C) particle flux; and (D) Activity Parameter (AP, a measure of transport intermittency) recorded at 1 Hz for a three-hour measurement period on the foredune crest on 4 May 2010. Wind speed and direction are from the 3D sonic at 0.2 m above the bed adjacent to the vertical array of LPCs. Smoothed trend lines are 5-minute moving averages. Particle flux is the vertically-integrated instantaneous (1 Hz) count summed over six LCPs in the vertical array (left-hand scale). Upper (grey-shaded) panels show 15-minute mean counts (right-hand scale). Every 15-minute segment has six vertical bars that indicates AP for each of the six sensors in the vertical array (left-most bar is the lowest LPC and right-most bar is the highest LPC in the array).

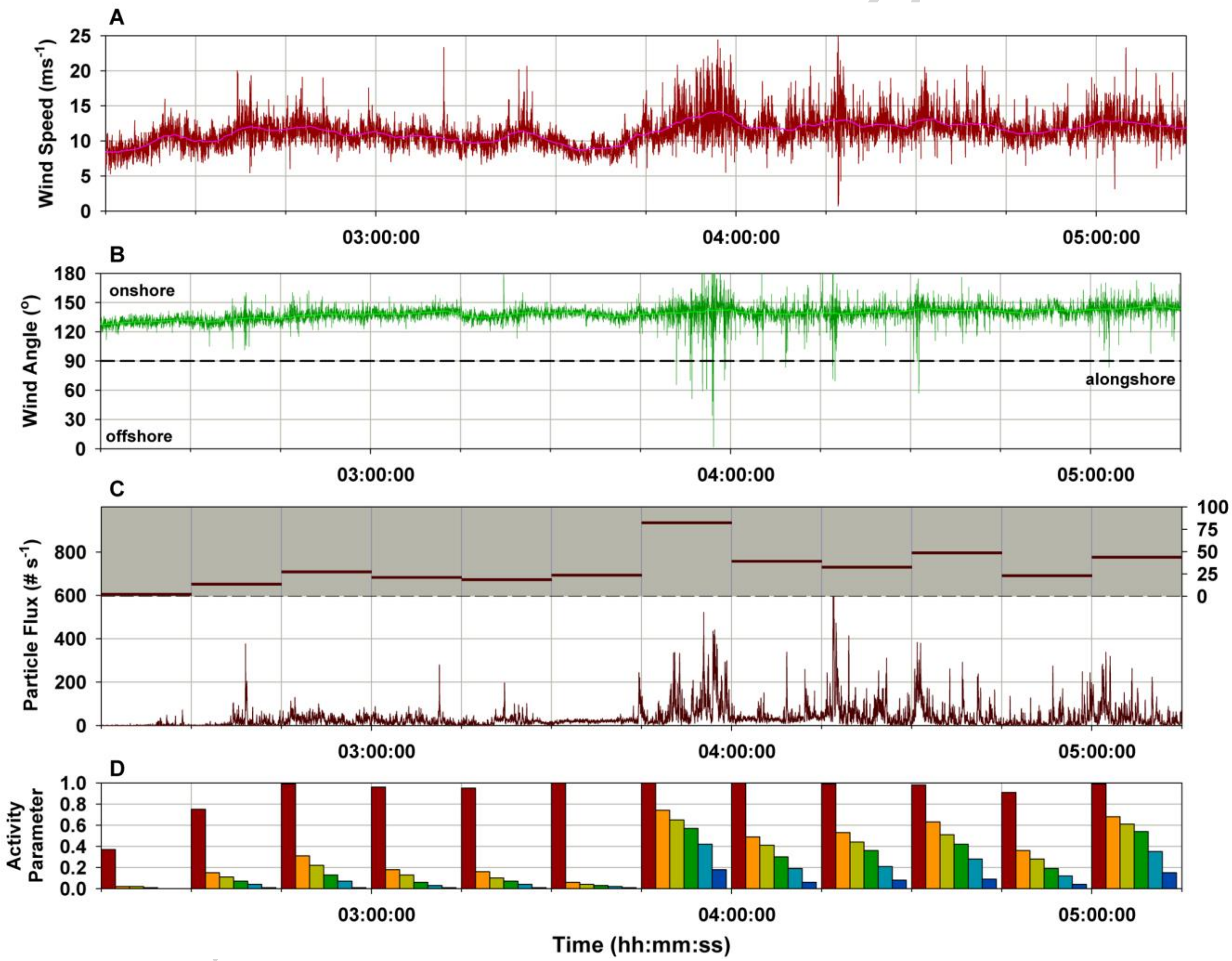
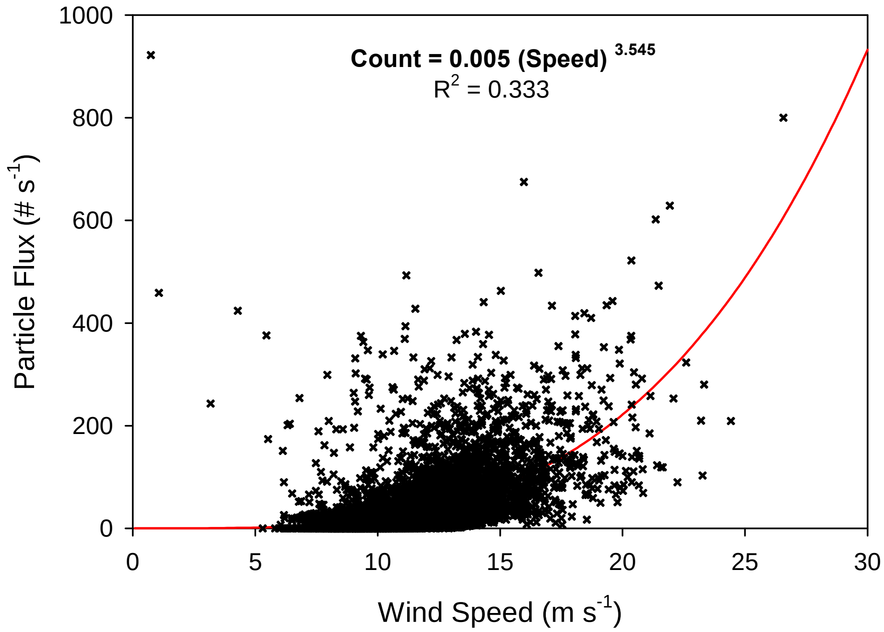


Figure 15: Regression of wind speed against particle flux (data in panels A and C, respectively, in Fig. 14) during an intense wind event on 4 May 2010 at the PEI field site.



4.2.6. Observations of flux divergence and spatial-temporal patterns of erosion and deposition

The introduction and use of relatively affordable, fast-response sediment transport sensors in aeolian geomorphology has facilitated the deployment of dense arrays of instruments that enable the characterization of spatial-temporal patterns of transport rate across an entire beach-dune profile. Not only has this provided insight into the vertical structure of the saltation layer, as described above, but also into the potential correlation between fundamental scales of fluid events (i.e., coherent flow structures) and transport events such as aeolian streamers (Baas and Sherman, 2005; Bauer et al., 2013) or 'flurries' (Bauer and Davidson-Arnott, 2014).

Initially, the objective of horizontal arrays of transport sensors was to quantify the spatial variability in transport rate relative to predictions from equilibrium models. Ellis et al. (2012)

noted that there were only five field-based studies addressing this problem at that time, and four of them used integrating traps rather than fast-response sensors. The horizontal spacing between traps was usually several metres. Baas (2003) was the first to utilize a very closely spaced horizontal instrument array, which included both fast-response piezo-electric impact sensors (e.g., 'Safires') and hot-wire anemometry. Collectively, these studies demonstrated that there can be considerable spatial variability in transport rate, with the coefficient of variation across the array of traps ranging from about 0.1 to 1.0. In the Baas (2003) study, the sand surface in front of the array was meticulously groomed, thereby reducing the likelihood that the spatial variation in transport rate was due to surface controls. Nevertheless, it proved impossible to link the scales of the transport events (i.e., streamers) to the scales of fluid structures embedded in the wind field in a statistically reliable way.

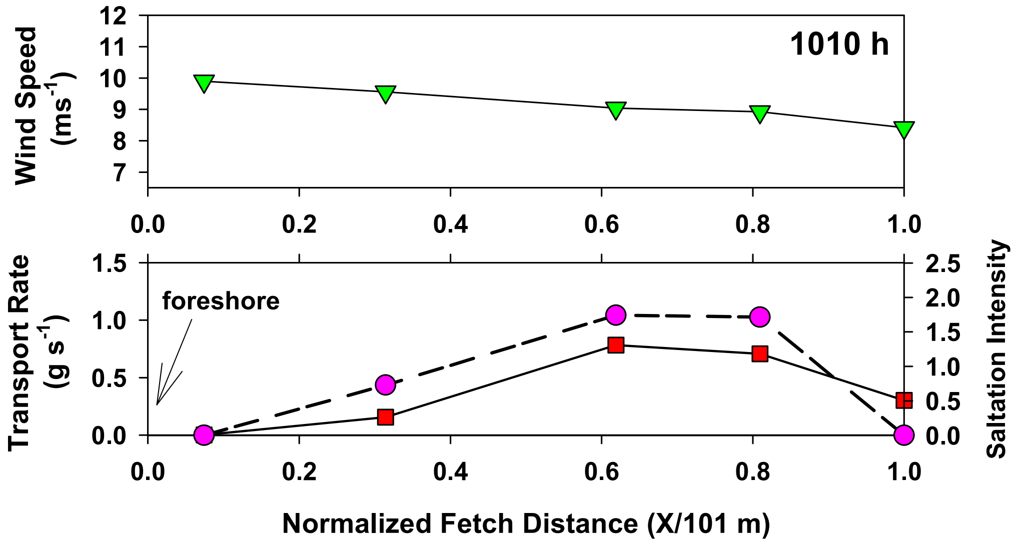
One of the more useful applications for data derived from spatial arrays of LPCs is to derive the sediment flux divergence, $\nabla \cdot q_s$, which is the spatial gradient (d/dx, d/dy, d/dz) in sediment volume flux (q_s). The flux divergence is used in a simplified version of the sediment conservation relation referred to as the Exner equation (Paola and Voller, 2005),

$$\frac{\partial h}{\partial t} = -\frac{1}{(1-p)}\ \nabla \cdot q_s \tag{1}$$

where h is elevation of the bed, t is time, and p is sediment porosity. Figure 16 shows the cross-beach pattern of mean wind speed and mean sediment transport (time averaged flux and transport intensity) at four trap locations oriented along prevailing streamlines during an obliquely onshore wind event at the Greenwich Dunes site on 11 October 2004. Estimates of sediment transport were from integrating traps as well as co-located Safire sensors and both methods showed the same trend in transport. The rate of sand transport increased from a minimum at the foreshore, where conditions were extremely wet and fetch-limited, to a maximum on the upper beach, where conditions were dry and closer to equilibrium. An unexpected, but recurring decline in transport rate toward the dune toe was also measured, which is explained by the downstream reduction in wind speed associated with the vertical growth of the boundary layer, and flow stagnation imposed by the dune, thereby yielding a concomitant decrease in near-surface shear stress (Bauer et al., 2009; Walker and Hesp, 2013; Hesp et al. 2015). The flux divergence between neighboring trap locations suggests that there would be net erosion from the foreshore and across most of the beach, which is required to drive the increase in saltation flux in the downwind direction. However, the decrease in sediment flux between the last two stations indicates that this is a zone of deposition, which is typically observed during onshore transport events across beaches.

Bauer et al. (2015) demonstrated that a similar pattern of sand accumulation at the toe of the dune occurs during offshore wind events because of eddy recirculation over the seaward (lee) slope of the foredune. A methodology for isolating the cross-shore sediment flux from the total sediment flux using the wind vectors was proposed. These patterns of flux divergence over beaches are critical to infilling wave-cut scarps at the dune toe and to rebuilding dune ramps that are essential to facilitating sand transport pathways onto the stoss slope of the foredune and toward the crest.

Figure 16: Patterns of mean wind speed, mean sediment transport rate derived from sand traps (squares), and transport intensity measured from safire sand transport probes (circles) at several trap locations from the foreshore to dune toe during an obliquely onshore wind event at the Greenwich Dunes, PEI site on 11 October 2004. Normalized fetch distance for each trap (aligned into local flow streamlines) is provided on the x-axis.



The complexity of spatial-temporal patterns of erosion and deposition across beach-dune systems during single events is becoming widely appreciated and increasingly quantified, yet the linkages between plot scale investigations and landform scale perspectives remain more challenging. Logistically, it is not yet feasible to conduct experiments at the intensity of the plot scale with continuous high-frequency monitoring over periods of years. Nor is it reasonable to maintain high-density instrument deployments over very large areas because of financial constraints. As a consequence, there can be substantial data gaps during periods in which significant geomorphic change may occur in locations where we did not (or were unable to) monitor. Thus, plot scale studies provide only a limited vignette within the broader frequency-magnitude-effectiveness regime that governs dune morphodynamics, and yet it is the broader landscape scale perspective that is of greater concern to coastal resource management. The next section explores research at the landform scale that attempts to bridge the divide between the plot scale and the landform scale (Table 1).

4. Landform scale

The objectives of research at the landform scale in the PEI study were motivated by the need to make observations and to obtain data that provide insights into which processes at the plot scale may be most relevant for understanding and managing issues related to foredune morphodynamics in partnership with Parks Canada Agency (e.g., dune evolution, dune migration, coastal erosion). At the landform scale (Table 1), beach-dune sediment budgets and foredune growth over months to years are controlled initially by the volume of sand on the beach that is available to be transported to the foredune by aeolian processes and/or the propensity for sediment to be eroded from the upper beach and foredune by high-water events.

Much effort has gone into developing predictive models based on standard deterministic equations used in plot scale studies (e.g., Hunter et al., 1983; Kroon and Hoekstra, 1990; Wahid, 2008). The approach employed in the PEI research at the landform scale is similarly 'reductionist' (Bauer and Sherman, 1999) as it splits the problem of predicting aeolian transport into smaller and smaller components with the intent of scaling back up. Over time, it has certainly offered insights into the relationships between aeolian sediment transport and a range of controlling variables, but it has also highlighted other non-trivial issues such as how to combine multiple supply-limiting factors, or to upscale findings from plot scale studies to the landform scale. Ultimately, a main focus of research at the landform scale is to predict (model) sand delivery from the beach to the foredune, and then to examine the broader processes involved in beach-dune interaction as controls on foredune evolution.

4.1 Modeling sediment delivery to coastal dunes

4.1.1 Classic approaches to modelling long-term aeolian sand drift

The most widely used approach for predicting aeolian sand supply and resulting dune form is the 'Fryberger method' (Fryberger and Dean, 1979), which uses the Lettau and Lettau (1977) equation to calculate aeolian sand transport (drift) at an annual scale. The Fryberger method was applied initially to desert dunes (e.g., Fryberger, 1980; Carson and Mclean, 1986; Wang et al., 2002) but has also been adopted for coastal dunes (e.g., Chapman 1990; Wal and McManus, 1993; Davidson-Arnott and Law, 1996; Hesp and Hyde, 1996; Blumberg and Greeley, 1996, Walker and Barrie, 2006; Miot da Silva and Hesp, 2010). The wind speed at 10 m drives ‘drift potentials’ (DP) in compass directional classes to express total potential sand drift and resultant drift potential (RDP) associated with the wind regime in a particular area. In turn, these quantities can be related to dune size, shape, and mobility using statistical expressions. The method is relatively simple and only requires wind data from standard meteorological stations. Details of the method and discussion of procedural limitations, including inaccuracies resulting from how data are converted (e.g., units as knots vs. m s^{-1}, Bullard, 1997) and/or categorized during the calculations to introduce ‘frequency bias’ (Pearce and Walker, 2005), are provided elsewhere.

The main limitation of the Fryberger method, especially in vegetated coastal foredune settings, is that it does not account for key supply-limiting factors, such as surface moisture or fetch effects (Nickling and Davidson-Arnott, 1990; Bauer and Davidson-Arnott, 2003), transport –limiting factors such as vegetation or beach wrack, or near-surface secondary flow effects such as topographic steering. As a result, measured transport and deposition is typically much less than predicted by the Fryberger method (e.g., Hunter et al., 1983; Sarre, 1989; Chapman, 1990; Davidson-Arnott and Law, 1996; Hesp and Hyde, 1996).

4.1.2 New efforts to assess the regime of aeolian transport events in beach-dune systems

Recognizing the limitations of the Fryberger approach to predicting sand supply from the beach to the foredune, Delgado-Fernandez et al. (2009; 2012; 2013a) developed a monitoring system that simultaneously measured wind velocity (hourly), sediment transport, and some key supply-limiting factors including surficial moisture content (see Darke et al., 2009) and beach width (cf. Lynch et al., 2006). Sediment transport was measured using several complementary methods, including piezoelectric (‘Safire’ style) saltation sensors and erosion-deposition pins permanently deployed from the upper beach toward the crest, bedframe volumetric surveys, and visual interpretation of aeolian transport conditions from fixed-mount camera imagery. The resulting data set permitted assessment of both the magnitude and frequency of wind events and associated sand transport events, as well as the development and testing of a modelling approach to predict sand supply to the foredune that accounts for supply-limited conditions operating at seasonal time scales (Delgado-Fernandez and Davidson-Arnott, 2011).

To develop models capable of calculating sand supply to a foredune and, in turn, to better predict foredune evolution at the landform scale, it is important to evaluate the relative significance of event frequency, magnitude, and effectiveness following the concepts described in Wolman and Miller (1960) and Wolman and Gerson (1978)(see section 2). An effective wind event is defined as a period during which wind speed exceeds the threshold of motion for dry sand for more than two hours (based on an hourly photo acquisition rate), thus providing the potential for significant transport to occur. Additional supply-limiting factors to consider are the moisture state of the beach sand, wind approach angle, and available fetch distance, among others. Ultimately, the potential hourly sand transport rate (Q) can be converted into a potential hourly sand delivery into the dune based on the cosine function (Davidson-Arnott and Law, 1990; Bauer and Davidson-Arnott, 2003):

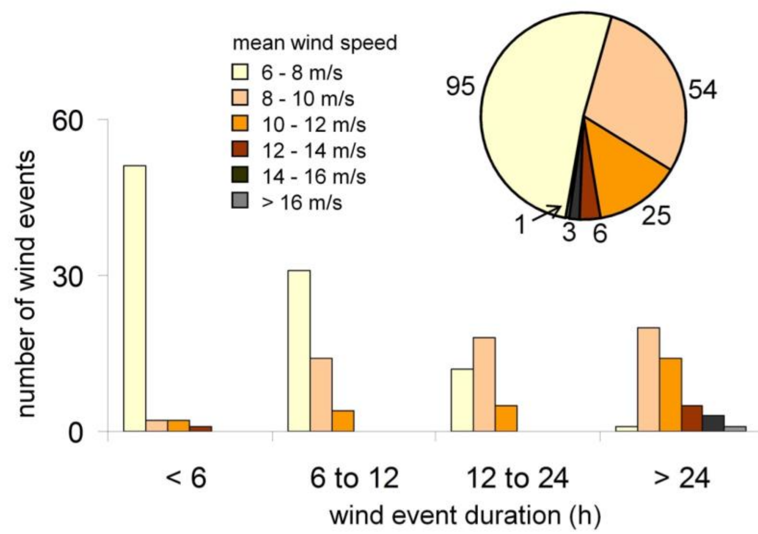
$$Q_n = Q \cos \alpha \tag{2}$$

where α is the angle of the wind to shore perpendicular and Q_n is the hourly average sand transport into the foredune per metre alongshore (kg h^{-1} m^{-1}). Equation 2 can be summed for each hour to give the total potential transport for the event. Note that a transport event does not necessarily coincide with the duration of the wind event because of the threshold condition, which defines when sand transport is active. As such, transport events may occupy all or only a portion of an associated wind event and some wind events may have no transport associated with them at all.

Delgado-Fernandez and Davidson-Arnott (2011) examined a total of 184 wind events during a 9-month period from 1 September 2007 to 31 May 2008. Most of these events

(95/184) had mean wind speeds of < 8 m s^{-1} and, therefore, were of insufficient strength to yield transport (Figure 17). Only about 25% of the events had wind speeds in excess of 12 m s^{-1} and of sufficient duration (> 12 hrs) to yield significant sediment transport.

Figure 17: Wind event categorization according to wind speed magnitude and storm duration (adapted from Delgado-Fernandez and Davidson-Arnott, 2011). In general, low magnitude events were more frequent and of shorter duration than large magnitude events, which were infrequent and of longer duration.



The relative magnitude of each wind storm ($Q_{m\%}$) as a potential sediment transporting event was calculated based on hourly wind transport rates and event duration, and expressed as percentage of the total amount of sediment transport predicted for the study period. An expression proposed by Wolman and Miller (1960) was adapted for this purpose:

$$Q_{m\%} = \frac{Q_i \, . \, F}{Q_{tot}} \, . \, 100 \qquad (3)$$

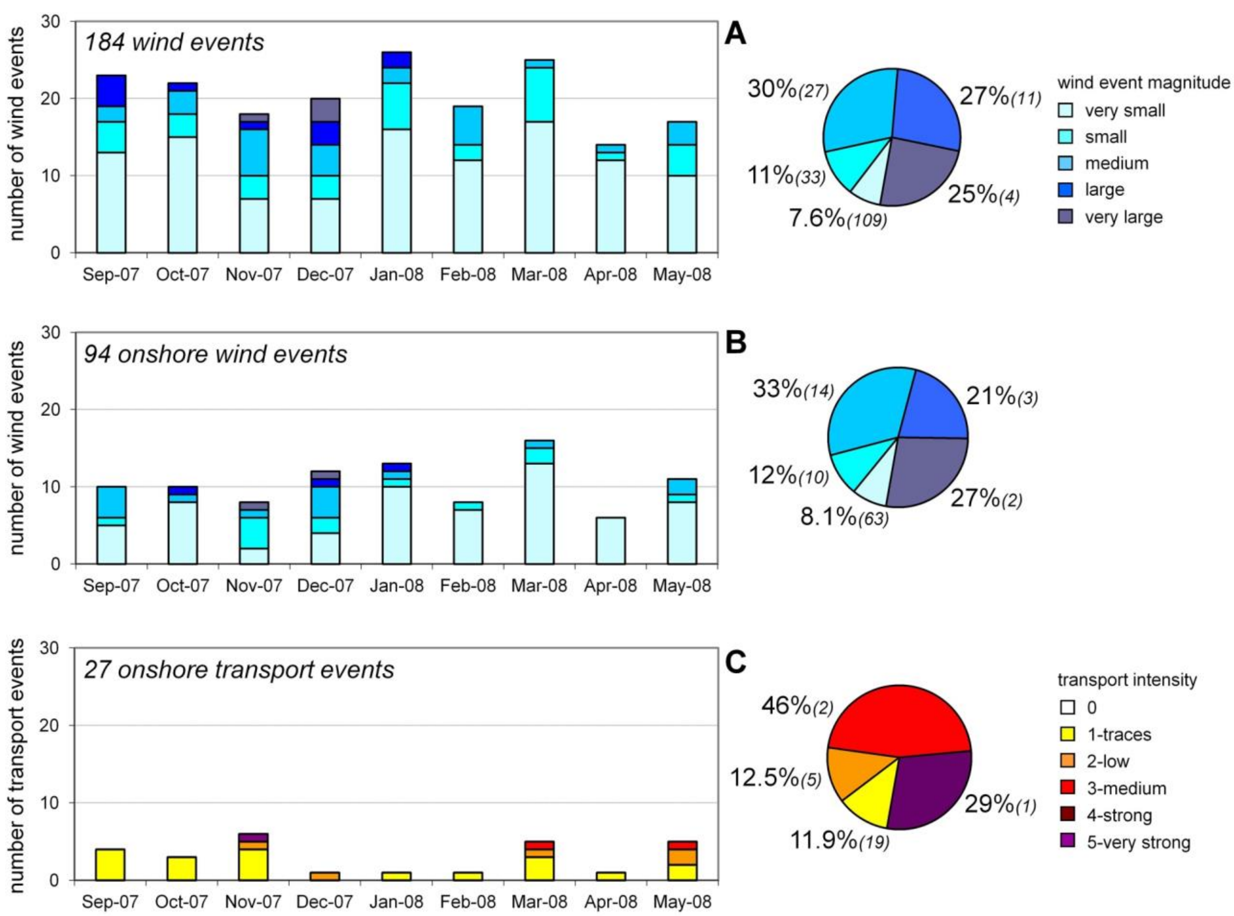
where Q_i is the sediment transport during a given event predicted by summing the potential transport for each hour (per Delgado-Fernandez and Davidson-Arnott, 2011), F is the frequency of the event, and Q_{tot} is the total sand transport predicted over the study period. Events were grouped into five magnitude classes prior to implementation of Equation 3 to simplify the frequency analysis. Roughly 50% of the potential transport was associated with large magnitude events (> 81 x 10^2 kg m^{-1}), while the smallest events (< 3 x 10^2 kg m^{-1}) contributed only 7.6% to the total potential transport despite accounting for more than 60% of the total number of events (Fig. 18A).

Large magnitude wind events occurred mostly during the late fall and winter months. Events with an onshore component accounted for about half of the number of events but about 71% of the total potential transport. However, when the potential transport is modified by the cosine function (Eq. 2), the net potential transport (Q_n) into the dune is about 41% of the total transport predicted for all events (Fig. 18B). Despite the overall reduction in the number of events with only onshore conditions and the magnitude of predicted transport for those events, the percentage distributions associated with each category were very similar to the total population of all wind events (Fig. 18A). Infrequent, large magnitude wind events were still concentrated during the late fall and winter months and were responsible for approximately 50% of potential sediment input to the dunes (as depicted in Figs. 2c, d).

Many of the wind events with very large transport potential (i.e., with extreme wind speeds) actually produced less (or no) total transport compared to more moderate events (Fig. 18). The influences of one or more supply-limiting controls, such as fetch, wind angle, surface moisture, storm surge, wave runup, tide level, and the presence of snow and ice during the winter months (January through March) are critical in determining whether aeolian transport is active or not and, thereby, effective in moving sediment into the foredune. The wind vector is but one of many important variables to consider at this scale, which is a much different result than that obtained using the Fryberger method in the absence of appropriate local controls on the transport process.

Figure 18C shows that three transport events were responsible for the majority of sand delivery to the foredune over the 9-month observation period. The largest amount of transport occurred during an event that lasted 90 hours (long duration) with an average wind speed of 12.5 m s^{-1} (moderate to low magnitude). The other two transport events occurred with average wind speeds of 8.2 and 9.2 m s^{-1} (low magnitude) and lasted 32 and 54 h (long duration), respectively. Active transport during these events was both time limited (i.e., only observed during a portion of the wind event) and magnitude limited (i.e., observed transport was less than predicted). Despite fetch-restricted and moisture-limiting conditions, these three events delivered 75% of the total amount of sand to the dune during the study period. The remaining 25% was delivered during 24 lesser transport events. Ten of the strongest wind events, which accounted for 45% of the total predicted sediment input to the dunes, produced no significant transport at all (Delgado-Fernandez and Davidson-Arnott, 2011). This study shows that although transport-competent winds may be frequent, only a sub-set of these events may be effective in transporting sediment toward the foredune given the complex suite of supply-limiting factors and their seasonal variations.

Figure 18: (A) Frequency distribution of potential sediment transport events for the 9-month study period. A total of 15 large or very large magnitude wind events during winter months were associated with over 50% of the total potential sand transport. (B) Frequency distribution of potential sand transport events with onshore flow direction and modified by the cosine function, which are believed to be the major contributors to foredune maintenance and growth. Large magnitude events still accounted for approximately 50% of potential sediment input to the dunes. (C) Observed sand transport towards the foredune measured using a combination of techniques (described in Delgado-Fernandez and Davidson-Arnott, 2011). Only one of the original large magnitude wind events (in November) actually produced strong transport. Two additional medium transport events occurred in the spring. Together, these three events accounted for approximately 75% of the total sand delivered to the dunes. Values outside the pie charts indicate the percent of potential transport and the number (in brackets) of events.



4.1.3 Advances in modelling the effect of supply-limited conditions on predicted sand transport to foredunes

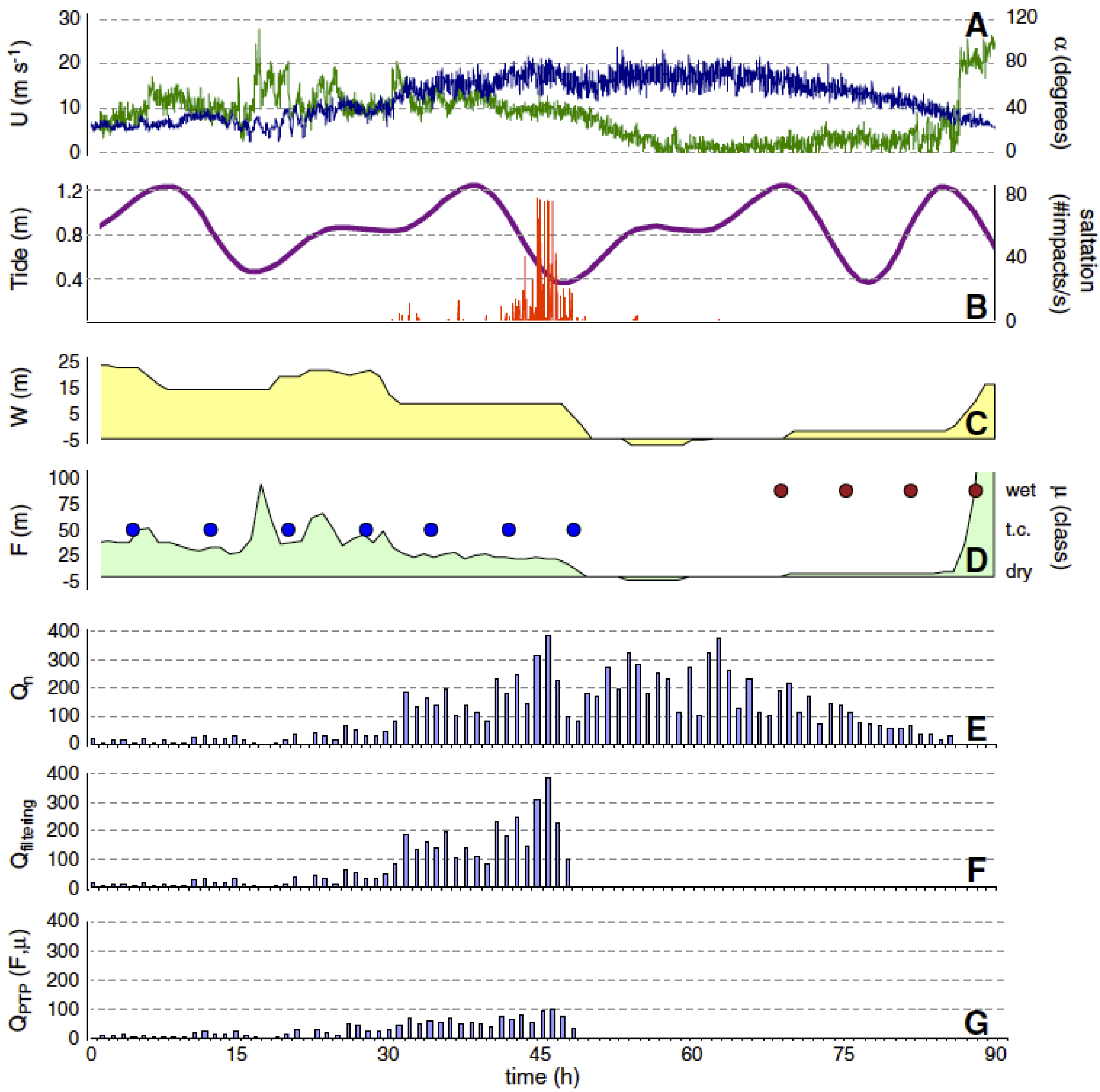
As discussed above, supply limitations play a key role in determining the actual sediment transport associated with a wind event in coastal environments. This highlights a need to model supply limitations explicitly when predicting sediment supply to foredunes over periods of weeks to years. Delgado-Fernandez (2011) used the same dataset to test a supply-limited modelling approach, which involved filtering the time series to remove all periods when: i) wind speed was below the threshold for dry sand, ii) when winds were offshore, and iii) during periods of high surface moisture or coverage by snow and/or ice. Using the theoretical framework for assessing the impact of the fetch effect (per Bauer and Davidson-Arnott, 2003, sections 4.2.2 and 4.2.3) the critical fetch length, F_c, was first determined for 'dry' conditions, where the surface moisture content was <2%, and then for situations of greater moisture content, between 4 and 10%, to allow for the lengthening of F_c with increasing surface moisture. When $F_c > F$, the effects of supply limitation can be modeled by any of the expressions presented in Bauer and Davidson-Arnott (2003), whereas when $F_c < F$, transport rate is considered to be at its maximum potential.

An example of the model output for a 90-hour storm in November 2007 is shown in Figure 19. Sediment input to the foredune based on wind speed and direction alone was over-predicted at Q_n = 9,470 kg m^{-1} for this event, with maximum transport rate coinciding with the peak of strong onshore winds (Fig. 19A,E). Large moisture content and short fetch distances (Fig. 19C,D) imposed a constraint on sediment transport at around 50 hrs into the event and, when these factors were included in the model (Fig. 19G), the predictions improved considerably. The initial filtering approach reduced the total predicted input to the foredune from approximately 86,000 to 36,000 kg m^{-1}. The value was further reduced to about 19,000 kg m^{-1} once the supply-limiting effects of fetch and moisture were applied. Values for deposition measured by erosion-deposition pins and bedframe stations over the same period ranged from about 4,000 to 15,000 kg m^{-1}. The uncertainty in measured deposition reflects the difficulty in accounting for the effects of wave erosion and some landward sand transfers (losses) beyond the foredune, although it is evident that the filtered estimates from the model are much closer to the measurements than to original, unfiltered predicted values.

These modelling results at the landform scale, combined with the plot-scale investigations described above, highlight the need to include supply-limiting factors when predicting sand transport from the beach to the foredune. The modeling approach reviewed here considers the effect of increasing moisture content on lengthening the critical fetch (F_c) necessary to achieve saturated transport, however, it is also possible to model this effect as a limitation on sand supply from the surface directly (cf., de Vries et al., 2014). Other studies have attempted to scale up sediment supply to coastal dunes and model their evolution by simply calculating subaerial barrier volumes and comparing these to foredune volumetric change measurements (e.g., Miot da Silva and Hesp, 2010), or by using computational approaches that solve simplified aerodynamics and sand transport relations (e.g., Luna et al., 2011; Duran and Moore, 2014) or cellular automata approaches (e.g., Baas, 2002; Baas and Neild, 2007; 2010; Zhang et al., 2015; Keijsers et al., 2016). The utility of these modelling efforts is limited, however, by fundamental assumptions of saturated flux and the effectiveness of onshore winds over seasons and years. While incorporation of complexities such as moisture and vegetation may improve some simulations (e.g., Luna et al., 2011, Zhang et al., 2015; Keijsers et al., 2016), realistic parameterization of vegetation growth (e.g., seasonal phenology, gradual succession) and related roughness effects and sand trapping efficiency are generally lacking. In addition, there are limited field measurements to inform and validate such models, which increases the risk of using expedient oversimplifications (Barchyn et al., 2014).

Finally, sand input from the beach is just one component of the foredune sediment budget. Controls on dune evolution at the landscape scale must also consider the broader framework of beach-dune interaction, which includes wave erosion during storms and berm construction and the onshore welding of nearshore sand bars to the foreshore, as discussed below.

Figure 19: Modelling output for a 90-hr storm at the Greenwich Dunes, PEI site starting on 9 November 2007. A) 2-min records of wind speed, U, and direction, α; B) saltation intensity and tidal elevation; C) beach width, W; D) fetch distance, F, determined by beach width and wind direction, and classified (optically derived) moisture values, μ; E) hourly potential transport based on wind speed and direction, Q_n; F) output of the filtering step, $Q_{filtering}$; G) calculated transport over isolated potential transport periods, Q_{PTP}, including fetch distance and moisture. Transport in E–F expressed in kg m^{-1}. Modified from Delgado-Fernandez (2011: Fig. 10).



5.2. Beach-dune interaction

5.2.1. Classic understanding of sand supply and coastal dune evolution

While cycles of foredune erosion during extreme storms and subsequent rebuilding by aeolian processes over long inter-storm periods have been recognized for decades, incorporation of this understanding into a holistic conceptual framework stems from the proceedings of a symposium on beach-dune interaction edited by Psuty (1988). Studies of beach-dune interaction typically employed either one or some combination of repeated topographic surveys, mapping from aerial photography, stratigraphic analysis from trenches or cores, or interpretation of shallow seismic logs (e.g., Olson, 1958; Bigarella, 1979; Thom and Hall, 1991; Gares and Nordstrom, 1995; Bristow et al., 2000; Hesp, 2013). In the last two decades, the development of Ground Penetrating Radar (GPR) and airborne or terrestrial LiDAR has enhanced mapping of landforms in great detail. In addition, analysis of digital imagery using GIS has greatly increased our ability to use both historical aerial photography and modern imagery (e.g., Figs. 11, 23) to map landform change at time scales of days to decades. These technologies provide a compelling means to fill the information gap between the plot scale and the landform scale. Nevertheless, a key challenge remains in correlating observed morphological changes of foredunes with the key forcing variables.

At the plot scale, localized erosion and deposition patterns can be understood and crudely predicted on the basis of the near-surface wind vector and the contributions of a host of supply-limiting factors listed as 'independent' variables in Table 1. But at the landform scale, all of the 'independent' variables at the plot scale become 'dependent' variables and, therefore, the patterns of erosion and deposition that ultimately lead to broad-scale foredune evolution are governed by such factors as the nature of shoreline progradation or erosion, the emergence or removal of vegetation cover, and seasonal to decadal changes in the morphodynamic state of the nearshore system fronting the foredune. Thus, the detailed nuances of sediment transport at the scale of seconds and hours (i.e., the intra-event dynamics of interest at the plot scale) become largely irrelevant as attention must shift toward event characterization (i.e., kinds of events), inter-event conditions (i.e., processes active between events), and the time-sequencing of events. In effect, the focus becomes parameterizing the changing nature of geomorphically effective events over time as conditioned by the broader context within which beach-dune interaction takes place.

Events leading to dune erosion and potential overwash are controlled by meteorological factors that govern wave generation, storm surge, and aeolian transport (e.g., Kriebel and Dean, 1985; Vellinga, 1986; Morton, 2002; Forbes et al., 2004; Thornton et al., 2007; Pye and Blott, 2008; Roelvink et al., 2009). An additional factor is alongshore variations in beach width associated with rip current circulation and megacusps (e.g., Komar, 1971; Thornton et al., 2007), intertidal bar welding (e.g., Aagaard et al., 2004; Anthony et al., 2006) and longshore sandwave migration (e.g., Inman, 1987; Davidson-Arnott and Stewart, 1988; Davidson-Arnott and Law, 1990; Ruessink and Jeuken, 2002; Davidson-Arnott and van Heyningen, 2003; Houser et al, 2008). These controls operate at time scales of months to years.

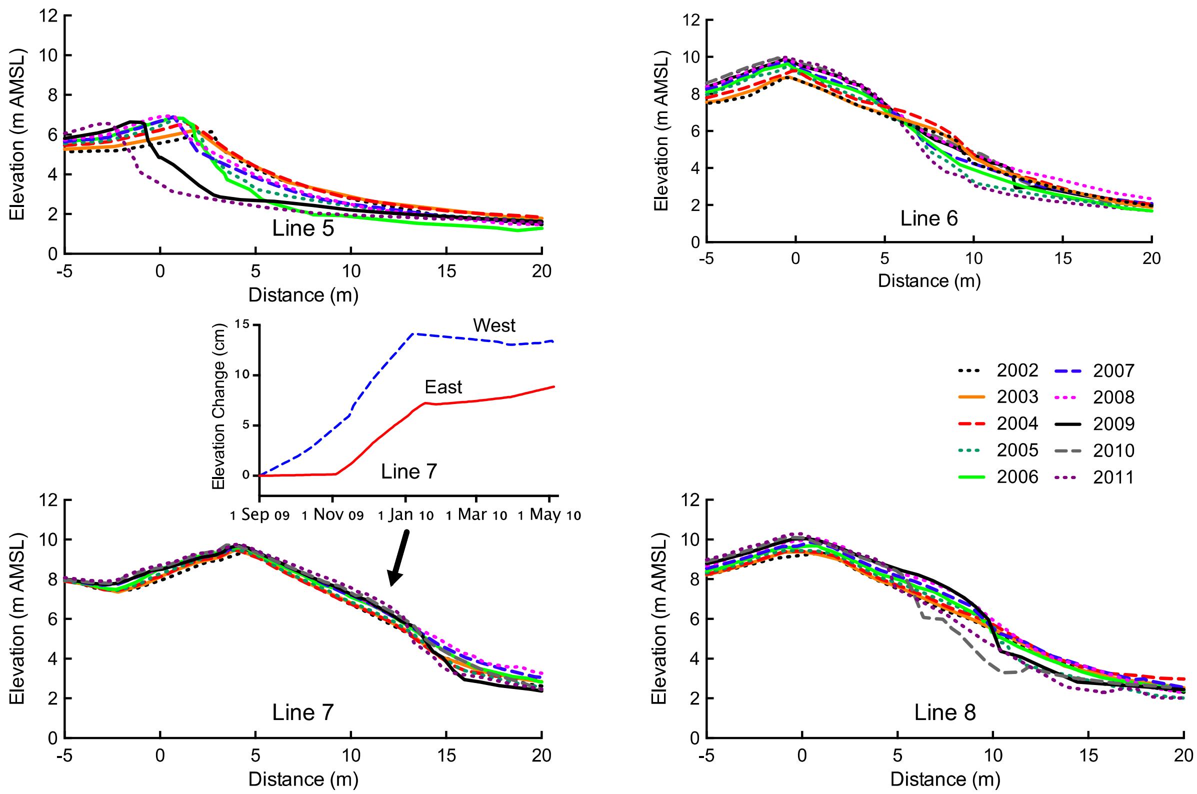
5.2.2. Assessing annual to decadal beach-dune interaction

Changes in beach-foredune morphology were quantified at the PEI site using: (i) repeat topographic surveys of cross-shore profiles, (ii) detailed bedframe measurements of volumetric changes along each transect (per Davidson-Arnott and Law, 1990), and (iii) hourly remotely sensed measurement of erosion-deposition pins on the stoss slope of the foredune (see Delgado-Fernandez et al., 2009; Ollerhead et al., 2013). The site was partitioned into 3 distinct reaches (see Fig. 1d). Reach 1 extended W about 2 km from the E boundary of PEI National Park and was oriented ~100°-280°. Reach 2 extended about 3 km to the inlet of St. Peters Bay and was oriented at 60°-240° and reach 3 is about 1 km long and extends SE into St. Peters Bay.

Examples of annual topographic profiles are shown in Figure 20 for Lines 5 to 8 in Reach 2 for the period May 2002 to May 2011. These data illustrate how profile response differs in the alongshore direction (E to W) due to variations in the littoral sediment budget fronting the beach-dune system. Sediment accretion is evident on the lee slope for all profiles in Reach 2, indicating landward sand transfers, but the pattern of topographic change on the stoss slope and back beach is highly variable. On the eastern margins of Reach 2 (e.g., Line 5), the profile was displaced landward over time due to a negative littoral sediment budget. On the western margins where there is a positive littoral budget (e.g., Line 8), accretion occurred on the stoss slope of the foredune while the crest remained relatively stationary. Between Lines 6 and 8, there is a transition from a negative to a positive littoral budget and, as a consequence, there was relatively little change in the foredune profile at Line 7.

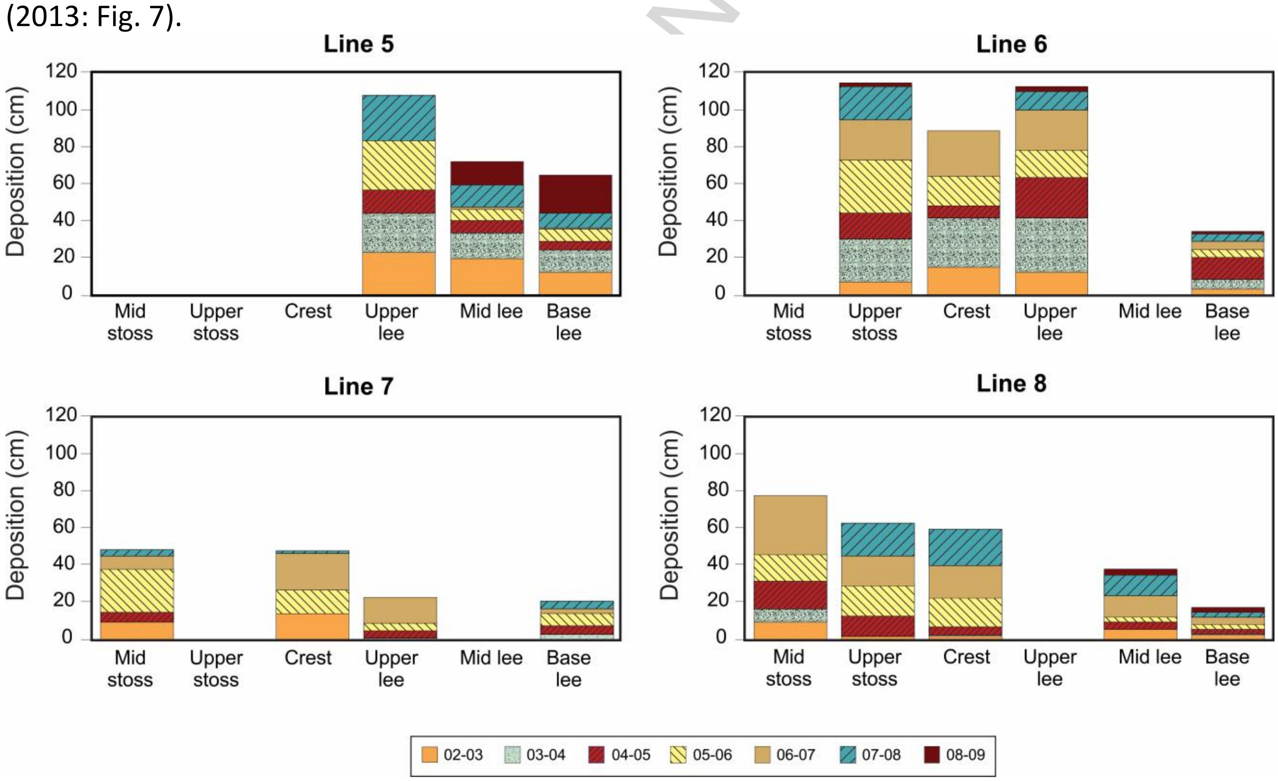
During the winter 2008-2009 season, there was a major dune scarping event that eroded the toe of the entire foredune along Reaches 1 and 2. As a result, little sediment moved onto the foredune at all four lines over the following year, which illustrates how dune scarping and ramp rebuilding processes pre-condition the broader dune profile response. In short, if the foredune is scarped, usually during late fall and early winter storms, a dune ramp must re-establish in order to provide a path for any significant volume of sediment to move up onto the upper stoss slope. Figure 20 includes an inset graph that shows time-series (2009-2010) trends from erosion-deposition pin lines installed on the E and W sides of Line 7. The data are mean values derived from all pins deployed on each line spanning most of the stoss slope. On the E line, there was a period of relatively little change on the stoss slope (Sept – Nov) followed by rapid accretion (Nov – Feb) and then no change (Feb – May), whereas on the W line there was continuous accretion (Sept – Jan) followed by no change afterward. Examination of annual profiles from 2009 and 2010 shows that this was a period of ramp rebuilding. By mid-November 2009, the ramp had built sufficiently in front of the E line to permit transport onto the stoss slope and sediment accretion occurred on the stoss slope at both lines of pins thereafter. The onset of winter terminated the accretion period.

Figure 20: Cross-shore topographic profiles for the period 2002-2011 showing differences in foredune evolution at Lines 5-8 in Reach 2 at the study site (see Fig. 1d for general locations).



The patterns illustrated in Figure 20 are also apparent in more detailed bedframe data (Fig. 21). At Line 5, no net deposition was recorded on the stoss slope or crest because the profile was being displaced landward semi-continuously. Sediment was transported from the beach, across the stoss slope, and onto the lee slope, except when there was no ramp present (e.g., 2006-2007). Line 5 data also show that when there is a significant scarping event, as in 2008-2009, sediment can still move onto the lee slope in association with landward displacement of the entire profile. At Line 6 there was also no recorded deposition on the mid-stoss slope over the seven-year interval. At Lines 7 and 8, where the littoral sediment budget transitions from negative to positive, deposition was recorded on the mid-stoss at both lines in most years, particularly at Line 8 at the W end of Reach 2. Little deposition was recorded on the mid and upper stoss slope at Lines 7 and 8 in years like 2003-2004 because sediment was being trapped in the incipient foredune though, in subsequent years, sediment was able to move up and over the foredune. Seasonal bedframe data and erosion-deposition pin datasets in 2009-2010 (not shown) also show that most sediment transport onto the foredune occurs during the fall and early winter months. There is a secondary peak in the late winter to early spring when the snow and ice cover disappears and the vegetation cover is dormant and of low density (Ollerhead et al., 2013).

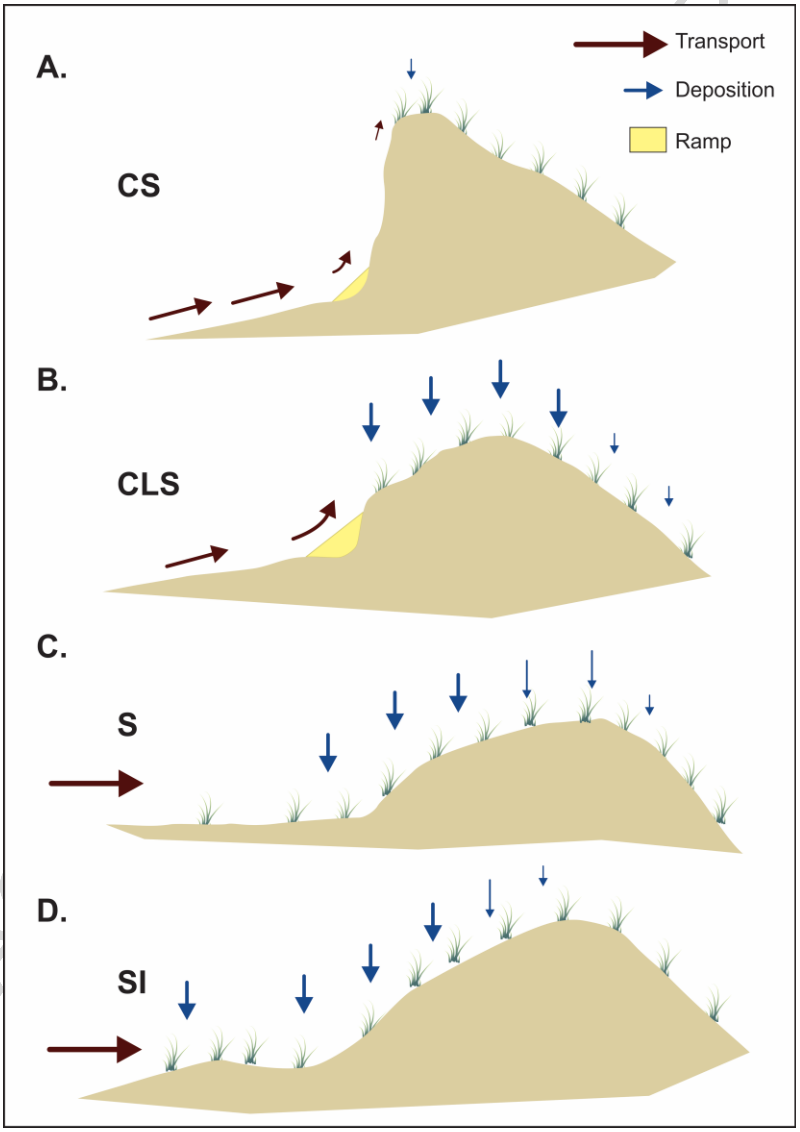
Figure 21: Stacked bar graphs showing the amount and variability of sediment deposition over Lines 5-8 for each year period from 2002-2003 to 2008-2009. Modified from Ollerhead et al. (2013: Fig. 7).



The broad picture of beach-dune interaction and evolution that emerges from this dataset is one of large inter-annual variability driven by: (i) the magnitude and timing of wind events that yield aeolian transport, (ii) the magnitude and timing of storms that alter the beach configuration and potentially scarp the dune toe, (iii) the severity of winter temperature and snow cover conditions, (iv) the spatial variability in beach width at the plot scale due to surface moisture, wind approach angle, and foreshore accretion/erosion, and (v) landform scale variation in the littoral sediment budget. Similar patterns have been documented for many other mid-latitude coasts (e.g., Law and Davidson-Arnott, 1990; Byrne 1997, McKenna Neuman 1990a, 1990b, 1993; Ruz and Allard, 1994; van Dijk and Law, 1995; 2003; Aagaard et al., 2004; Anthony et al., 2006; Pye and Blott, 2008; Yurk et al., 2014). A specific nuance that became evident in the PEI research, however, was the role of foredune scarping and the subsequent development of dune ramps in either precluding or facilitating the transfer of sand to the upper foredune slope. Similar observations were made by Christiansen (2003) and Christiansen and Davidson-Arnott (2004). Dune scarp and fill processes and ramp building are understood conceptually (e.g., Carter et al., 1990), although there are very few field studies of the processes involved. Plot-scale research on alongshore winds and topographic steering (section 4.1.4) as well as seasonal-interannual topographic profile responses (Ollerhead et al. 2013, section 5.2.2) provide some insights. If the foredune is scarped, flow deflection may be significantly different near the scarp than for a non-scarped dune. Winds above the scarp may be deflected onshore towards the crest while wind flow seaward of the scarp may be deflected semi-parallel to the beach during oblique and alongshore winds (Hesp et al. 2013). These wind patterns likely result in dune ramp development because of extended fetch distances that mobilize sediment on the upper beach, which is deposited near the foredune toe to infill the eroded areas. Echo dune formation is also common at the base of scarps and is often the first stage of scarp fill development (Carter et al., 1990; Christiansen and Davidson-Arnott, 2004). Sand deposition in the lee of echo dunes occurs during onshore winds just above sand transport threshold and during more oblique winds. Additionally, slumping or avalanching of the scarp face can occur. All processes lead to infilling of the scarped zone and eventual rebuilding of the foredune toe ramp. Once the ramp is reconstructed, sediment pathways onto the stoss slope are re-established.

A conceptual model based on the beach-dune interactions described above is shown in Figure 22 (see also specific intervals in Fig. 21), which illustrates the following associations: (i) when the foredune is cliffed and a relatively small ramp is present (CS), very little sediment reaches the dune crest or lee slope (e.g., Line 5 2006-07); (ii) where the dune is cliffed and a dune ramp extends over a substantial portion of the lower slope (CLS), moderate to large amounts of sediment can be delivered to the upper stoss slope, crest and lee slope (e.g., Line 6 2002-03, 2003-04; Line 8 2005-06, 2006-07); (iii) where there is a continuous, vegetated stoss slope but no substantial incipient foredune (S), moderate amounts of sediment reach the upper stoss and crest and more limited amounts reach the lee slope (e.g., Line 6 2007-08; Line 7 2002-03, 2006-07); and (iv) where there is a continuous vegetation cover and a well-developed incipient foredune (SI), substantial amounts of sediment are trapped in the incipient dune and lower stoss slope, with small to moderate amounts reaching the crest and little if any reaching the lee slope (e.g., Line 6 2008-09; Line 7 2007-08, 2008-09; Line 8 2002-03, 2003-04). Further details of these responses are described in Ollerhead et al. (2013).

Figure 22: Conceptual diagram of the four characteristic foredune profile forms found at the study site. (A) “CS” is a fully cliffed form with stoss slope > 40° that results from high magnitude wave or storm surge erosion, whereas (B) “CLS” is a cliffed lower stoss slope form resulting from lower magnitude storms. A ramp may or may not be present at any given time. (C) “S” is a stoss slope form that has a continuous vegetated slope of < 40° from the dune toe to the crest and vegetation may extend onto the upper beach, while (D) “SI” is a stoss slope form that has a continuously vegetated slope and that is fronted by a vegetated incipient foredune that is capable of trapping significant quantities of aeolian sand transported off the beach. Modified from Ollerhead et al. (2013: Fig. 6).



5.2.3. Decadal scale observations of extreme overwash and foredune recovery

The field surveying observations at the Greenwich Dunes, PEI site span a period of only one decade, yet, several major storms occurred during this time and resulted in significant erosion of the seaward base of the foredunes. The effects were often locally constrained, but in the winter of 2008-09, the entire shoreline of the study area (and beyond) experienced pronounced erosion. Barrier breaches and washover fans occurred elsewhere along the north shore of PEI, but the foredune at Greenwich was not breached. To place this event into context, examination of historical aerial photographs and local newspaper articles was conducted to see how often erosive storms had occurred in the past. This research indicated that foredune breaching had occurred historically, with a particularly intense storm accompanied by a very high storm surge documented in October 1923 (Mathew et al., 2010). During this time, the entire dune system along the Greenwich peninsula was eroded and overwashed, creating a continuous washover terrace that extended up to 600 m inland. Such inundation overwash is the most severe form of overwash (Sallenger, 2000; Morton, 2002; Donelly et al., 2006) and signifies an extreme event in the spectrum of event magnitude. Moreover, evidence from elsewhere along the north shore of PEI (Simmons, 1982) indicates that this event destroyed the foredune over most, if not all, of the region. Evidence of this event provided an outstanding opportunity from which to assess rates of landform recovery (Wolman and Gerson, 1972) in a beach-dune system.

Mathew et al. (2010) analyzed ortho-rectified mosaics of aerial photographs from 1936, 1953, 1971, and 1997 and produced digital elevation models (DEMs) for each photo year.

Figure 23 shows the orthophoto mosaics for each year that were analyzed to assess landscape change and Figure 24 shows extracted topographic profiles at profile locations 6 and 8 (see Figs. 1, 20). In 1936, over a decade after the storm, a wide, unvegetated intertidal zone and beach is seen with aeolian sand accumulating at the landward edge of the transgressive dunes. Relief landward of the beach along the shoreline was generally <1 m above mean sea level (aMSL). Between 1936 and 1953, vegetation established along much of the shoreline and initiated foredune development in some locations, although there were still several zones of active overwash. It is likely that the slow rate of vegetation establishment in this 20-year period reflects almost complete removal of pioneer vegetation by the storm and, thus, the absence of a nearby source of seeds or reproductive material to re-colonize the area. Less severe overwash events do not produce such intense 'sterilization' of the substrate (Saunders and Davidson-Arnott, 1990; Snyder and Boss, 2002). By 1971, a continuous, broad foredune was present along almost all of the shoreline with a relief of 2-6 m aMSL. In 1997, the foredune ridge had grown to 6-10 m aMSL and the crest was more continuous and located closer to the beach. Thus, while washover healing can take place in less than a decade for relatively small events (Cleary and Hosier, 1979), an event of the magnitude of the 1923 storm required more than five decades for recovery to a form similar to that found at the site today.

Figure 23: Historical aerial photographs showing landscape changes at Greenwich Dunes, PEI from 1936 to 1997. Locations of cross-shore profiles 6 and 8 (from Figures 1 and 20) are indicated as well as extent of established foredunes (yellow polygons). Cross-shore topography along profiles 6 and 8 extracted from the related stereo imagery is shown in Fig. 24. Modified from Mathew et al. (2010: Fig. 4).

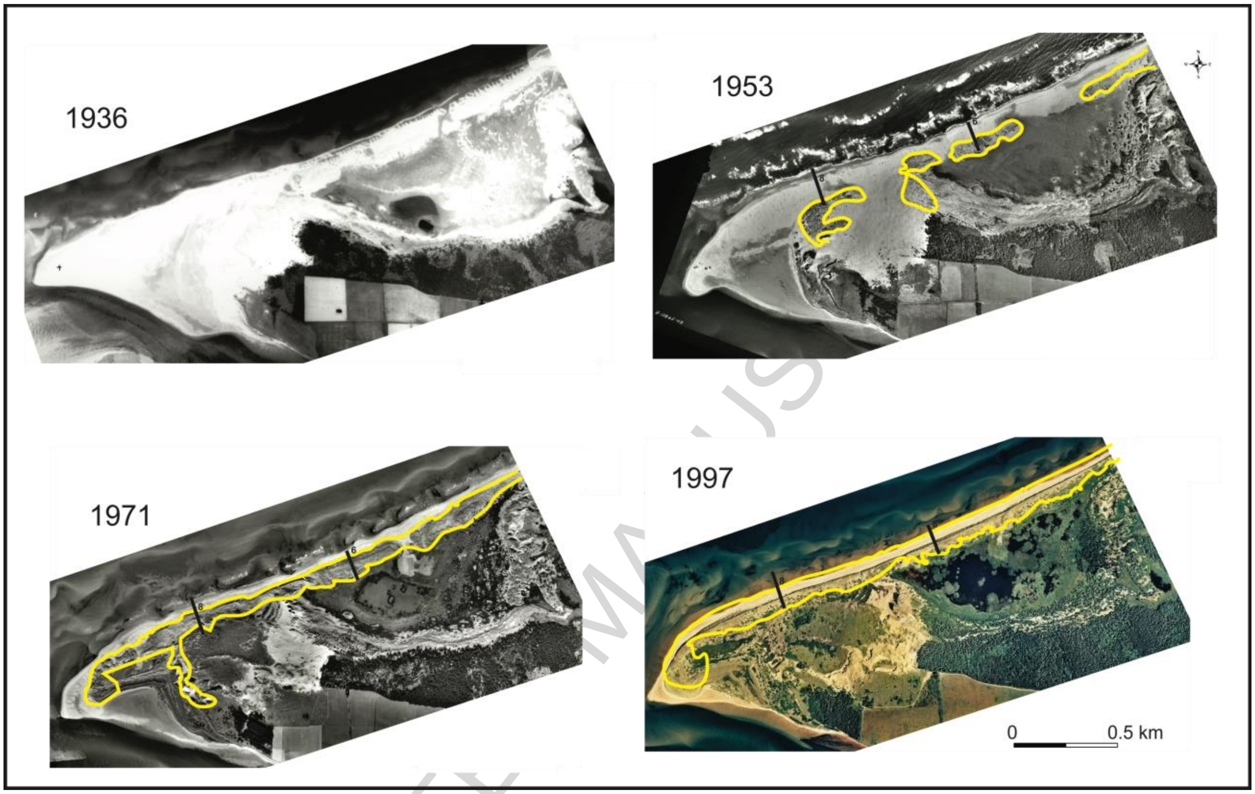
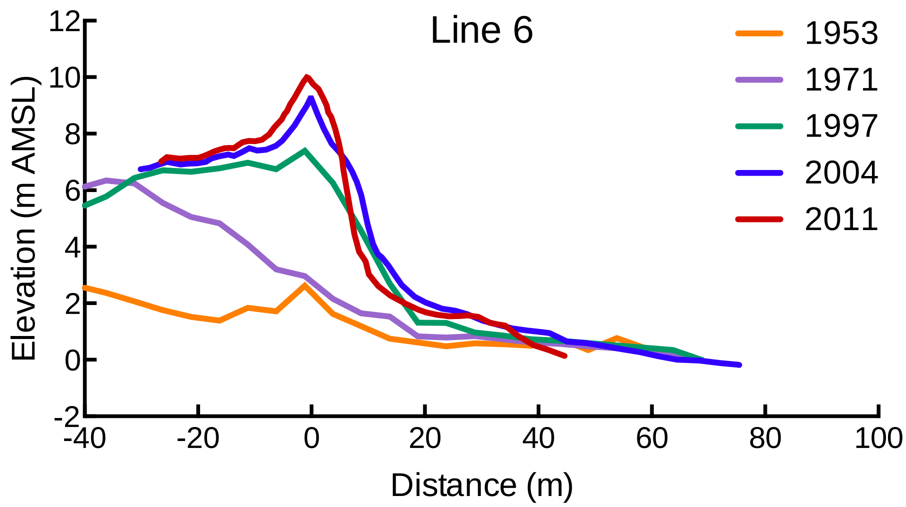
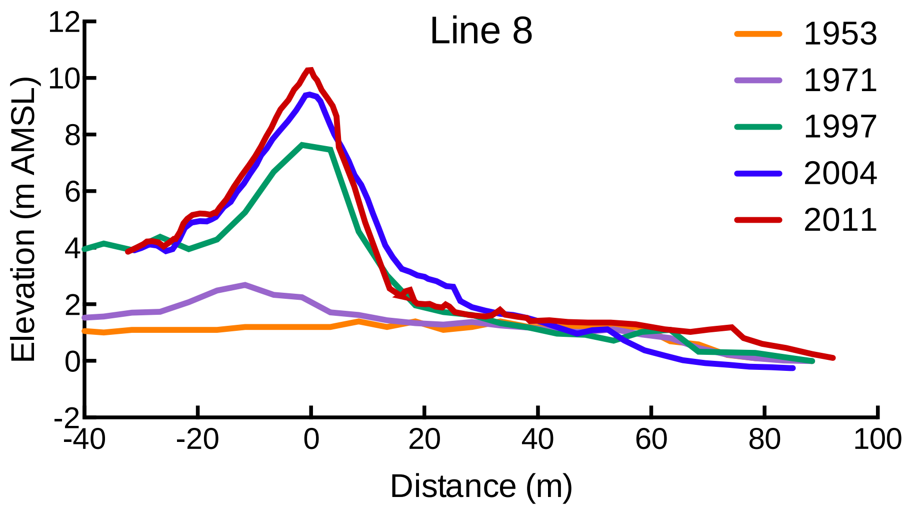


Figure 24: Cross-shore topographic profiles extracted from stereo aerial photography by Mathew et al. (2010) at lines 6 and 8 (see Figs. 1 and 20) from 1953 to 2011 depicting the extent of vertical accretion and foredune recovery following the major overwash event that occurred prior to 1936.





At both the plot and landform scale, the potential for foredune erosion and rebuilding at Greenwich, PEI is highly dependent on the frequency and magnitude of seasonal storm events, most of which occur in the fall and early winter months (Forbes et al., 2004). While a very large storm surge accompanied by large waves is necessary to produce the inundation overwash of the 1923 storm event, the impacts of smaller, less severe storms are also controlled by factors such as surfzone and beach slope and morphology, foredune height and morphology, littoral sediment budget, and the time interval between storms (e.g., Houser et al., 2008; Esteves et al., 2012; Heathfield et al., 2012; Hesp and Smyth, 2016b). The extent and severity of erosion from an individual storm cannot be predicted by modeling storm surge elevation and wave height alone. Other factors, such as the presence and effects of dune ramps and incipient dunes, all influence the extent of erosion and, subsequently, the rate and nature of dune recovery. There are now a number of approaches to modeling dune erosion and overwash from relatively simple models based on a few broad beach and water level parameters (e.g., Komar et al., 1999; Kriebel and Dean, 1993; Larson et al., 2004; Mull and Ruggeiro, 2014) to much more computationally complex 2D cross-shore models such as XBeach (Roelvink et al., 2009; Splinter and Palmsten, 2012; de Winter et al., 2015) or 3D models such as SWAN offshore and XBeach in the nearshore (Dissanayake et al., 2014). Rigorous field-testing of these models, however, requires considerable data on morphology before and after the event, and of ongoing processes during the storm. None of these models adequately couple nearshore processes to aeolian processes in the true sense of beach-dune interaction. Very recently, Zhang et al. (2015) coupled a process-based nearshore model and a cellular automata aeolian model to simulate historical foredune change on the Baltic Coast. Due to the extent and limitations of model calibration, however, accurate prediction of future coastline change at scales of years to decades remains elusive.

5. Landscape scale

In PEI, two controls dominate beach-dune morphodynamics and evolution at the landscape scale. First is the regional RSL trend, which has been rising at a rate of about 0.3 m century^{-1} for the past 6,000 years. Second is the rapid erosion of relatively soft bedrock leading to recession rates of 0.3-1.0 m a^{-1} and high sand supply to the littoral system (Forbes et al., 2004; Webster, 2012). The focus of the PEI research at the landscape scale was on understanding the effects of the ongoing RSL transgression and the influence on the littoral and dune sediment budgets and resulting evolution of the beach-dune system. Two particular questions were addressed: 1) How do observations of decadal scale dune dynamics align with expected responses per the Bruun (1962) model of coastline response to sea-level rise?, and; 2) What is the nature of foredune morphological change (i.e., equilibrium shape and size), if any, associated with ongoing sea-level transgression?

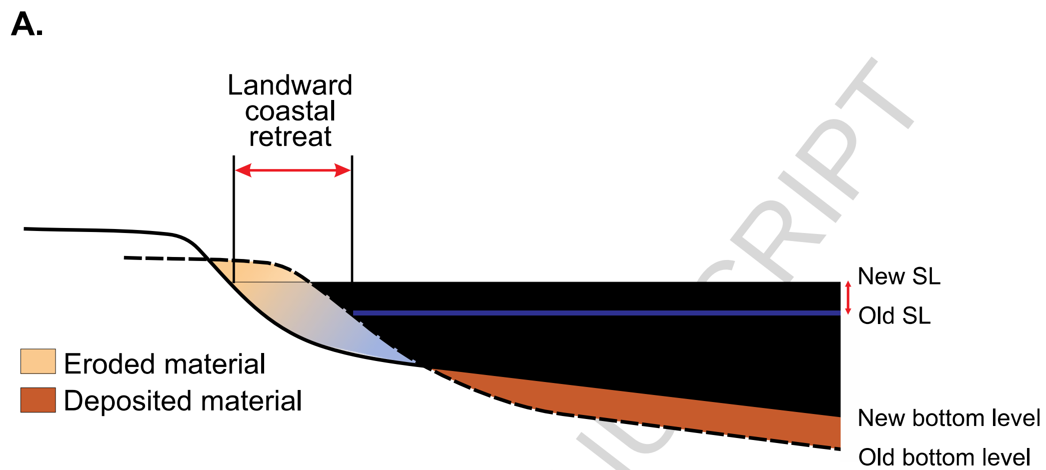
5.1 The classic view of the response of coastlines to sea-level rise: the Bruun model

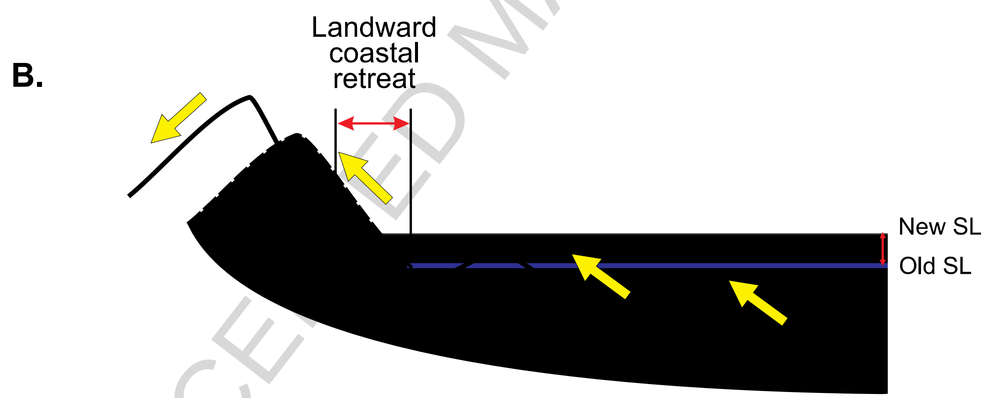
For decades, much effort has been centred on understanding and predicting the response of sandy coastal systems to sea-level rise using the “Bruun Rule” (Bruun, 1962; Schwartz, 1967; SCOR Working Group 89, 1991; Mimura and Nobuoka, 1995; List et al., 1997; Zhang et al., 2004; Pilkey and Cooper 2004; Rosati et al., 2013), which predicts that a sandy coast will respond to progressively rising sea levels by shoreline erosion and recession. The volume of eroded sediment is transported offshore and deposited as a layer with a thickness equal to the rise in sea level (Fig. 25). Thus, the sink for sediment is offshore, which implies that the sediment is lost to the nearshore system as further sea-level rise forces the wave-base upwards. In PEI, sea-level rise over the past 6,000 years should have resulted in very large volumes of sand deposited in the nearshore. However, it is evident from surveys by Forbes et al. (2004) that the shoreface is sediment starved beyond the nearshore bar system. Furthermore, a huge volume of sediment is stored in inlet tidal deltas and in beach and dune deposits on the mainland or on barriers. Thus, the Bruun model appears not to apply to the PEI coastline, and similar conclusions have been reached for a handful of other coasts (e.g., Rosati et al., 2013; Aagaard, 2014).

5.2 A new perspective on the response of beach-dune systems to sea-level rise: the RD-A model

Based on previous work on the dynamics of nearshore bar systems and research on beach-dune morphological changes in PEI, Davidson-Arnott (2005) proposed a conceptual model (aka the RD-A model) of the response of mainland sandy beach and dune systems to sea-level rise that envisions onshore migration of sediments in the nearshore and consequent landward and upward translation of the beach-dune profile. In the RD-A model, the primary sediment sinks are landward of the nearshore, not offshore, and the equilibrium nearshore profile is maintained as its spatial position migrates (Fig. 25).

Figure 25: Schematic illustrations of the Bruun (1962) model of beach profile response to rising sea level (A) showing erosion of the upper beach and deposition in the nearshore to a thickness equivalent to the rise in sea level, and the RD-A model (B) showing erosion and landward migration of the nearshore profile and transgression of the beach and foredune under the same amount of sea-level rise. Modified from Davidson-Arnott (2005: Figs. 1 and 2).





It is now recognized from various lines of evidence that RSL on sandy coasts is generally accompanied by erosion of sand from the outer shoreface and onshore transport resulting in the accumulation of large sediment bodies on land and in shallow water. For instance, seismic profiling and coring off the East and Gulf coasts of the USA and elsewhere in the world have revealed that the Holocene transgression resulted in reworking and onshore transport of sediments on the shelf as RSL rose, leading to the formation of ravinement surfaces that persist to the present (e.g., Belknap and Kraft, 1981; Niederoda et al., 1985; Siringan and Anderson, 1994; Rodriguez et al., 2001; Dillenburg and Hesp, 2009; Goff, 2014; Schwab et al., 2014, Goff et al., 2015). This is also supported by recent analysis of shoreline change and profile evolution (e.g., Short, 2010; Schwab et al., 2013; Rosati et al., 2013; Houston and Dean, 2014).

Sediment budget studies based on long-term monitoring and individual field experiments at Skallingen, Denmark, have shown that sand is transferred from the lower to upper shoreface in response to ongoing sea-level rise (Aagaard et al., 2004; Aagaard and Sorensen, 2012, 2013). These empirical studies are supported by numerical modeling of sand transport by wave events (Aagaard, 2014) at five different sites. A simulation of sand transport for one year using the model showed net sediment transfers that compared well to transport rates estimated from nearshore bar migration and aeolian accretion (Aagaard, 2014). This work provides a mechanism through which the landward transfer of sediments, necessary for translation of the nearshore profile in equilibrium with sea-level rise, occurs as envisaged by the RD-A model.

There is also increasing recognition that on gently sloping coasts, landward translation of barriers often involves overwash and inlet processes that move large volumes of sediment landward (e.g., Dean and Mauremeyer, 1983; Rosati et al., 2013), such as the accretion of barriers on the east coast of Australia towards the end of the Holocene transgression (e.g., Roy et al., 1994; Hesp and Short, 1999; Cowell et al., 1995, 2003). On mainland dunes, landward translation of the foredune occurs by aeolian transport over the dune crest and deposition on the lee slope, as our annual surveys and other studies show (Ollerhead et al., 2013; Hesp and Walker 2013: Figs. 10 and 11, see also section 5.2.2). Appreciable amounts of sand may be transported tens of metres landward of the crest during strong wind events, thus providing a deposit over which the foredune can subsequently migrate (Arens, 1996; Aagaard et al., 2004; Christiansen and Davidson-Arnott, 2004; Hesp et al., 2009, 2013; Petersen et al., 2011; Ollerhead et al., 2013). This is also the mechanism by which landward transgressive dune systems can be fed (e.g., Anderson and Walker, 2006). According to the RD-A model, if the sediment budget is relatively neutral (e.g., Line 7), the beach-dune profile will translate landward in equilibrium with sea-level rise and the transgression distance, R, associated with the rise in sea level can be measured by migration of the dune toe position. In so doing, one would need to account for the local nearshore context over shorter time frames, as demonstrated by the variable responses observed in Fig. 20 (Line 5 to Line 8).

The RD-A model is a simple 2-D model that is best applied to specific cases such as confined headland-bay beach systems where there are no significant alongshore transfers of sand. On most exposed coasts, however, it is necessary to consider the complexities introduced by negative or positive littoral sediment budgets and other factors that may influence the dynamics of beach-dune interaction locally. The positive littoral sediment budget at Line 8 appears to have counter-balanced the landward translation of the profile due to RSL rise for at least a decade, while at Line 5, where the sediment budget is negative, ongoing translation of the shoreline is clearly taking place (Fig. 20). Although these associations are apparent when examining a decade of topographic profile changes, these trends may be significantly altered over periods of centuries. Thus, at the landscape scale, even a large overwash event such as the storm of 1923 can be viewed simply as part of the ongoing 'normal' processes leading to landward translation of the profile (Mathew et al., 2010) according to the RD-A model.

On shorelines characterized by barriers, tidal inlets, estuaries, and lagoons with large accommodation space, the controls on shoreline displacement become highly complex. This applies to much of the north coast of PEI, to the east coast of the USA, and to areas such as the Wadden Sea in western Europe. In these situations, alongshore transfers and accommodation space in lagoons are major controls on coastal evolution and it is increasingly recognized that these 3D sediment transfers have to be modeled explicitly in order to understand the morphodynamic character of the beach-dune profile and nearshore system (e.g., Stive, 2004; Stive et al., 2009; Hinckel et al., 2013; Ranasinghe et al., 2013; Moore et al., 2014). Consideration must also be given to other factors that influence alongshore variations in post-storm dune recovery (e.g., Houser, 2013) such as controls imposed by shallow bedrock outcrops on the potential for shoreline transgression.

6. Summary and Conclusions

6.1 The persistent challenges of scale in beach-dune geomorphology

A continuing challenge for geomorphology, as with many multidisciplinary Earth sciences, is that most knowledge about how natural systems function is bounded by scale-specific constraints inherent to the theories, methodologies, and objects of study that are adopted or constructed in the scientific process. Occasionally, efforts are made to broaden perspectives by considering knowledge from closely allied fields or sub-fields, which often have different methodological and/or theoretical underpinnings. In so doing, a more nuanced understanding of the dynamics of natural systems is often derived that is informed by alternative perspectives and different scales of inquiry. This paper attempts to provide such a 'scale aware' perspective on coastal-aeolian morphodynamics and evolution, based in part on the vast literature on aeolian processes on coasts and deserts worldwide, but primarily on a decade-long research program at the Greenwich Dunes, PEI, Canada. This research program incorporated experimental and monitoring methods spanning the plot (micro), landform (meso), and landscape (macro) scales. It is argued that this approach has led to a more holistic (i.e., multi-scalar), focussed, and comprehensive (albeit incomplete) understanding of a discrete beach-dune system than has been undertaken previously.

An example of the dilemma posed by the scalar boundedness of empirical geomorphic knowledge in this research is the disconnect in knowledge gained between the detailed process observations of sand transport activity and related beach-dune conditions (sections 5.1.2, 5.1.3) and the morphological response observations provided by the seasonal cross-shore beach-dune profiles (section 5.2.2). The former provides key information on the regime of sand transport, erosion, and deposition processes presumed to be representative at a seasonal scale, while the latter yields key insight on the magnitude and direction of seasonal to interannual topographic and sediment budget changes in the beach-dune system. Despite some spatial and temporal overlap (e.g., profile 7 exists in the area of coverage of the camera monitoring system) and the respective richness of these datasets, there remains significant scalar incompatibility or incompleteness between them. For instance, in the absence of information on how and when sediments were mobilized between all of the surveys and at all locations, it is only with much caution and many limitations that one can extrapolate how the temporally-limited and spatially-discrete observations of the transport regime might translate from seasonal or decadal trends in beach-dune morphology or sediment budgets. Similarly, it is incredibly difficult to retrodict prior system states that preconditioned the present observed conditions. So, despite great efforts here to span spatial and temporal scales of process-response interactions, there remain some appreciable gaps at scale transitions, in particular.

6.2 Plot scale complexities encourage consideration of landscape scale linkages

The plot scale research at the Greenwich Dunes, PEI, provided significant insights into the widely recognized inability of conventional sediment transport models to predict sand mass flux moving across the beach-dune system as a function of wind strength alone (e.g., Sherman and Li, 2011). Multiple supply-limiting constraints (e.g., surface moisture, grain size and texture, bed roughness, salt crusts, vegetation) and transport-limiting factors (e.g., vegetation, coarse lag deposits, wrack) collectively result in sand transport intermittency. Many of these factors, and their effects on sand transport, are spatially variable as the wind field transitions from the nearshore to the foreshore, across the back beach, and on to the foredune. However, the plot scale research also showed that these factors are coupled and co-evolve both in space and time, with often counter-intuitive outcomes depending on feedback relationships. For example, the veering of wind direction from cross-shore to oblique approach angles can strongly influence the delivery of sand to the foredune by way of the fetch effect. Generally, sand transport increases across the back beach due to increasing fetch distance, however, less may be delivered to the foredune because total sediment transport across the dune toe line decreases in proportion to the cosine of the wind angle (e.g., Bauer and Davidson-Arnott, 2003; Delgado-Fernandez, 2010). Similarly, as the incident wind begins to interact with foredune topography, and the vegetation thereon, there can be considerable changes in wind speed and direction as a result of flow deflection, streamline compression or expansion, flow acceleration or deceleration, vegetation density and distribution, and related secondary flow patterns (e.g., flow steering, separation, reversal, jet formation) with significant implications for sand transport pathways (e.g., Walker et al., 2006; Walker et al., 2009a; 2009b; Bauer et al., 2012; Hesp et al., 2009; 2015; Hesp and Smyth, 2016). Turbulence within the boundary layer is influenced significantly by these wind-topography interactions and this research, along with the findings of other researchers, suggests there is some commonality in the turbulent signatures found at key locations such as the foredune crest and toe (e.g., Chapman et al., 2012; 2013; Wiggs et al., 1996b; Wiggs and Weaver, 2012). This work also indicates that there is sufficient uncertainty surrounding the association of turbulent Reynolds Stress with sand flux to question whether or not this parameter is a reliable indicator of wind strength for predicting aeolian sand transport over complex terrain.

There also remain significant gaps in our knowledge with regard to how and when sediment is moved from the nearshore to the foreshore and, eventually, to the foredune (Houser, 2009; Houser and Ellis, 2013). Furthermore, it remains unclear as to what types of events are most significant in growing or maintaining foredunes. For instance, the importance of offshore winds in maintaining foredunes or contributing to the development of sand ramps and healing wave-cut dune scarps has become recognised increasingly (e.g., Jackson et al., 2011; Lynch et al., 2009; 2010; Bauer et al., 2015). Other external factors such as wave run-up, tidal and storm surge innundation, salt spray, rainfall, snow/ice cover, and progressive sediment stripping and deflation during transport events also present spatial and temporal complexities in the aeolian sand transport process. At times, therefore, it is possible to have some portions of the beach where there is no transport because wind strength is insufficient to entrain sediments, other portions where wind strength is adequate but surface controls restrict the rate of sand supply (leading to intermittency), and yet other areas where there is sufficient wind and sand available to yield substantial transport. All of the complexities resulting from flow-landform-transport interactions over beaches and dunes at the time scales of single events and seasons, thus, beget consideration of broader landscape scale controls.

6.3 Bridging the plot to landform scale transition

In response to the complexities at work at the plot scale, standard equilibrium models of sand transport often fail to produce accurate estimates across beaches and over foredunes.

This serves as a reminder that conceptualizing and modelling sediment transport across beach-dune systems as controlled by singular factors in isolation is an inadequate approach. The collective body of research reviewed in this paper also highlights how information about the broader (landform scale) context is critically important. The conceptual scheme in Table 1 shows that, in order to understand sediment transport rate and patterns of erosion and deposition across the beach (i.e., the dependent variables), one requires knowledge of the independent variables (e.g., wind speed, wind approach angle, surface debris, vegetation, salt crusts, surface moisture, snow or ice cover, beach width and slope) as well as knowledge of a few key parameters such as foredune size and geometry and vegetation species, distribution, cover density, height, and morphology. However, at the landform scale, these all become dependent variables (i.e., things we want to predict or better understand) that are governed by a range of independent variables (e.g., wave climatology, climatic conditions, geological setting). In turn, these independent variables dictate the overall supply of sediment available for beach-dune development. In other words, to improve understanding of sediment transport and beach-dune morphodynamics at a particular site, landform-scale factors that influence plot-scale dynamics must be factored in. Essentially, a typology of events is required that distinguishes them according to their character and effectiveness in yielding geomorphic change on the beach-dune profile, and that includes information on their magnitude, frequency, and duration of occurrence.

The research at Greenwich Dunes attempted to provide information that links the plot scale to the landform scale. For example, the monitoring and modelling work of Delgado-

Fernandez and Davidson-Arnott (2011) and Delgado-Fernandez et al. (2009, 2012, 2013a) demonstrates that a simple aeolian transport regime assessment, such as the Fryberger and Dean (1979) model, is inadequate for predicting long-term sand supply to foredunes as it does not consider the range of supply-limiting conditions in coastal regions that occur. For instance, intense winter storm events with powerful winds, which would yield significant transport per the Fryberger and Dean (1979) model, are often insignificant in terms of sediment delivery to the foredunes simply because the beach is covered by snow and ice. Similarly, strong wind events must also be considered in relation to the surface moisture state of the beach, which is controlled by precipitation amounts, relative humidity, solar forcing, tidal stage, and storm surge. During the nine-month photographic observation period of transport activity at the PEI study site, only three (of 184) wind events accounted for 75% of the total sand delivered to the foredune. Sand transport over the foredune was further influenced by the density of vegetation cover (Ollerhead et al., 2013), which also has a seasonal signature that must be accounted for in long-term modelling.

6.4 Bridging the landform to landscape scale transition

The transition from landform to landscape scales is similarly critical, as demonstrated by the cross-shore profiles measured for over a decade at the Greenwich Dunes site (Figure 20). Some regions (e.g., line 5) suggest a continuous and progressive landward migration of the dune with little change in overall profile form. In contrast, other sites (e.g., line 8) shows a relatively stable crest location with seaward progradation of the stoss slope, while others (e.g., line 7) remained stable and virtually unchanged. This suggests a transition from a negative littoral sediment budget to the East (line 5) to a positive budget to the west (line 8). Clearly, a linear extrapolation of any individual trend from these locations would suggest very different styles of shoreline evolution over the next century. Thus, the littoral sediment budget affects the beach width, which in turn influences: i) the fetch effect and, hence, potential sediment delivery to the foredune as well as, ii) the propensity for wave scarping of the dune toe during high water storm events. The presence of a scarp or a sand ramp at the base of the dune also strongly influences the ability of sand to move onto the stoss slope and toward the dune crest.

To understand how this coastline might evolve over the next century requires additional information at the landscape scale on rates of relative sea-level rise, the broader geological context of the north shore of PEI, as well as the history of regional coastal change as documented in archives (e.g., Mathew et al. 2006) and via proxy data in the sedimentological record. As uniformitarianism would suggest, these are perhaps our best indicators of what may happen in the future, but ideally this information could be integrated back into the landform scale and then to the plot scale, so as to provide a deeper understanding of how we might reliably predict the future rather than simply extrapolate trends. For example, framing the understanding of coastline evolution within the RD-A model for the response of sandy beach-dune systems under rising sea levels challenges scientists to predict the events that yield the landward (and upward) translation of the beach-dune profile from year-to-year. In turn, this requires the capacity to predict the nature of sediment transport processes across beach-dune systems at the plot scale, which leads us back to the uncertain nature of the relation between sediment flux and wind strength.

Much has been written about the unlikely prospects for 'upscaling' micro-scale knowledge of Earth surface system function obtained via scientific reductionism to macro-scale system outcomes (e.g., Sherman, 1995; Bauer and Sherman, 1999; Bauer et al., 1999), especially in light of such challenges as error propagation in models, chaotic behavior in non-linear systems, and climate non-stationarity. One might ask then, “If we are doing science in the service of coastal resource managers who are interested primarily in landform and landscape scale outcomes, why even bother with plot-scale experiments?” The answer, it seems, is to provide a more holistic understanding of the system under investigation (or under management), in terms of the range of processes, feedbacks, controls, and linkages between scales of interaction, to thereby reduce the probability of making incorrect assumptions or predictions about the future. Knowledge and understanding at each of the scale domains is not independent of the other, and there has to be consilience or unity of knowledge (Wilson, 1998). So, from a management perspective, given the increasing pressures and impacts of global climatic and environmental change, there is a clear need for more applied, integrated, multi-scalar knowledge. Ignoring the context provided by knowledge at shorter and longer scales then seems like a perilous course of action.

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