The monitoring and modelling of the impacts of storms under sea-level rise on a breached coastal dune-barrier system

VOLUME I

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Acronyms

ANOVA	Analysis of variance
AR	(IPCC) assessment report
CAD	Computer aided design
CANUPO	CAractérisation de NUages de POints (point cloud classification system of Brodu and Lague (2012))
CIM	Continuous injection method
CMIP	Coupled model intercomparison project
DEM	Digital elevation model
DOD	DEM of difference
DTM	Digital terrain model
dGPS	Differential global positioning system
EBK	Empirical Bayesian Kriging
EDM	Electronic distance meter
EPA	Environmental Protection Agency
ftp	File transfer protocol
GCM	General circulation model
GHG	Greenhouse gas
GIA	Glacio-isostatic adjustments
GIS	Geographical information system(s)
GMSL	Global mean sea-level
GMSLR	Global mean sea-level rise
GNSS	Global navigation satellite systems
GPS	Global positioning system
HDS	High-definition surveying
ICZM	Integrated coastal zone management
IDW	Inverse distance weighting
IMU	Internal measurement unit
INFOMAR	Integrated mapping for the sustainable development of Ireland's marine resource
IPCC	Intergovernmental Panel on Climate Change
IRSL	Infrared stimulated luminescence
LAT	Lowest astronomical tide
LiDAR	Light detection and ranging
LPA	Lowest points analysis
MIKE21	Coastal modelling package developed by DHI
MSL	Mean sea-level
NAB	North Atlantic basin
NN	Natural neighbor
ODM	Ordnance datum Malin Head
OPW	Office of public works
PCA	Principal components analysis
PSA	Particle size analysis
RCM	Regional circulation model
RCP	Representative concentration pathways
RF	Radiative forcing
RMS	Root mean square
RMSE	Root mean square error

RTK GPS	Real-time kinematic global positioning system
SAT	Surface air temperature
SIM	Spatial integration method
SLC	Sea-level change
SLP	Sea-level pressure
SLR	Sea-level rise
SRES	Special report on emissions scenarios
TIM	Time integrated method
TIN	Triangular irregular network
TLS	Terrestrial laser scanning (can be used interchangeably with the term "ground-based LiDAR")
UAV	Unmanned aerial vehicle
WAM	Wave Prediction Model
WCRP	World Climate Research Programme

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Abstract

Little is known about the impacts of storms on breached barriers, and virtually nothing is known about the impacts of storms under a rising sea-level on these systems. This PhD research aims to help fill this gap. In 2008, barrier breaching at Rossbehy, Co. Kerry resulted in the establishment of a new tidal inlet. Semi-diurnal tidal exchange through the new channel has been on going since this event. Rossbehy provides an excellent opportunity to study the influence of storms on barrier evolution post-breaching.

A two-year monitoring campaign was undertaken to assess the morphological impacts of storms on Rossbehy and a neighbouring barrier, Inch. Multiple topographic surveys were conducted using terrestrial laser scanning (TLS) technology. The logistics of collecting, storing, processing, and analyzing this data were addressed in this research project. Major volume losses were recorded over the duration of the monitoring period at Rossbehy, but not at Inch. This difference was likely due to the orientation of the sites in relation to the main inlet channel.

Meteorological data and numerically simulated nearshore wave data were used to identify and characterize storm events that occurred during the monitoring period. Strong negative statistically significant correlations were observed between rates of dune volume change and storm duration for events that occurred during the monitoring period. Additional statistical analyses revealed that event duration in combination with maximum significant wave height were the best predictors of dune volume change at Rossbehy.

A novel experiment was set up to assess the impacts of storms under future sea-level rise (SLR) on Rossbehy using numerical modelling and TLS data. Numerical modelling was performed in MIKE21. TLS data was used to evaluate the effectiveness of the model in simulating dune volume changes near the breach. The results of the experiment indicate that under future SLR, storms will contribute to a net offshore movement of sediment in the near shore zone of Rossbehy. This will inevitably lead to shoreline retreat and could result in the

possible drowning of the barrier if back barrier saltmarsh sediments cannot accumulate fast enough to keep up with rising sea-level.

Based on the results of the monitoring campaign and modelling experiments, a conceptual model of the evolution of the system was developed – the S-SLR model. The model integrates the influence of storms under a rising sea-level into a previously developed conceptual model put forth by O'Shea (2015). The new model accounts for sediment deficits in the near shore zone caused by storms under a rising sea-level.

This is the first assessment of the potential impacts of storms under sea-level rise on a breached barrier system in Ireland. It is envisaged that this study will serve as baseline from which to compare future process studies of similar systems.

This is to certify that the work I am submitting is my own and has not been submitted for another degree, either at University College Cork or elsewhere. All external references and sources are clearly acknowledged and identified within the contents. I have read and understood the regulations of University College Cork concerning plagiarism.

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Sarah Kandrot

1 Introduction

Coasts are perhaps the most active of all geomorphic environments and are subject to a wide range of complex and dynamic processes that operate over various spatial and temporal scales. Effective adaptation to this dynamism through coastal management practices necessitates a scientific understanding of these processes, especially as coasts respond to climate change (Wong *et al.*, 2014). Already coastal environments the world over are experiencing the adverse consequences of hazards related to a warming climate and rising sea-level (Meehl *et al.*, 2007; Parry *et al.*, 2007; Trenberth *et al.*, 2007; Füssel and Jol, 2012; Field *et al*., 2014). In 2012, Hurricane Sandy caused an estimated $82 billion in damage along the New York and New Jersey coasts alone (Hernandez, 2013). In Europe, the annual cost of coastal erosion mitigation measures is estimated to be about €3 billion per year (EUROSION, 2004). Without taking adaptation measures, the direct cost of sea-level rise to the EU could reach €17 billion per year by 2100 (Hinkel *et al.*, 2010). Coastal research is now more economically relevant than ever.

Coastal environments, however, are complex systems with multiple feedbacks and understanding the interrelationships between and among the different variables and parameters associated with these systems is not always straightforward. Therefore, coastal researchers often rely on the development and use of models to gain insights into coastal structure, organisation, and functioning (Trenhaile and Lakhan, 1989). These models may be conceptual, empirical, or numerical. Regardless of the type of model, if they are to be of any use, empirical observations are required to develop and validate them. Unfortunately for many of the world's coasts, coastal process data from monitoring procedures is lacking (Christie *et al.*, 2005; Devoy, 2008; Swift, 2008; Tribbia and Moser, 2008; Rosenzweig *et al.*, 2011).

According to the Intergovernmental Panel on Climate Change's most recent assessment report (AR5),

> *"Coastal systems and low-lying areas will increasingly experience adverse impacts [due to climate change] such as submergence, coastal flooding, and coastal erosion due to relative sea level rise*"
>
> (Wong *et al*. 2014, p. 364)

In order to effectively deal with these impacts, an understanding of coastal functioning is critical. This research project will deliver data that will help further our understanding of coastal processes in Ireland. The study is specifically concerned with the Inch and Rossbehy barrier spits, located in Dingle Bay, Co. Kerry (figure 1.1). The barriers contain pristine wide, sandy beaches and are locally important sites on the Wild Atlantic Way tourist route. In 2008, barrier breaching occurred at Rossbehy. This resulted in the formation of a new tidal inlet through which semi-diurnal tidal exchange continues as of 2016. Much of this study is based around Rossbehy. This is due to the fact it was the site of breaching and therefore offers a unique opportunity to gain insight into the morphodynamic evolution of a breached barrier. Breaching could affect sediment transport patterns and the delivery of sediment to neighbouring Inch. The impacts of breaching of Rossbehy on Inch (if any) have not yet been evaluated. One of the aims of this study is to address this. In addition, little is known about the influence of storms on the morphodynamic evolution of breached barrier systems. Moreover, the potential impacts of storms under sea-level change on such systems have not yet been fully addressed in the coastal or engineering literature. Another aim of this PhD is to help fill this gap.

Eustatic (global) mean sea-level is projected to rise by 0.26 to 0.82 m[1] by 2100 (Church *et al.*, 2013b), therefore rising seas will have a significant impact on coastal barriers in the coming decades (Pilkey and Young, 2009; Pilkey, 2011).

[1] Depending on emissions
[2] Here, extreme storms are referred to generally as infrequent events characterized by record-

In addition, extreme storms[2] are projected to become more frequent and intense (Kiely *et al.*, 2005; Beniston *et al.*, 2007; Mori *et al.*, 2010) and, according to Devoy (2008, p. 327), these changes "are likely to cause Ireland's coastal wetlands and other soft-sedimentary systems to be among the first in Europe to respond to storm-led sea-level rise." As such, results from this type of research are critical to inform sound coastal management policies in a changing climate.

This chapter introduces the aims and objectives of the PhD and defines the basic rationale for the work. Chapter two introduces the basic theoretical framework required for interpreting the research presented in later chapters. Chapters three to six go on to critically review the literature specific to the Inch-Rossbehy barrier system (chapter 3), sea-level rise (chapter 4), storms and their impact on coastal barriers (chapter 5), and terrestrial laser scanning (chapter 6). Chapters seven to ten present methods and results. Chapter seven focuses on the morphological monitoring campaign undertaken during this PhD, including the collection, processing, and analysis of terrestrial laser scanned data. Chapter eight examines the relationships between observations of morphological change at the sites and storms that occurred during the morphological monitoring campaign. Chapter nine presents sediment tracer experiments undertaken to better understand sediment transport patterns at Rossbehy. Chapter ten presents process-based modelling experiments which were run to examine the relative morphological influence of events under various sea-level rise scenarios on Rossbehy. Chapter eleven delivers a critical interpretation of the results of this research and presents a new conceptual model of the influence of storms on a breached barrier under SLR based on the findings of this study. Finally, chapter twelve concludes the thesis.

1.1 Research Summary and Objectives

The primary aim of this research is to evaluate the importance of storms as a driver of morphologic change on a breached barrier system under present and potential future sea-levels. This was achieved using field monitoring and numerical modelling techniques. A two-year monitoring campaign, which involved the use of terrestrial LiDAR surveying, was undertaken to assess the

[2] Here, extreme storms are referred to generally as infrequent events characterized by record-setting meteorological characteristics.

morphological impacts of storms on the system. The characteristics of events observed during the monitoring period served as inputs in a 2-dimensional horizontal (2DH)[3] process-based modelling experiment in which the relative impacts of storm events were assessed under different SLR scenarios. While it is acknowledged that caution must be exercised in the interpretation of 2DH process-based modelling results, especially when limited data are available for model validation and calibration, it is argued in this thesis that such results can be useful, and may even be critical, to inform sound coastal management policies in a changing climate.

The system under investigation is somewhat unique in terms of morphology, origin, and functioning. In fact, few pre-existing models of barrier development apply specifically to systems with similar characteristics and/or history to Inch/Rossbehy. As such, the system offers a unique opportunity to test these models, and, perhaps, build on them or create new ones.

A series of objectives related to the research aim have been laid out. These are described and commented upon as follows:

- **Assess the viability of terrestrial laser scanning as a monitoring technique in vegetated coastal dune environments** - A relatively new form of surveying, terrestrial laser scanning, was used to monitor the morphological response of the Inch and Rossbehy barrier dunes to storms. Given its limited use in such environments, this is an opportunity to assess the viability of the technique.
- **Identify potential statistical relationships between storm characteristics and observed morphological changes at Inch and Rossbehy** – The purpose of this is to quantitatively examine the potential existence of relationships between storms and observed changes in foredune morphology in an effort to evaluate the relative importance of various storm characteristics, including frequency, time between events, significant wave height, etc.
- **Undertake a sediment tracing experiment at Rossbehy –** The motivation for the tracing experiment is to better understand the transport

[3] 2DH = 2-dimensional vertically (depth) averaged model

processes in operation the site and to help verify that numerically modelled transport pathways are broadly in agreement with observations.

• **Develop a numerical modelling experiment for assessing the potential impacts of storms under future SLR on Rossbehy** – A numerical modelling approach was undertaken to support the achievement the aim set out in this research.

1.2 Rationale

The influence of storms under a rising sea-level on breached barriers has not yet been fully addressed in the coastal or engineering literature. The Inch-Rossbehy system presents an opportunity to study these drivers. While a conceptual model of the morphodynamic evolution of Rossbehy was recently proposed by O'Shea (2015), that model did not take into account either the influence of storms or SLR on the system. There is a need to better understand the influence of these drivers, especially given the following:

1. Rates of SLR are stated to have been important determinants of morphosedimentary behaviour at Rossbehy in the past (Cooper *et al*., 1995; Delaney *et al*., 2012);
2. Relative sea-level is projected to rise in the region by between c. 45-70 cm (Lowe *et al.,* 2009; Grinsted *et al*., 2015);
3. The negative impacts of both storms and SLR on similar type systems, such as barrier islands, are well documented (*e.g.* Morton, 2008; FitzGerald *et al*., 2007); and
4. A scientific understanding of the influence of these drivers is required for effective and sustainable coastal management, which is unfortunately lacking at present in Ireland (Gault *et al.*, 2011; Marchand *et al*., 2011).

In addition, an innovative aspect of this study is that it tests the viability of a relatively new form of technology – terrestrial laser scanning – as a morphologic monitoring tool in vegetated dune environments. The use of TLS in coastal environments is still fairly novel and best practice standards are practically non-existent. There is a need for the development and/or enhancement of methods for collecting and processing TLS data, including methods for registering multi-

temporal scans, filtering vegetation from TLS point clouds, and generating digital elevation models (DEMs) and DEMs of difference (DODs) and calculating volumetric change.

2 Theoretical Background

This chapter provides the theoretical framework required for interpreting the research presented in later chapters. The approaches and theories used to study coastal environments are multidisciplinary and require a basic knowledge of concepts in coastal geomorphology. In an effort to contextualise the wider literature presented later, a basic introduction to relevant coastal concepts follows with examples from Inch and Rossbehy. Given that the aim of this research relates to understanding the morphological evolution of the system, an introduction to the types of sedimentary environments present within the study area and the processes responsible for driving the evolution of the system is relevant. In addition, basic concepts and equations that form the basis of modern numerical modelling, which has been employed in this research, are presented.

2.1 Coastal sediment barrier systems

Coastal barriers are accumulations of sand or gravel that form as a result of the combined action of wind, waves and currents. In terms of morphology, they are usually linear in shape and low in elevation, although they may aggrade to several tens of metres above sea-level (Davis and FitzGerald, 2004). The Inch and Rossbehy barriers are comprised of fine- to medium-sized sand, with dunes reaching elevations of 15-20 m.

Barriers form part of a larger sedimentary system, the barrier system. A barrier system is a complex of sedimentary environments shaped by wind, waves, and currents. Sediment exchange occurs within and between these environments as a result of transport by hydrodynamic and/or aeolian processes. While there are various types and morphologies of barrier systems, many share the following six general environments[4]:

- the mainland
- a back-barrier lagoon or bay
- the barrier or sand/gravel body itself
- one or more inlets and inlet deltas

[4] These have been delineated by Oertel (1985), specifically, with respect to barrier island systems, but they may be extended to barrier systems in general.

- a nearshore platform
- the shoreface

Barrier systems may also (but do not always) support other sedimentary environments such as dunes, salt marshes, and tidal flats. The sedimentary environments present at Inch and Rossbehy are illustrated in figure 2.1.

2.1.1 Types

Three general types of barriers can be distinguished: barrier islands, welded barriers and barrier spits (Davis and FitzGerald, 2004). These may be defined based on their connection to the mainland, with barrier islands completely isolated from the mainland, welded barriers attached at both ends, and barrier spits attached at only one end. Inch and Rossbehy are barrier spits. The spits support large dune barriers characterised by Holocene progradation. These terms are explained as follows:

Barrier Spits

Spits develop where there is a sudden change in the shape of the coastline, *eg.* at re-entrants in the shoreline or at major headlands (figure 2.2). Refraction of waves around the updrift corner of the re-entrant results in a reduction in wave energy and therefore deposition of sediment. The accumulation of sediment at this corner initiates spit growth. Spit growth then continues until the point where currents, often directed perpendicular to the spit, are sufficiently strong to limit sediment transport or until sediment supply runs out. As such, sediment supply is a principal control on the evolution of spits.

While Inch and Rossbehy are referred to as spits in the literature, it has been argued that such a categorisation may be misleading due to the fact that the barriers exist within a narrow, swash-aligned bay, which limits the influence of longshore sediment transport on their evolution. As such, Devoy (2015) refers to them as either “spit-like barriers” or “beach-dune barriers.”

Dune Barriers

Barriers comprised of sand dunes are called dune barriers. These are accumulations of sediment that form above the high water mark as a result of

wind action. They may consist of a single linear ridge (a foredune) or a series of ridges. Foredunes should not be confused with beach ridges (Hesp *et al.* 2005). Hesp *et al*. (2005) have distinguished between the two: While both typically form above high tide and may appear similar in morphology, especially on aerial photographs, beach ridges are swash and storm wave built deposits, while foredunes are formed by aeolian processes. Figure 2.3 shows an example of a gravel beach ridge and foredune at Inch near the study site. As a general rule of thumb, Short and Hesp (1982) have found that foredunes are best developed on flat, sandy dissipative shorelines (where wave energy is more widely distributed) and poorly developed on steep, reflective beaches (where wave energy is more concentrated), with a continuum of other intermediate states in between.

Prograding, retrograding, and aggrading barriers

Barriers can be prograding (migrating seaward), retrograding/transgressive (migrating landward), or aggrading (building vertically). The existence of historical recurves (drift aligned ridges) at Inch and Rossbehy (figure 2.4) suggests that, at least for some of their history, the barriers were prograding. Two historical recurves have been identified within the barrier interior at Rossbehy by O'Shea (2015). The recurves are thought to represent either earlier northern limits of dune progression or southern limits to a historical breaching event. The existence of these ridge patterns may suggest that spit accretion at the sites occurs episodically in response to changes in sediment supply. Presently, minor drift-aligned recurves can be found at the distal ends of Inch and Rossbehy (figures 2.4). Further discussion of the evolution of the barriers is provided in chapter 3.

2.1.2 Barrier morphology

The overall morphology of coastal barriers in general is controlled by various parameters, including tidal range, wave energy, sediment supply, accommodation space, sea-level trends, and basement geometry. The ratio of wave height to tidal range dictates the presence and distribution of barriers (Davis and FitzGerald, 2004). According to the Hayes model, there are three major types of depositional coastlines, which are categorised based on this ratio – wave-dominated, mixed energy and tide-dominated (Hayes, 1979). Barriers occur almost exclusively in the wave-dominated and mixed energy settings

(Davis and FitzGerald, 2004). With a spring tidal range of 3.2 m and a mean offshore significant wave height of 2.8 m, the Inch-Rossbehy system is considered a mixed energy environment (Sala, 2009).

Barriers can also be characterised based on their orientation with respect to incoming waves and sediment supply (figure 2.5). On this basis, they can be either swash-aligned (orientated in the direction of the incoming waves) or drift aligned (orientated oblique to incoming waves). According to Davies (1972), drift-alignment occurs on barriers where the down-drift sediment supply is adequate to fulfil the longshore power for transport. Such barriers are characterised by high sediment supply, spit growth, and spatial stability. Alternately, swash-alignment occurs where the down-drift sediment supply is insufficient to fulfil the longshore power for transport. These barriers therefore align themselves with the dominant wave direction. Equilibrium is said to occur on swash-aligned barriers when longshore sediment transport potential is equal to zero, at which time the barrier becomes vulnerable to breakdown, migration and spatial instability (Orford *et al*., 1996). Prior to Rossbehy breaching, both Inch and Rossbehy were predominantly swash-aligned.

The principal process control on the plan-form orientation (*e.g.* swash-alignment or drift-alignment) of coastal barriers is likely longshore sediment supply (Orford *et al*., 1996; Sala, 2009). Working in Nova Scotia, Canada, Orford *et al*. (1996) related changes in the orientation of gravel-dominated spits to a reduction in the longshore sediment supply. In this model, the growth of a drift-aligned feature is controlled by the rate of increase in sediment supply. Switching (from drift-alignment to swash-alignment) may occur when there is a reduction in sediment supply, initiating a process termed cannibalisation. Cannibalisation occurs when this reduction in sediment supply leads to erosion of material from an updrift source and deposition of that material in a downdrift sink. Refraction induced changes in longshore power gradients facilitate the concomitant development of wave-sediment cells (figure 2.6). Orford *et al*. (1996) differentiate between two types of cannibalisation – macro-scale cannibalisation occurs when the entire length of the barrier is restructured as one cell, while micro-scale cannibalisation occurs when several cells are created along the length of the barrier. Within these cells, erosive zones are associated with zones of accelerating sediment

transport, whereas accretive zones are located in areas of weaker sediment transports. Further discussion on sediment cells is provided in section 2.4.3 (sediment budget).

While not developed for sand dominated barriers, there are some key similarities between the Nova Scotia and Inch/Rossbehy barriers. Firstly, all have similar origins, *i.e.* they were created from glacially derived sediments. Second, similarities in the behaviour of Rossbehy and the Nova Scotia barriers have been identified by Sala (2009) and O'Shea and Murphy (2013). In the early 2000s, Rossbehy was stable and swash aligned. This corresponds with the macro-scale cannibalisation phase of Orford *et al*. (1996)'s model, during which time a single sediment cell is thought to have existed along the length of the barrier. During the micro-scale cannibalisation phase (2007-2008), the development of multiple cells culminated in breaching, after which time the drift-aligned zone began to grow at the expense of the more stable swash-aligned zone (which is still on-going as of 2016). O'Shea and Murphy (2013, p. 44) argued "as the drift-aligned section continues to grow at the expense of the stable swash-aligned section, macro-scale cannibalisation can be deemed to be ongoing. Whether micro-scale cannibalisation occurs again is unclear." Some key differences between the Rossbehy barriers and those of Nova Scotia are (1) the Nova Scotia barriers are gravel dominated, while the Rossbehy barrier is sand dominated and (2) Rossbehy is shifting from swash-alignment to drift alignment, while the opposite was observed in the case of the Nova Scotia barriers. Orford *et al*. (1996) did state that it was unclear whether or not the phases they described reflected an evolutionary trend.

2.1.3 Barrier breaching

Barrier breaching in 2008 significantly altered the dynamics of Rossbehy and its associated ebb-tidal delta. As such, the nature and role of barrier breaching is examined here.

Breaching occurs when a new opening forms across a barrier allowing water to flow freely between the water bodies on either side (Kraus, 2003; Kraus and Wamsley, 2003). If tidal exchange is weak and longshore sediment transport

strong, a breach may infill quickly. If not (which may be the case for Rossbehy), it may become a more permanent feature (*e.g.* a new tidal inlet).

Barriers can either be cut from their seaward or back barrier bay side, although the latter is more common. Breaching from the back barrier bay side usually occurs during storm events, when a surge combined with high waves causes water levels to increase to heights capable of overtopping the barrier. Once this occurs, elevated bay waters flow across the barrier, gradually incising the barrier and cutting a channel. Breaching from the seaward side occurs as a result of direct attack by waves. In this case, wave energy gets concentrated along a very narrow part of the shore. If conditions allow, the storm wave may overtop the beach ridge or foredune, resulting in flow toward the back of the lagoon. If frictional losses are minimal, available energy may be sufficient to cut a channel (Pierce, 1970). Analysis of aerial photographs suggests that breaching at Rossbehy was attributable to forcing from the western (seaward) side (O'Shea and Murphy, 2013).

Once established, breaches can be characterised based on their basal level relative to mean water levels. For example, Hartley and Pontee (2008) have distinguished between three different types of breach based on breach depth relative to mean high water spring (MHWS) and mean low water spring (MLWS). While the classification was developed for gravel barriers, it has been deemed appropriate for sand barriers because it doesn't depend on grain size (Sala, 2009).

According to Hartley and Pontee (2008), a level 1 breach (figure 2.7) is defined as one in which the basal level is near MHWS. If sediment supply from longshore drift is sufficient, a level 1 breach will infill. If not, it can deepen to form a level 2 breach. A level 2 breach occurs when the basal level of the channel is near MLWS. This type of breach may be more permanent than level one, but could still infill or deepen to form a permanent inlet. A permanent inlet occurs when the basal level of the channel is at or below MLWS and tidal exchange occurs on a daily basis. Permanent inlets form when tidal exchange is strong and longshore sediment transport weak (Kraus and Hayashi, 2005).

As indicated by Hartley and Pontee (2008), breach development is usually not a linear process. For example, during initial breach growth, tidal exchange and river discharges may widen the initial opening, but it may subsequently infill. At Rossbehy, Sala (2009) observed a shift from a level 2 breach in the winter of 2009 to a level 1 breach in the summer of 2010, which she interpreted as a prelude to a possible complete infilling. However, semi-diurnal tidal exchange has been observed on numerous occasions since under fairweather conditions (most recently, in June 2016). This suggests the new inlet may be a more permanent feature. Given the strong tidal exchange at the site and the limited influence of longshore transport due to the swash-aligned orientation of the (main section) of the barrier, this might be what one would expect.

Breaching is the most common mechanism by which new tidal inlets form today (this – and the other mechanisms of inlet formation - are discussed in section 2.3.2). The economic and environmental consequences of breaching, which include the reduction or loss of protective natural beach and dunes, loss of property by flooding, wave attack, and erosion, and potential loss of access to property, can be devastating (Kraus and Wamsley, 2003). As a result, a research priority has been the development of predictive models of breach risk, growth and stability. A good first estimation of breach stability is the inlet throat area-tidal prism relationship first recognized by O'Brien (1931):

$$A = CP^n \tag{1}$$

Where:

- A = minimal cross-sectional area of inlet below MSL (m^2)
- C and n = linear regression coefficients
- P = tidal prism (m^3)

The tidal prism is the volume of water entering and leaving the inlet on the flood and ebb flow. It can be roughly estimated by multiplying the area of the back bay by the tidal range or, if there are extensive tidal flats, from tidal current and cross-sectional area measurements. Its size varies with the system's hydraulic efficiency, or the ability of the system to transport water between the ocean and the bay. Systems with large inlets and short channels are more hydraulically

efficient because the effect of hydraulic friction is reduced. When breaching occurs, hydraulic efficiency increases, resulting in a larger tidal prism and enhanced exchange of water (Giese *et al*., 2009). According to the O'Brien relationship, breaches can only be maintained (*e.g.* turn into new inlets) if the cross-sectional area is large enough to accommodate the tidal prism or the tidal prism small enough to maintain the cross-sectional area. Studies have revealed that the O'Brien relationship exists for stable inlets all over the world, and it is therefore an important first-order determinant of breach stability.

Another important relationship describing breach stability is that of Escoffier (1940). Escoffier (1940) proposed the inlet closure curve (figure 2.8), in which in which the maximum flow velocity in the tidal inlet, $\hat{U}$, is a function of the cross sectional flow area, A. Within this model, an equilibrium velocity, $\hat{U}_{eq}$, exists, which is defined as the flow velocity required to maintain the channel. The parameters A_1 and A_2 are the stability thresholds. When $A<A_1$, $\hat{U}<\hat{U}_{eq}$. In this case, the breach closes because the maximum flow velocity isn't sufficient to maintain transport. When $A_1<A<A_2$, there is sufficient energy to transport sediment, but whether or not the channel will become stable depends on the cross sectional flow area. For it to become a stable inlet, sediment must be removed from the channel until $A>A_2$. When this occurs, $\hat{U}<\hat{U}_{eq}$, sediment transport becomes too weak to result in inlet closure, and the inlet becomes stable.

O'Brien's and Escoffier's relationships are only valid for single inlet bay systems, and therefore may not be applicable to Rossbehy, where a main inlet already existed prior to breaching. This is known as a multiple inlet bay system. Many studies suggest that multiple inlet bay systems are unstable (*e.g.* if a new inlet forms, one of the inlets has to close). Stable multiple inlet systems do, however, exist. Examples of stable multiple inlet systems have been documented at Gasparilla Sound, Florida (Escoffier, 1977), Chatham, Massachussetts (Friederichs *et al*., 1993; Liu and Aubrey, 1993), the Dutch Wadden Sea (Louters and Gerritsen, 1994), and Ria Formosa, Portugal (Salles *et al*., 2005). The characteristics of existing multiple inlet systems are examined in section 2.3.1. The causes and conditions that determine stability at multiple inlet systems are still unclear (Pacheco *et al*., 2010).

A final relationship of practical importance related to barrier breaching is the ebb tidal delta volume-tidal prism relationship (also known as the Walton and Adams relationship). This may be used to determine how nearby beaches will respond when a tidal inlet is formed due to storm breaching. Walton and Adams (1976) showed that a strong correlation exists between the magnitude of the tidal prism and the volume of sand in an inlets ebb tidal delta. Essentially the relationship says the larger the tidal prism, the more sand contained in the ebb tidal delta – *e.g.* the volume of sand in the delta is controlled by the tidal prism.

Growth of an ebb-tidal delta immediately following breaching has implications for adjacent shorelines because the additional sand diverted to the delta is removed from the longshore transport system, thus resulting in erosion of the downdrift shoreline (Davis and FitzGerald, 2004). This resonates with the situation at Rossbehy, where breaching has been accompanied by the expansion of the swash bar fronting the barrier. However, high rates of shoreline erosion are being experienced on the updrift (and not downdrift) side of the inlet.

In summary, breaching has resulted in the formation of a new multiple inlet system at Rossbehy. Recent observations of semi-diurnal tidal exchange at the new inlet suggest the feature may be more permanent than previously thought. The post-breaching evolution of adjacent barrier dunes at Inch and Rossbehy is examined in chapter 7.

2.2 Coastal Dunes

Sand dunes represent an important sediment store on many beaches, as evidenced at Inch and Rossbehy. Their formation is primarily a function of sediment supply and onshore wind, although sediment supply is usually the limiting factor (Davis and FitzGerald, 2004). As such, an examination of dune morphologies and processes is relevant here.

Aeolian processes are responsible for dune building. Wind-blown beach sand accumulates on the back beach when any type of obstruction, such as driftwood or or vegetation, blocks it from moving any further. If the accumulation is above the high water mark, it may accrete to form embryo or coppice dunes, small mounds held together by vegetation. If they continue to accrete uninterrupted,

they may coalesce into a foredune ridge to become a barrier dune. Barrier dunes are the first line of protection against storm induced erosion and flooding, and are therefore locally important to many coastal communities.

2.2.1 Morphology

The variety of dune morphologies has been described extensively (Nordström *et al*., 1990; Hesp, 2002; Martínez and Psuty, 2004) and is related to sediment supply and aeolian and hydrodynamic processes. Sloss *et al*. (2012) provide a comprehensive description of dune morphologies, which is briefly summarised as follows, with examples shown from Inch and Rossbehy.

A general classification of coastal dunes into primary and secondary dunes can be made, where primary dunes are defined as dunes composed of sand blown directly from the beach face and secondary dunes are those that develop as a result of the subsequent modification of primary dunes (Doody, 2012). Primary dunes make up the barrier dune – the first landward sand dune formation along a shoreline. These develop adjacent to the beach above the high water mark, and they are thus significantly modified by wave action. There are three general types of foredunes: incipient, established, and relict. Incipient, or embryo dunes, are low accumulations that form immediately above high spring tide (figure 2.9). Incipient dunes can be ephemeral, grow into established foredunes, or become relict and stable as new incipient foredunes develop seaward (a process known as progradation). Established foredunes (figure 2.10) tend to form at the rear of dissipative beaches. These can accrete to significant heights (>20 m), but high dunes are at risk of an increased likelihood of blowouts (figure 2.11), erosional features resembling saucers or trough shaped hollows. Established foredunes may eventually become isolated from their sediment supply when incipient foredunes develop seaward and accrete, replacing the former ridge and rendering it relict and stable. Relict dune ridges mark former shoreline positions and many of those that exist today, such as those at Inch (figure 2.12), formed as a result of Holocene sea-level change. These may also represent the limits of extreme events.

The modification of primary dunes by aeolian processes results in the development of secondary dunes. Secondary dunes include blowouts, parabolic

dunes, and transgressive dunefields, although parabolic dunes and transgressive dunefields may be classified as primary dunes if they exist adjacent to the beach. Parabolic dunes (figure 2.13) are U-shaped dunes that form as a result of continued transport of sand through blowouts. Their trailing arms are often anchored by vegetation. Transgressive dunefields are formed by the downwind and/or alongshore movement of sand over vegetated to semi-vegetated terrain. They are highly mobile and occur frequently in temperate humid climates where there is sufficient sediment supply and powerful onshore winds.

2.2.2 Relationship to sediment supply

Dune morphology is closely related to sediment supply and dune-beach exchange (Martínez and Psuty, 2004), and therefore warrants attention. The relationship between sediment budget and dune morphology has been described conceptually by Psuty (2004) and is illustrated in figure 2.14. In this model, foredune development is a function of the erosional status of the shoreline. Under positive beach budget, the dimensions of the foredune are inversely proportional to the rate of beach accretion. This is because high rates of progradation do not allow sufficient time for sand transport to the foredune. Maximum development occurs under a slightly negative beach budget. This is because at this point, scarping of the dune occurs, freeing up sand for transport on the seaward face. This, however, eventually leads to transfers inland (*e.g.* sand is removed entirely from the system), and results in dissection, blowouts, and parabolics. As the influence of inland transport wanes, the dimensions of the foredune decrease until it becomes a washover/sand sheet. This model represents a continuum of development, over which the sequence of forms is entirely dependent upon the changing combination of beach and foredune sediment budget.

Sediment supply to the foredunes is controlled by the sediment budget of the barrier complex as a whole. As such, if the barrier complex is suffering from a restricted sediment supply, this would be reflected in the foredune morphology. Presently this is the case at Rossbehy, where erosion of the foredunes in the drift aligned section of the barrier is taking place as part of the cannibalisation process. This material is not being returned to the dunes, but accumulating on

the ebb-tidal delta and, to a lesser extent, at the distal end of the barrier in the form of a minor drift-aligned shoreline spit.

2.2.3 Dune erosion

Dune erosion is a natural process triggered by waves during storm events. During storms, water levels may rise as a result of tides, wind- and wave-induced forces (setup), and surge, the combined effect of which is attack by incoming waves (van Rijn, 2009). Material eroded from the foredune is dragged down the slope by the downrush, which ultimately undermines the dune toe (a process called wave undercutting). Eventually, this will destabilise the upper part of the foredune, causing slumping, which can be eroded again by wave-induced processes. The material eroded from foredunes is deposited on the beach and in deeper water, forming an offshore bar. Figure 2.15 shows graphically typical examples of pre- and post-storm beach profiles. The post-storm foredune is steeper than the pre-storm foredune, with a wider and higher beach, a result of the redistribution of sediment from the foredune (Van Thiel de Vries, 2009). Eventually, fair-weather waves and swell return the sediment shoreward, where it can be remobilised by wind during the process of dune recovery. At Inch, the seasonal exchanges between the foredunes and offshore bars is generally balanced, at least in the short term, which means the foredunes have remained relatively stable in terms of their position. In the drift-aligned zone of Rossbehy, erosion and recovery of foredunes is more complicated due to barrier breaching, and eroded material from the foredunes is not being returned.

Various mechanisms are responsible for dune erosion. Based on field observations, Nishi and Kraus (2001) describe four such mechanisms (shown graphically in figure 2.16):

a) **Layer separation and collapsing** – occurs when a near-vertical dune face is subjected to wave impact. Repeated wave impact causes vertical fault lines (cracks) to develop. The outer layer gradually separates from the landward portion of the dune, eventually becoming unstable and collapsing suddenly (figure 2.17).
b) **Layer separation and overturning** – as above except the dune crest is overturned rather than collapsing.

c) **Notching and slumping** – occurs when wave attack results in a notch being cut into the base of the dune, causing it to eventually become unstable and collapse. Notching and slumping tend to occur on near-vertical dune slopes, slopes held in place by roots, compacted slopes, or slopes composed of rocks.

d) **Sliding and flowing** – occurs on gently sloping (close to the angle of repose) dunes when modest wave impact at the base of the dune causes a thin layer of sand to slide down the slope.

The mechanisms are important because they are each responsible for delivering different amounts of sand to the beach, thus modifying the beach profile response in various different ways. Once the foredune is eroded though, it can take considerable time (months or even years) to be returned.

2.2.4 Post-storm dune recovery

Post-storm dune recovery refers to the processes that a dune or dune system goes through to return to its original morphological, sedimentary, and/or ecologic state prior to a major storm event. There are a number of controls on recovery, including:

- sediment supply (important at Rossbehy),
- orientation with respect to wind and wave climate,
- dune height and pre-storm morphology (Priestas and Fagherazzi, 2010),
- degree of storm damage,
- subsequent storm magnitude and frequency (Houser and Hamilton, 2009),
- the prevalence of drift line material (Gerlach, 1992) and
- time between events (Priestas and Fagherazzi, 2010).

While in some cases, recovery can take several years (Thom and Hall, 1991; Morton *et al.*, 1994), in others, it may never happen at all.

An early simple conceptual model of post-storm recovery was proposed by Carter *et al.* (1990). In this model, three phases associated with erosion of dunes by waves are identified: undercutting, slumping, and the eventual reforming of the dune face (figure 2.18). During storms, basal undercutting of the foredune by

wave action (phase 1) induces slope failure (phase 2). This results in scarping and slumping of the dune face (figure 2.19). During the beach/dune recovery cycle (phase 3) sediment eroded during scarping is gradually returned to the slope face in a process initiated by echo dune formation between the high water mark (HWM) and the dune scarp. Later, wind-blown material may accumulate at the crest and/or on the mid-slope. Once crestal deposition reaches some threshold, slope failure occurs due to overloading, and material is redeposited on the lower slope and upper beach. Wind-blown material from these deposits gradually accumulates up-slope and eventually covers the scarp. Once sufficient accumulation has occurred such that the slope face is lying below the angle of repose (32° to 43° for loose sand grains), vegetation can begin to colonise the dune face. Where vegetated slump blocks occur, plants may establish new roots within the surrounding and underlying material, accelerating the recovery process. Evidence of this can be found at Rossbehy (figure 2.19) and may be a key recovery mechanism here.

2.3 Tidal inlets

A tidal inlet may be defined as "an opening in the shoreline through which water penetrates the land thereby providing a connection between the ocean and [back barrier] bays, lagoons, marsh, and tidal creek systems" (FitzGerald and Buynevich, 2003). Crucially, tidal exchange is required to maintain the channel. The volume of water that enters and leaves the inlet on the flood and ebb flow is known as the tidal prism. This is a function of the open water area in the back barrier (approx. 135 km^2 for Castlemaine Harbour) and the tidal range (MHWS = 3.76 m ODM for Inch). It can be estimated either by multiplying the area of the back bay the tidal range or, for coasts where there are extensive back barrier tidal flats, from tidal current and cross-section measurements (Davis and FitzGerald, 2004). This section describes the morphology of idealised inlets versus that of the Inch-Rossbehy system. It goes on to describe theories relating to the formation and evolution of inlets. The formation and evolution of the Inch-Rossbehy system is described in Chapter 3.

2.3.1 Morphology

Tidal inlets can be located between two barriers, a barrier and a bedrock headland, or a barrier and a glacial headland. The sides of inlets are commonly formed by the recurved edges of spits or barrier islands. These consist of sand deposited by refracted waves and flood tidal currents. Figure 2.20 shows the morphological components of an idealised tidal inlet. In this model, the inlet separates a recurved spit from a barrier island. The inlet throat is the deepest part of the inlet and is usually located where the bordering barriers constrict the inlet channel to a minimum width and minimum cross-sectional area. This is also where tidal currents reach their maximum velocity. The orientation of the main channel in the back barrier basin controls the initial direction of ebb outflow into the inlet throat (Elias and van der Spek, 2006).

When inlet tidal currents exceed the effects of longshore tidal currents, the onshore/offshore movement of sediment results in the deposition of ebb- and/or flood-tidal deltas. These are depositional horseshoe (flood-delta) or lobate (ebb-delta) features characterised by sand shoals and tidal channels. Flood-tidal deltas form on the landward side of the inlet as a result of the lateral expansion of currents, which lose their velocity and therefore deposit sediment. Ebb-tidal deltas form on the seaward side of the inlet, where they are more vulnerable to modification by wave action. The source of sediment for both types of deltas can be from erosion of the main channel and/or derived from longshore drift. The main components of an ebb-tidal delta are (see also figures 2.20 and 2.21):

- the main ebb channel - which is shaped by ebb currents;
- swash platforms – located between the main channel and the adjacent barriers;
- channel margin linear bars – linear wave formed bars on either side of the ebb channel;
- swash bars – wave formed bars located on the swash platforms behind the channel margin linear bars;
- marginal flood channels – channels separating the bars from the adjacent barriers; and

- the terminal lobe – the most seaward feature of the ebb-delta – here, flows decrease, resulting in deposition, and waves break during storms or at low tide, resulting in sediment suspension.

The morphology of ebb-tidal deltas is controlled by the relative effects of longshore drift versus tidal currents. When the influence of longshore currents exceed the effects of tidal currents, swash bars and channel margin linear bars tend to align themselves with the predominant direction of longshore drift. If inlet tidal currents exceed the effects of longshore currents, the bars align themselves in the direction of the inlet tidal currents (Oertel, 1988). This is the case for the main inlet at Inch/Rossbehy.

While most inlet systems are single inlet systems, multiple inlet systems do exist, as mentioned previously in this chapter. Two examples of these systems and their morphological characteristics are outlined as follows:

- **Nauset, Massachussetts** (figure 2.22): Nauset spit is one of a series of spits and barrier islands that make up a 30-km long chain on the southern outer coast of Cape Cod, Massachusetts. The southern end of the spit is segmented due to several storm breachings, the most recent of which occurred in 2007 (Giese *et al*., 2009). Presently, tidal exchange is through two tidal inlets – South Inlet (established in 1987) and North Inlet (established in 2007). Bedforms and shoals associated with formation of the two inlets have been documented by FitzGerald and Pendleton (2002) and Giese *et al*. (2009), amongst others. Following the formation of South Inlet, sand washed into the back barrier was reportedly reworked into shoals and bedforms and the associated ebb-tidal delta grew in volume. The decreased sediment supply as a result of the interruption in longshore drift caused significant shoreline recession downdrift (to the south). Following the formation of North Inlet, an extensive flood tidal delta developed landward of the inlet as well as a smaller ebb delta seaward of the inlet.
- **Ria Formosa, Portugal** (figure 2.23): The Ria Formosa system is a chain of 5 barrier islands (Barreta, Culatra, Annona, Tavira and Cabanas) and two barrier spits (Ancao and Cacela) separated by six inlets (New Ancao

Inlet, Faro Inlet, Annona Inlet, Fuzeta Inlet, Tavira Inlet, and Cabanas or Cacela Inlet). The inlets link a large tidal lagoon with the open ocean. The lagoon is characterized by salt marshes and sand flats (including overwash fans, minor recurved spits, and flood tide deltas), and a complex of lagoonal channels. Ebb-tidal deltas front each of the inlets, except at Faro, where the steep offshore bathymetry has prevented the development of an ebb-delta (Salles, 2001).

These examples suggest that the morphological features of multiple inlet systems are similar to those of single inlet systems.

The main inlet separating Inch and Rossbehy displays some of the typical morphological features discussed previously. However, the system differs from the idealised morphologies in some important ways. Firstly, Inch and Rossbehy are (predominantly) swash-aligned spits, whereas idealised models of tidal inlets put them between a spit and a barrier island or two barrier islands. The *de facto* assumption within the idealised model appears to be that barriers are located on open coasts, where there is a dominant direction of longshore transport. This is not the case for Inch and Rossbehy, which are located within a narrow, swash-aligned embayment whereby onshore-offshore transport dominates. Second, the main inlet separating Inch and Rossbehy extends into Castlemaine Harbour north of Cromane, where the Rivers Maine and Laune drain. The Caragh River also drains into the bay via a series of tidal channels behind Rossbehy. Many idealised models of inlets often do not include rivers and flanking systems. Rivers, especially the Caragh, may be important at Rossbehy. This is because river flow may affect the growth and stability of the breach (Kraus *et al*., 2002). For example, during times of unusually high precipitation, flows from rivers and streams could promote channel scour on the ebb flow. Alternatively, rivers could also act as a source of sediment for breach infilling. Finally, while the Inch/Rossbehy system shares some morphological features of the multiple inlet systems described previously, the process controls on the systems differ. Conceptual models of multiple inlet systems have been described by various authors, most notably in Aubrey and Giese's (1993) volume *Formation and Evolution of Multiple Tidal Inlets*. One such model is examined in the following section.

2.3.2 Formation and Evolution

Prerequisites for the formation of tidal inlets are (1) an embayment and (2) barriers. Inlets can form either (1) as a result of breaching; (2) as a result of spit building across a bay; or (3) as a result of the drowning of river valleys (Davis and FitzGerald, 2004).

As mentioned previously, breaching is the most common mechanism by which tidal inlets form today. Breaching can result in the formation of ephemeral inlets (usually opened during large storms and subsequently infilled quickly) or permanent inlets. For a new inlet to be established through breaching, subsequent tidal exchange between the ocean and the bay must be sufficient to maintain the channel. Breaching is commonly responsible for the formation of dual or multiple inlet systems, as in the case of Nauset spit and Ria Formosa. Giese (1988) and Giese *et al*. (2009) proposed a conceptual model for the formation of dual inlets based on observations of the historical behaviour of Nauset Spit. This model describes a cyclic, two-phase evolution, in which an inlet development phase prompted by breaching is followed by an inlet migration phase, characterised by spit elongation and an eventual return to a single inlet system. The inlet remains stable until a new breach is formed, thus initiating repetition of the cycle. Some key differences between the dual inlet systems of Nauset spit and Rossbehy are the open coast versus narrow embayment orientation of the spits, respectively, and the fact that the main inlet channel at Rossbehy appears to be geologically fixed (*e.g.* non-migrating) (O'Shea and Murphy, 2013).

A second mechanism for inlet formation is spit building across a bay (Davis and FitzGerald, 2004). This usually occurs early in the evolution of a coast. As spit growth ensues, the opening to the bay gradually decreases in cross sectional area. As the cross sectional area decreases, there is a coincident increase in tidal flow. Current velocities must therefore increase. The tidal inlet is formed once the bay reaches a stable configuration (*e.g.* cross-sectional area is balanced by tidal prism). This may be the mechanism by which the main inlet between Inch and Rossbehy formed. Further discussion on this is provided in chapter 3.

A third mechanism for tidal inlet formation is the drowning of river valleys (Davis and FitzGerald, 2004). In this case, rising sea-level gradually floods a former river valley. Spit and/or barrier island growth ensues until it reaches the paleochannel, where tidal currents scour sediment vertically, removing sediment from the channel. Often the valley fill sediment is less resistant that the sediment making up the valley walls, which facilitates this removal by tidal currents through the inlet channel. The growth of the barrier is also limited by this deepening inlet channel. The channel becomes an inlet when tidal currents passing in and out control overall sediment transport and the dimensions of the inlet throat (*e.g.* cross sectional area) are controlled by the tidal prism.

Tidal inlets may be migrating or stable. Inlets greater than 8 m depth are usually stable, because sediments on either side are more resistant to erosion. Shallower inlets (<3-4 m) are more likely to migrate because the inlet channels are usually eroded in sand, making migration easier (Davis and FitzGerald, 2004). Migration occurs when longshore transport of sand results in deposition on one side of the channel. This restricts flow through the channel, allowing tidal currents to scour the channel. The overall effect is that the channel moves downdrift. Some inlets may migrate updrift. This occurs where a major back barrier tidal channel approaches the inlet at an oblique angle. If the bend in the channel is on the updrift side of the inlet, ebb currents directed there may result in erosion and migration updrift. Inlets that migrate updrift are usually small to moderately sized and occur along coasts with small to moderate net sand transport rates (Davis and FitzGerald, 2004).

2.4 Morphodynamics

Morphodynamics refers to the study of the dynamic interactions between hydrodynamic processes and the morphology of the seabed over a range of time-space scales (Short and Jackson, 2013). While this definition is widely cited, it should be noted that beach morphodynamics also includes the study of aeolian processes and landforms, such as coastal dunes, as dune systems represent an important sediment store within the sediment budget of beaches where they exist (Anthony, 2008). The morphodynamic approach to the study of beaches is based on the process-response model – a fundamental relationship that describes the

processes of wind, waves, and currents acting upon sediment to produce coastal landforms (Psuty, 2008). The basic process-response model for beaches of adequate sediment supply is the beach profile (figure 2.24), extending from the offshore bar, through the surf and swash zones, and into the coastal foredune (Short, 1999). Within this system, sediment is episodically stored and released in a dynamic flux regime. Sediment is delivered first to the beachface by nearshore (surf and swash) processes (Houser, 2009). Then, during low tide, when the foreshore is dry and vulnerable to wind action, beach sand is lifted and transported by aeolian processes to the backshore zone, where dune formation occurs (King, 1972; Bird, 2011). During storm events, strong waves rework the foredune and transport sediment back to the beach and offshore. Coastal sand dunes thus act as a sand reservoir, intermittently receiving, storing, and releasing excess beach sand, resulting in a range of beach morphologies. These and other processes that control nearshore morphology on barrier beaches are described in more detail in the following sections.

2.4.1 Coastal zone hydrodynamics

Hydrodynamics can be simply described as the mathematical description of fluid motions in nature (Svendsen, 2006). In the context of coastal geomorphology, an understanding of hydrodynamics is essential to describe sediment transport, and in turn, resulting changes in beach morphology, in response to the passage of ocean waves. Some basic hydrodynamic concepts are outlined as follows.

The nearshore zone

In the cross-shore direction, the seaward limit of the nearshore zone is marked by the point where the wave base begins to interact with the seabed (L/2 > depth; L = wavelength) and wave heights start to increase to conserve energy flux (wave shoaling). Figure 2.25 shows schematically important nearshore wave processes, which are described as follows:

Refraction

Refraction is the bending of wave crests as a result of variation in the underlying bathymetry. This occurs because the portion of the wave in shallower water moves slower than the portion in deeper water, thus when wave crests approach

the shoreline, they adjust to reflect the underlying bathymetry. At the study location, this is important because refraction over nearshore shoals can change the direction of incoming waves with respect to the shoreline, potentially leaving the foredunes vulnerable to wave attack.

Diffraction

Diffraction is the sideways bending of wave crests when they encounter an obstacle, such as a jetty, headland, or island. At the study site, diffraction around shoals at low tide may affect the direction of sediment transport by waves.

Reflection

Reflection occurs when waves bounce back from an obstacle, such as a jetty or a steep rocky shore. When reflected waves interact with oncoming waves, they form standing waves. These are important for offshore bar formation (Davidson-Arnott, 2010). Offshore bars are important at the study site because they represent an important sediment store.

Shoaling

Shoaling occurs when waves enter a depth at which the wave base begins to interact with the seabed, resulting in an initial increase in wave height. In this way, energy flux is conserved and wave frequency remains constant. The location in which waves begin to interact with the seabed is important because it marks the zone in which sediment transport by waves begins.

Breaking

Wave breaking is the process by which the surface of a wave folds or rolls over and intersects itself, causing large amounts of wave energy to be dissipated (Babanin, 2011). Waves may break as a result of steepening during their propagation into shallow water, interaction with other waves, or through the input of energy from wind. Galvin (1972) referred to breakers as either spilling (surf rolls gently over the front of the wave; occurs on beaches with relatively gentle slopes), plunging, (wave curls over, forming a tunnel; occurs on beaches with moderately steep slopes), collapsing (bottom face of the wave steepens and collapses, but crest never fully breaks; occurs on beaches with steep slopes), or

surging (wave rolls onto beach retaining a fairly stable shape; occurs on beaches with steep slopes).

Wave shoaling and breaking are important on the terminal lobe of ebb tidal deltas because they lead to deposition on the swash platform, channel margin linear bars, and ebb shoals.

Longshore currents

Currents travelling parallel to the shoreline are longshore currents. Longshore currents are the mechanism by which longshore drift of beach sediments occurs. They are generated by waves orientated obliquely to the shoreline (Longuet-Higgins, 1970). Longshore currents move water and sediment parallel to the beach in the direction of wave approach. Their ability to move sediment depends on various factors, including beach length, wave height, wave velocity, orientation of incoming waves, and the relative strength of tidal currents. At Inch and Rossbehy, the strength of longshore currents decreases as you move toward the tidal inlet, where tidal currents dominate transport.

Surf and swash zone processes

Landward of the initial breaker zone, surf and swash zone processes play an important role in nearshore morphodynamics. This is because the shallow water depth along with the turbulence induced by wave breaking enhances the amount of sand in suspension. In addition, longshore currents and undertow have a significant impact on sediment transport in these zones (Fredsøe, 2002).

The surf zone extends from the initial breaker zone to the shoreline, with the swash zone being a sub region of this zone located landward of the line of breakers. The swash zone is characterised by alternate wetting and drying by swash uprush and backrush. Important processes that occur in the surf zone have been summarised by Schwartz (2006) and are briefly outlined as follows:

1. **Creation and breaking of a wave roller** – A roller is a "region of highly turbulent, aerated water that moves at roughly the wave celerity" and forms at the surface at onset of wave breaking (Schwartz, 2006; p. 930).

Rollers are responsible for dissipating energy and are important for balancing energy, mass, and momentum.

2. **Maintenance of residual turbulence in the water column** – Residual turbulence is the turbulence left behind by the wave roller. This expands downward through the water column and is important as it plays a role in the suspension of sediment and the mixing of surf zone currents.
3. **Creation of set-up in the mean water level** – In the surf zone, currents can be driven by gradients in radiation stress components. As a wave and roller shrink, radiation stress in the cross shore direction decreases and a "set-up" (increase) in mean water level is created. Winds can also affect local wave induced set up.
4. **Generation of currents (cross-shore, longshore, rip currents) and low frequency motions** – Cross-shore currents (undertow), longshore currents, rip currents, and low frequency motions, including surf beat, edge waves, and shear waves, occur in the surf zone and represent important sediment transport mechanisms.
5. **Entrainment and suspension of sediment** – Entrainment of sediment in the surf zone depends on both the sediment characteristics (particle size, density, etc.) and wave and current characteristics (wave height and period, current strength). Sediment concentrations in the surf zone are generally much higher than those outside the surf zone due to turbulence induced by wave breaking.
6. **Sediment transport** – Changes in beach morphology result from cross-shore and longshore changes in spatial gradients in sediment transport. These variations occur as a result of oscillatory wave motions, mean currents, the presence of obstructions such as jetties or groynes, and bathymetry (important at the study site, where variations in the morphology of the channels and shoals in the ebb-tidal delta affect transport).

2.4.2 Sediment transport processes

Sediment transport describes the movement of sediment in a fluid (the fluid usually being either air or water). Whether sediment will be eroded, transported, or deposited depends on bed characteristics (slope), particle characteristics (size,

density, etc.) and flow characteristics (characteristics of the fluid, laminar vs. turbulent flow, etc.).

Perhaps the earliest modern studies of sediment transport were undertaken by Filip Hjulström, a Swedish geographer who described the thresholds for erosion and deposition of particles in running water based on experimental data in the form of the now famous Hjulström diagram (Hjulström, 1939; figure 2.26). The Hjulström diagram is a log-log curve showing critical erosion and deposition velocities as a function of particle size. The curve has undergone modifications, most notably by Sundborg (1956), to reflect variations in flow depth. While developed to describe fluvial transport, the Hjulström diagram is often cited in the coastal literature (*e.g.* Novak, 1972; Hearn, 2008; Davis, 2013; Harris, 2014). The diagram shows that for large particles (>fine-medium sands) grain size is proportional to the velocity required to either erode or keep particles suspended. Small particles (<0.1 mm), however, require greater velocities for entrainment to occur. This is due to grain-to-grain adhesion and electrostatic charges. The presence of cohesive sediment, such as clay and mud, thus increases resistance against erosion. It should be noted that while the Hjulström diagram illustrates these principles, it is only an approximation of the transportability of sediments, and is somewhat limited with regard to the prediction of sediment transport, especially in coastal environments where flow velocity is not constant and particle distribution is usually not uniform.

The following sub-sections describe some basic concepts related to sediment transport. Many of the formulae derived in the sediment transport literature form the basis of modern numerical modelling, which has been employed in this research, and thus, an introduction is appropriate here. It is important to note that many of the following formulations are based on empirical parameters, which were derived using experimental or laboratory data. This means that model results are often sensitive to calibration parameters specific to the location of interest, reinforcing the need for preliminary model validation studies particular to the area of interest (Kulkarni, 2013).

Entrainment

Three forces are responsible for sediment entrainment in a fluid: gravity, drag (or friction) and lift (or shear stress) (figure 2.27). The gravitational force (F_G) acts vertically downwards, opposing the lift (F_L) and drag forces (F_D). Theoretical expressions for these forces have been derived (*e.g.* Hardisty, 1990) such that the critical flow conditions, or conditions resulting in the displacement of a sedimentary particle, can be calculated given specific sediment and fluid characteristics.

Critical flow conditions are usually described in terms of the critical bed shear stress (the minimum force per unit area acting on the sediment surface required for entrainment) or the critical Shield's parameter (a nondimensionalisation of shear stress, τ) and the Reynold's number. Shield's parameter is given by equation 1:

$$\tau_* = \theta = \frac{\tau}{(\rho_s - \rho)gD} \qquad (1)$$

Where:

- τ_*=critical Shield's parameter (shear stress required for entrainment)
- τ / θ =a dimensional shear stress;
- ρ_s = the density of the sediment;
- ρ = the density of the fluid;
- g = acceleration due to gravity;
- D = characteristic particle diameter of the sediment

Bed shear stress is derived experimentally and depends on bed roughness and particle, fluid, and bed characteristics.

The Reynold's number is the ratio between driving (inertial) and resisting (viscous) forces (Ritter *et al*., 2006) and is given by:

$$Re_p = \frac{U_p D}{v} \qquad (2)$$ Where:

- U_p = particle velocity;
- D = grain size;
- v = kinematic viscosity (dynamic viscosity divided by fluid density)

When plotted against the Reynold's number, the critical Shield's parameter gives the threshold curve for sediment motion under unidirectional flow conditions (figure 2.28). This diagram (known as the Shield's diagram) separates conditions of established movement (above the curve) from conditions of no movement (below the curve). The diagram shown in figure 2.28 was updated with experimental data from Miller *et al*. (1977). It should be emphasized that the Shield's diagram is only valid for uniform flow on a flat bed. Kulkarni (2013, p. 18) points out "some effects such as bed ripples or the effect of the combination of unidirectional and oscillatory flow on initiation of motion are largely unknown."

Modes of transport

There are three modes of sediment transport in a fluid:

1. **Rolling or sliding (creep)** – particles roll or slide along the bed
2. **Saltation** – particles bounce along the surface
3. **Suspension** – the weight of moving particles is wholly supported by fluid forces (Hardisty, 1990)

Particles transported by rolling/sliding or saltation together represent bedload transport. Various formulae have been developed to describe suspended and bedload transport. The structure of most of the formulae is similar, with the differences between them mostly due to the fact that they were derived for specific situations and/or based on different sets of assumptions (e.g. uniform grain size). This is important because say, for example, a sediment transport formula takes into account only beach slope, sediment size and peak wave period. This formula would probably not be useful in places where the influence of tidal currents is pronounced, such as at Rossbehy. As such, the choice

between which is appropriate for use in a numerical model setting depends on the situation and is usually made based on prior experience or by trial and error (Kulkarni, 2013).

O'Shea and Murphy (2013) performed an analysis of sediment transport formulae using data from Rossbehy. They found that different formulae were more appropriate for different parts (swash versus drift aligned sections) of the barrier. These results were taken into account when it came to the set up of the numerical model used in O'Shea (2015) and subsequently used for this study.

Fall velocity

The fall (or settling) velocity (w_s) of particles refers to the velocity at which the force due to gravity exceeds fluid forces (*e.g.* the velocity at which deposition occurs – represented by the lower curve on the Hjulström diagram). This depends on both the particle and fluid characteristics. Because most particles are irregularly shaped, it is difficult to obtain an exact mathematical transcription of fall velocity. Van Rijn (1993) derived a formula for spherical particles with a diameter between 100 μm and 1000 μm. This is based on the kinematic viscosity (the ratio of absolute viscosity to density, ν) and other properties of the particle and water density and is given by equation 3:

$$w_s = \frac{10v}{d}\left[\left(1 + \frac{0.01\left(\frac{\rho_s}{\rho_w} - 1\right)gd^3}{v^2}\right)^{0.5} - 1\right] \quad (3)$$

Where:

- ν = the kinematic viscosity
- d = sphere diameter
- ρ_s = sediment density
- ρ_w = water density
- g = acceleration due to gravity

Fall velocity is incorporated into numerical models as part of a mass balance equation describing the transport of suspended sediment load.

The equilibrium beach profile and the Bruun Rule

The concept of an equilibrium beach profile was first described by Bruun (1954) and has been used widely in engineering applications. The equilibrium profile is an approximation of the cross-shore shape of the zone over which net sediment transport takes place. The term "closure depth" was introduced to describe the depth beyond which no net transport takes place. Based on an examination of field data from Mission Bay, California and the Danish North Sea coast, Bruun concluded that the cross-shore profile in the vertical direction for any sandy beach could be expressed as follows:

$$h = Ay^{\frac{2}{3}} \tag{4}$$

Where:

- h = water depth
- A = a sediment scaling parameter (related to grain size or fall velocity)
- y = cross-shore distance from the shoreline

This model has been widely criticised, most notably by Pilkey *et al.* (1993), who have cited fundamental issues with regard to the underlying assumptions of the model and its poor performance in several instances.

The equilibrium profile model has undergone various modifications over the years, for example by Dean (1991) and Rosati *et al.* (2013), but, despite its criticism, it remains popular today.

Bruun later built upon his equilibrium profile model and proposed the following model, known as the Bruun Rule (Bruun, 1962), which describes equilibrium shoreline retreat of sandy coasts as a function of SLR. It states that the equilibrium profile of an active beach will respond to rising sea-levels in such a way that the volume of eroded material from the upper shoreface will be equal to the volume of material deposited on the more extensive lower shoreface. It was described mathematically by Bruun (1988) as follows:

$$R = \frac{SL}{h + B} \quad (5)$$

Where:

- R = shoreline retreat
- S = SLR
- L = cross-shore width of the active profile (extends from the closure depth to the furthest landward point of sediment transport)
- h = the closure depth (maximum depth of sediment transport)[5]
- B = the elevation of the beach or dune crest (maximum height of sediment transport)

This model has also received significant criticism, notably by Cooper and Pilkey (2004), Pilkey and Pilkey-Jarvis (2007) and Stive (2004), due to its simplicity and restrictions and misuse in many engineering applications.

2.4.3 Sediment budget

Before defining sediment budget, it is useful to explain the concept of a coastal or sedimentary cell (also sometimes called a littoral cell). Van Rijn (2010, p. 6) describes coastal cells as "self-contained micro, meso or macro units within which sediment circulates with cycles of erosion and deposition." These discrete units can be thought of as closed systems, whereby there is no net transport into or out of the system. The boundaries of coastal cells are usually defined by the topography and shape of the coastline, although they may not necessarily be fixed (Goudie, 2004). The concept of coastal cells was first introduced by Bowen and Inman (1966), with examples given for a stretch of coast in southern California (figure 2.29). For each of the 5 cells shown in figure 2.29, Bowen and Inman (1966) identified sediment sources, sinks, and transport pathways and estimated to the best of their ability the relative magnitudes of sediment transport into and out of the cells. From this information, a sediment budget can be constructed. Sediment budgets describe the mass balance of inputs (sources) and

[5] Various authors have proposed alternative definitions of closure depth. For example, Krauss *et al*. (1988) state "the depth of closure (DoC) for a given or characteristic time interval is the most landward depth seaward of which there is no significant change in bottom elevation and no significant net sediment transport between the nearshore and the offshore."

outputs (sinks) for sedimentary cells within a specified period of time. Sources can include input by rivers, estuaries, cliffs, dunes, shelves, or artificial nourishment and sinks can include dead zones, depressions, canyons, mining, etc. Both include longshore transport, which is usually estimated using either the CERC formula (US Army Corps Of Engineers, 1984) or the Kamphius formula (Kamphuis, 1991). The sediment budget equation can be expressed in terms of volume or volumetric rate of change as follows (after Rosati, 2005):

$$\sum Q_{source} - \sum Q_{sink} - \Delta V + P - R = Residual \qquad (6)$$

Where:

- Q_{source} = source to control volume
- Q_{sink} = sink to control volume
- ΔV = net change in volume within the cell
- P = material placed in the cell (*e.g.* nourishment)
- R = material removed from the cell (*e.g.* dredging, mining, etc.)
- Residual = the degree to which the cell is balanced (for a balanced cell, residual=0)

Usually, sediment budgets are constructed by first developing a conceptual budget, then assimilating available data to validate this model. Oftentimes, however, accurate and/or precise data are not available for all sources and sinks, or, worse, a source or sink may not have been identified correctly. Some of the challenges associated with constructing sediment budgets highlighted by Rosati (2005) include delineation of cell boundaries, definition of possible sediment transport pathways, and accurately quantifying the relative magnitude of sources and sinks. Multiple sediment cells can be distinguished at Inch and Rossbehy. From wave modelling, Cooper *et al*. (1995) identified at least two cells fronting Inch. Post-breaching, Sala (2010) delineated four cells at Rossbehy. These are presented in chapter 3.

Coastal sediment budgets are strongly related to relative SLR at the meso-scale (10^1-10^3 years) (Van Rijn, 1998; Van Rijn, 2010). Studies suggest that long-term beach erosion as a result of accelerated SLR may eventually lead to the deterioration of coastal barrier islands (Williams *et al.*, 1992; FitzGerald *et al.*,

2007), with nearshore tidal deltas, capes, and the inner continental shelf acting as sediment reservoirs (sinks) (Komar, 1998). The sediment budget approach to predicting shoreline changes as a result of SLR is becoming an attractive alternative to the Bruun Model (Bruun, 1962) because it takes into account longshore sediment transport fluxes, which become important over long stretches of coast (Cooper and Pilkey, 2004; Davidson-Arnott, 2005; FitzGerald *et al.*, 2008).

3 The Inch-Rossbehy Barrier System

The Inch-Rossbehy barrier system and the adjacent estuary have been the subject of extensive geomorphological research, particularly since the 1990s (*e.g.* Cooper *et al.*, 1995; Devoy, 1995; Orford *et al.*, 1999a; Orford *et al.*, 1999b; Sala, 2010; Gault *et al.*, 2011; Delaney *et al.*, 2012; O'Shea and Murphy, 2013). Having been the focus of a number of large-scale EU funded projects (*e.g.* SEA LEVELS, IMPACTS, Conscience and CoastAdapt), the area has been branded a "coastal research laboratory" by Devoy (2015). Table 3.1 provides a brief overview of the research that has been undertaken here. Many of these studies, however, were conducted prior to breaching. As a result, our understanding of the impacts of storms on the behaviour of the system post-breaching is limited.

Inch and Rossbehy are interesting from a geomorphological perspective in that they are both (morphologically) representative of many neighbouring areas along the mid European North Atlantic margin (Delaney *et al.*, 2012) yet unique in their functioning and behaviour, at least in recent times (O'Shea and Murphy, 2013). The barriers are locally important in that they serve as recreational amenities, support a number of protected habitats (NPWS, 2010; NPWS, 2014), and protect low-lying land in the Castlemaine harbour area from erosion and flooding. Local concern over the future of the area following the breaching of Rossbehy in 2008 has grown, especially after significant damage was caused by the extreme back-to-back storms of 2013/2014 (Hickey, 2014). So, with relative sea-level in the local area projected to rise by between 45-50 cm (Lowe *et al.*, 2009) and extreme events expected to become more frequent and intense (Kiely *et al.*, 2005; Beniston *et al.*, 2007; Mori *et al.*, 2010), the question of "what is the future of the barrier?" has become all the more pertinent. This is a question that this research aims to address. But first, this chapter will set the local physical and morphodynamic context.

3.1 Local physical and paleoenvironmental setting

Inch (Irish: *Oiléan Inse*, "the island island") and Rossbehy (Irish: *Ros Beithe*, "peninsula of the birches") are north-south trending barriers located within a narrow, swash-aligned embayment (Dingle Bay). They are both connected to the mainland at their proximal ends and separated separated by a deep, narrow (approx. 2.5 km wide and 10 m depth) tidal inlet (Chapter 1, figure 1.1). The barriers are fronted by an extensive ebb-tidal delta (figure 3.1) and backed by a large back-barrier estuary (Castlemaine Harbour). They are generally comprised of coarse sediments derived from both fluvial and glacial sources. The barriers support extensive high dune systems, which are fronted by wide, flat dissipative beaches comprised of subaerial sands (D_{50} = 0.235 at Rossbehy). The dunes are probably founded on underlying cobble or gravel deposits (Carter *et al.*, 1989b; Devoy *et* al., 1995; Sala, 2010; Delaney *et al*., 2012), although further coring and geophysical work is required to establish the basal stratigraphy in detail. As a result of the swash-alignment of the embayment, there is no regional longshore drift component in operation at the study area.

Inch and Rossbehy are part of a wider sedimentary system within the inner part of Dingle Bay, which includes the gravel spit, Cromane, and Castlemaine Harbour to the east (see Chapter 2, figure 2.1). For the purpose of understanding transfers of sediment, this area can be compartmentalised into two basins: an inner basin, defined as the area between Cromane in the west and the mainland in the east, and an outer basin, defined as the area enclosed by the seaward sides of Inch and Rossbehy in the west and Cromane in the east (Cooper *et al.*, 1995). The inner basin is characterised by low-energy intertidal sand and mud flats and an extensive saltmarsh fringe. Two rivers – the Maine and the Laune - drain into the inner basin at the easternmost extent of Castlemaine Harbour. The outer basin is characterised by intertidal mud and sand flats. Here, approximately 1 km east of Rossbehy, is where a third river, the Caragh, drains. Fluvial and tidal exchange, mainly through the main inlet channel, facilitates sediment transport within and between these environments.

The sedimentary environments presented above lie within the wider area of Dingle Bay on the southwest coast of Ireland. The region is characterised by a series of rugged headlands separated from one another by narrow bays. The

peninsulas are anticlines composed of sandstones, greywackes, and conglomerates, and Carboniferous limestones occupy the intervening synclines (Guilcher *et al.*, 1960). Extensive, unconsolidated glaciogenic deposits are widespread along the contemporary coastline as well as offshore across the inner continental shelf, with Quaternary sediment deposits seaward of Inch and Rossbehy having been found to be up to 70 m thick (Shaw *et al.*, 1994). Sediment exchange between the bays in the region is minimal (Cooper and Jackson, 2003).

The floor of Dingle Bay is occupied mostly by sand and coarse sediment, with extensive bedrock outcrops present across the majority of the bay (figure 3.2). Figures 3.3 and 3.4 show the bathymetry of the bay and Castlemaine harbour, respectively. Water depths in the bay deepen from 0 m to 140 m offshore. A variety of active bedforms has been documented on the bay floor, including gravel ripples, sand waves, and ebb- and flood- orientated structures near the inlet channel between Inch and Rossbehy (Shaw *et al.*, 1994). The majority of the bay (up to the 10 m depth contour) was mapped in 2009 as part of the INFOMAR (Integrated Mapping for the Sustainable Development of Ireland's Marine Resource) project, a joint venture between the Geological Survey of Ireland and the Marine Institute which aims to create a range of "integrated mapping products of the physical, chemical and biological features of the seabed in the near-shore area" (INFOMAR, 2015a). The data for figures 3.2 and 3.3 are derived from this source and are freely available online.

3.1.1 Paleoenvironmental history and the influence of sea-level

Controls on barrier development occur across a wide range of time and space scales. The shape, size, and extent of Inch and Rossbehy today are, to a large extent, the product of the mutual interactions between long-term (10^3) changes in relative sea-level, long- to short-term (10^3-10^0) changes in sediment supply, and the magnitude, frequency, and history of storms. While this study focuses on micro- to meso-scale drivers of change at the field sites - *e.g.* storms and short-term (10^1-10^2 years) RSLR – a holistic view of the processes and morphological impacts of those processes that are known to operate/have operated in the vicinity of the study area at various spatiotemporal scales provides the context in which to understand the processes in operation today.

Inch and Rossbehy and their adjacent environs have been the subject of a number of paleoenvironmental studies, and a Holocene coastal record has been constructed for the area (Carter *et al.*, 1989a; Wintle *et al.*, 1998; Delaney *et al.*, 2012). Table 3.2 summarises the findings from various studies chronicling the local history from >5,000 years BP to 500 years BP. Prior to 5,000 years BP, when sea-levels were approx. 3-5 m below present, the present-day area where Inch and Rossbehy are located was a terrestrial-dominated environment (Shaw *et al.*, 1986; Delaney *et al.*, 2012). But during this time, a progressive marine inundation was taking place in Dingle Bay (Delaney *et al.*, 2012). As sea-level rose, sediment present on the valley floor at that time would have subsequently become available for redistribution. This sediment would become the source for the development of the sedimentary environments present in the bay today, including Inch and Rossbehy.

An explanation for the development of Inch and Rossbehy as longshore, prograding spits is complicated by their existence within an embayment in which no regional longshore sediment transport takes place. As a result, competing narratives exist with regard to their origins and development. These are summarised as follows.

Development as spits

The first to suggest that Inch developed as a spit was Guilcher *et al*. (1960). They suggested that a re-entrant in the present configuration of the coastline comprised of solid rock cliffs may have been responsible for the interruption of sediment transport eastwards, prompting spit development. Cooper *et al*. (1995) found evidence to support this in the form of a systematic reduction in the radiocarbon age of sediments from the proximal end of Inch (age=4448+/-36 BP) to the distal end (age=1553 +/-42 BP). They interpreted this as congruent with spit formation, although others (Wintle *et al*., 1998; Orford *et al*., 1999; Devoy, 2013) caution this younging southwards could simply represent the reworking of sediments north to south. Presuming the former is true, Cooper *et al*. (1995) argued that the spit terminus was in place in its current location by about 1500 years BP, which means that the tidal inlet separating Inch and Rossbehy had formed by this time. This would suggest that by 1500 BP, the Inch-Rossbehy system had reached a stable configuration such that the cross-sectional area of the inlet was balanced by the tidal prism (*e.g.* the system became hydraulically

efficient). This explanation of the initial development of Inch follows the general model described in chapter two of spit building across a bay. There is no evidence to suggest, however, that the proto-Rossbehy formed in a similar manner.

Cooper *et al.* (1995) went on to argue that between 2000 BP-1500 BP, the proximal neck of Inch would have had to have been some distance seaward of its current position. Following the establishment of the inlet, the development of the ebb-tidal delta would have depleted Inch of sediment, resulting in cannibalisation of its proximal end. Fixed by the inlet, the southern terminus of the barrier remained in place, while the proximal end retreated landwards, resulting in an overall clockwise rotation of the barrier. Evidence for this scenario is described as follows by Cooper *et al*. (1995, p. 3): "Sandwiched between the glaciogenic material and transgressive sands and gravels in front of Inch are basins of interbedded fine sediments that are suggested to be back barrier/estuary in origin and reflect the past-mobility of the contemporary Inch barrier as it migrated landward under Holocene sea-level rise."

This change in orientation from a drift-aligned, prograding spit to a swash-aligned barrier (its present configuration) meant that Inch went from being dominated by longshore transport to offshore and cross-barrier sediment transport. It has been suggested by Cooper *et al*. (1995) and MacClenahan (1997) that this shift ultimately facilitated dune development. Based on evidence from infrared stimulated luminescence (IRSL) dating of dune sands, Cooper *et al*. (1995) argue that the onset of the dune barrier formation (which now overlies the swash-aligned barrier) occurred relatively recently, from 600 years BP, with the main dune barrier having been built 300-200 years BP.

Cooper *et al*. (1995) speculated little on the coincident development of Rossbehy. Evidence reported by Delaney *et al*. (2012) suggests that at 6100 BP, Rossbehy was probably seaward of its present position. It is unclear, though, whether or not Rossbehy previously underwent a similar change in orientation to Inch. Its formation as a longshore prograding spit is, therefore, up for debate.

Development over an end moraine

An alternative to the above has been proposed by Devoy (2013; 2015) and Delaney *et al*. (2012). In this scenario, Inch and Rossbehy developed over a glacial recessional end moraine extending across the width of the bay. It is

generally agreed that the gravel "spit" Cromane rests on a recessional end moraine, but there is less evidence to support the idea that Inch and Rossbehy do as well. Delaney *et al*. (2012) posit the surface of the moraine(s) was initially wave eroded under mid- to late Holocene rising sea level. Sediment from the moraine was reworked into a series of cobble sized wave-refracted ridges (Delaney *et al*., 2012). These ridges form the bases of the modern day "spits." Such ridges are also evident on Inch, emerging from beneath the dunes or in the interdune swales (Devoy, *pers. comm*).

From c. 6000 BP to present, Delaney *et al.* (2012) posit the proto-barrier(s) progressively moved on-land. There is certainly evidence of barrier rollover at both Inch and Rossbehy (*e.g.* Cooper *et al*., 1995 and Delaney *et al*., 2012). In fact, at Rossbehy, exposures of peat on the beachface and beneath the dune sands were observed on multiple occasions near the hinge point between the swash-aligned beach and the drift-aligned beach during field visits from 2013-2015 (figure 3.5). The truncated upper contact of similar woody, monocot peat from a core in the back barrier saltmarsh has been dated to 2781-2000 BP (Delaney *et al.*, 2012), indicating that the dune barrier at Rossbehy may have attained is present position by that time.

In this scenario, the formation of the main inlet was not as a result of spit building across the bay, but as a result of barrier breaching c. 3000 years BP (Devoy, 2015). An abrupt change from intertidal marsh sediments to channelised sands and silts in cores obtained by Delaney *et al*. (2012) in the back barrier environment at Rossbehy was presented as evidence of this breaching event. Organic material from the underlying marsh sediments was dated at 2873 years BP. Devoy (2013) posits the dune barrier probably began to develop on top of the wave-built ridges after the formation of the inlet (between 2000-3000 years BP). Devoy (2013, p. 5) speculated "some sand dune cover may have [even] began to develop prior to the arrival of the moraine barrier system, both [at Inch] and at Rossbehy."

This scenario is radically different from that of Cooper *et al*. (1995) in the following ways:

- It suggests the barriers did not form as longshore, prograding spits, but instead from the onshore movement of swash-aligned ridges.

- It suggests the development of the main inlet was not as a result of spit progradation across Dingle Bay, but as a result of a breaching event 3,000 years BP.

This has important implications for the present research. If the barriers developed according to the scenario of Cooper *et al*. (1995), then the geometry of the barriers with respect to incoming waves (*e.g.* drift alignment vs. swash-alignment) is controlled by changes in sediment supply. If the barriers fit the model of Delaney *et al*. (2012) and Devoy (2013, 2015), then storms (in combination with sediment supply) may play an important role in the evolution of the system. An aim of this study is to examine this role.

So which scenario is more likely? The N-S trending orientation of the barriers is consistent with the scenario of Delaney *et al*. (2012) and Devoy (2013, 2015), but the progressive reduction in the age of sediments from the proximal to distal end of Inch may suggest otherwise (Wintle *et al*., 1998). If the spit-building scenario is correct, it is possible that the Caragh River sediments acted as a barrier to onshore sand sediment flux, giving rise to spit development at Rossbehy. However, the development of two spits almost directly across the bay from one another could hardly be coincidental, a point that Devoy (*pers. comm*) argues lends credence to the glacial moraine scenario. In addition, the younging of sediments can be explained in the context of this scenario. Delaney *et al*. (2012) and Devoy (2013, 2015) posit that due to wave refraction at the proximal end of Inch, younger sediments may have been stripped away. More detailed work, perhaps involving ground penetrating radar (GPR), is required to confirm whether or not an end moraine exists beneath the barriers.

3.1.2 Recent behaviour

Table 3.3 gives a more recent (500 years BP to present) chronology of the barriers. Both Inch and Rossbehy are depicted on maps dating back to at least 1673 (*e.g.* figure 3.6). Evidence from IRSL dating by Cooper *et al*. (1995), which is supported by historical accounts from Smith (1756), suggest that the contemporary morphology of the system was attained by the mid-18th century. Documentary records show that a major breaching event occurred around this

time at Inch (Delaney *et al.*, 2012). This coincides with a major dune building event recorded in sediments (Cooper *et al.*, 1995).

Agricultural activity in Castlemaine harbour was widespread from the 1750s onward, and land reclamation at Rossbehy for agricultural purposes has been documented in numerous sources (*e.g.* Allanson-Winn, 1899; Herity, 1970). The mining of beach sands in the area was common from the 19th century onward, peaking around 1845. The cutting of marram for thatching was also common around this time, right up until the early 1900s (Delaney *et al*., 2012).

Cooper *et al.* (1995) analysed contemporary historical shoreline change for Inch (figure 3.7) and Rossbehy (figure 3.8) from 1841 to 1993 using historical maps, aerial photographs, and field survey data. At Inch, they found the maximum variation in shoreline position occurs at its distal end, with maximum rates of shoreline recession along the length of the barrier associated with the period 1949 to 1967 and maximum rates of accretion associated with the period 1973-1994. Orford *et al.* (1999a) interpreted these changes at Inch as being part of 30-50 year cycles driven by extreme storms and associated surge (see section 3.2.4). At Rossbehy, Cooper *et al.* (1995) noted a shoreline recession of 200-300 m occurred between 1841 and 1954 at the proximal end of the barrier. Subsequently, between 1954 and 1993, a shoreline advance of 20-30 m along the length of the barrier appears to have occurred. This corresponds with Guilcher *et al*.'s (1960, p. 322) account, which stated that, at its proximal neck, Rossbehy was "broken by two breaches through which the sea can pass during severe storms. The eastern part of the breaches is bare, and covered by the sea at high spring tide; the west part is beginning to be filled up by heaps of sand bearing some marram." O'Shea *et al.* (2011) undertook a similar shoreline change analysis extending from 1842 to 2010 covering both Inch and Rossbehy (figure 3.9). In agreement with Cooper *et al.* (1995), O'Shea *et al.* (2011) concluded that the morphology of Inch could be charactersised as dynamic on a decadal scale, but overall remained stable in position and shape. Rossbehy, however, was found to undergo long-term stability followed by rapid shoreline change, as was exhibited in 1842 and the early- to mid-2000s, by which point its distal neck had narrowed significantly since 1977. Figures 3.10 and 3.11 show aerial photographs and satellite imagery illustrating the evolution of Inch and Rossbehy, respectively, up to the present. These images exemplify the overall

conclusion of O'Shea *et al.* (2011) and show that while Inch continues to remain generally stable, Rossbehy has exhibited major changes at its distal end. In summary, the two barriers appear to behave independent of one another, which is striking considering they are subject to the same blanket boundary conditions (*e.g.* wind/wave climate).

During the weekend of 13/14 December 2008, following nearly a decade of narrowing at its distal neck, Rossbehy breached. Meteorologically speaking, the event responsible for breaching was not particularly significant (Met Éireann, 2008). To put it into context, two of the most extreme events to affect the local area were Hurricane Debbie (1961) and the Night of the Big Wind (1839). The minimum central pressure and wind speeds for those events were 964 hPa and 50 m/s (Hurricane Debbie) and 918 hPa and 51 m/s (Night of the Big Wind) (Orford *et al*., 1999a). The minimum central pressure and wind speed for the event responsible for breaching were 988 hPa and 10.8 m/s. More than 10 ha of dunes were levelled, resulting in the separation of the distal end of the barrier from the mainland and the formation of a new tidal inlet (figure 3.12). Semi-diurnal tidal exchange has been on-going since breaching, having been most recently observed in June 2016. As a result, the new channel can be considered a newly formed inlet.

Post-breaching, cannibalisation of Rossbehy has been on-going. The barrier can now be compartmentalised into a stable, swash aligned zone and an eroding drift aligned zone (Sala, 2010; O'Shea and Murphy, 2013). This erosion resulted in the destruction of an 18th century navigation tower at the distal end of the barrier in the winter of 2010. The breach widened from 600 m in 2011 to 900 m in 2015, and average rates of dune recession adjacent to the breach are presently on the order of 50 m/yr.

Finally, Rossbehy received a great deal of media attention during the 2013/2014 winter storm season, when the Irish coastline was impacted by a series of large storms – some of the largest on record. Simultaneous strong winds, tidal surges and low pressure resulted in substantial damage to Rossbehy, including damage to the main road providing access to the beach (figure 3.13), the destruction of the children's play area at the entrance to the strand, and the displacement of the Sunbeam shipwreck, which was buried on the beach since it ran aground in 1903 (figure 3.14). Cleanup operations were repeatedly interrupted by subsequent

storms. There is presently growing public concern in Kerry, as well as elsewhere in Ireland, over issues related to the management of erosion and flooding and damage to homes, infrastructure and heritage sites along the coast.

3.2 Local morphodynamics

Together, Inch and Rossbehy can be classified as a tidal-inlet midbay barrier system (Sala, 2010). The main inlet separating the barriers acts as a central hinge and appears to have been fixed in place since at least 1500 BP (Cooper *et al*., 1995). Sediment sources to the system include fluvial sources (from the Maine, the Laune, and the Caragh Rivers) and erosion of hard and soft cliffs. Sediment sinks include the Inch, Rossbehy, and Cromane barriers themselves (including the dune barriers at Inch and Rossbehy), the ebb-tidal delta complex, and the tidal flats and marshes. Caragh Lake, located 1.5 km upstream of the mouth of the Caragh River, may also act as a sink for sediment runoff from the upper part of Caragh catchment, limiting the amount of sediment coming downstream and entering the estuary behind Rossbehy. According to Cooper *et al*. (1995), the salt marsh in the inner basin acts as a medium-term (10^0-10^2 years) store of sediment, and the intertidal mud and sand flats of the outer basin represent short term (1-10 years) sediment stores. This is exemplified in figure 3.15, which shows changes in the shape and position of the ebb shoals fronting Inch and Rossbehy from 1894-2015. Since barrier breaching and subsequent widening occurred at Rossbehy, the ebb shoal fronting the breach has been observed to have increased in height and aerial extent, although the precise magnitude of which is unknown. A detailed assessment of changes in the aerial extent of the shoals relative to shoreline change has not yet been carried out. Such an assessment could help to explain concomitent changes in shoreline position and dune erosion. Further discussion on the dunes, local hydrodynamics, storms and barrier breaching is provided in the following sections.

3.2.1 Human impacts

Contemporary human influences on barrier morphodynamics at Inch and Rossbehy include shoreline protection works, recreation and tourism activities, and animal grazing. Coastal protection works are present at the entrances to both Inch and Rossbehy strands, where it is a priority to protect the main roads

providing access to the beaches. Groynes have been present on the beach fronting the proximal neck of Rossbehy since c. 1900. These were installed in an effort to protect the entrance to the strand such that it would not become separated from the mainland, which was a concern following breaching here around this time (Allanson-Winn, 1899). The groynes are presently reinforced by cobble- to boulder- size rock armour, which is maintained by Kerry County Council to protect the main road providing access to the beach. At Inch, a similar type of protection is present at the entrance to the strand, where, in response to safety concerns in 2005, Kerry County Council constructed sea defences to protect the narrow coastal arterial road which serves as one of the main tourism routes in Ireland (Gault *et al.*, 2011). This reactive approach to erosion management is typical in Ireland, where "the current absence of any national coastal management policy, the associated lack of an agreed monitoring approach coupled with antiquated foreshore legislation, makes it extremely difficult for local coastal managers (primary stakeholders) to provide a coherent response to coastal erosion" (Gault *et al.*, 2011, p. 931).

Pressure from tourism and recreational activities at Inch and Rossbehy is somewhat heavy, especially in the summer months (Devoy, 2015). The degradation of dune vegetation as a result of the use of motorized vehicles, and, to a lesser extent, walking and horseback riding has resulted in the formation of large dune blowouts (figure 3.16). Given little time for vegetation to take hold and stabilize the dunes, they remain vulnerable to wind action and further erosion.

Grazing on the dunes is common, although moreso at Inch, where the local landowner grazes sheep and cattle. While grazing is prohibited at Rossbehy, which is publicly owned land, it has been observed on numerous occasions. The combined impacts of grazing, the use of the dunes by people for recreational activities, and quad and motorbike riding have not been thoroughly evaluated at Inch or Rossbehy, although Devoy (2015) has drawn attention to these activities which are responsible for the degradation of many of the dune forms.

Proposals to develop a golf links course at Inch have been under debate for over ten years. Devoy (2013) evaluated the potential geomorphological and physical environmental impacts of such development. In this assessment, it was warned that the impacts of climate warming could lead to potentially negative

consequences at Inch, with knock-on effects from breaching at Rossbehy alone predicted to affect shoreline erosion and sediment redeposition at timescales within the order of the next 10-20 years (Devoy, 2013). Any potential negative impacts on the morphological development of Inch as a result of the golf course development (*e.g.* the interruption of sediment cell dynamics as a result of the stabilistation of barrier sands) would be compounded by this. As of November 2014, it was reported that a top American golf course owner was behind the latest golf course development plans at Inch (Radio Kerry, 2014), although a planning application has not been made as of 24 February 2016.

3.2.2 Dunes

Digital elevation models (DEMs) illustrating the Inch and Rossbehy dunescapes are shown in figures 3.17 and 3.18. On the seaward side of Inch, a low-angle dissipative beach gives way to a series of low to high, discontinuous foredune ridges, the most seaward of which has been extensively scarped by wave action, especially on the south-western area adjacent to the inlet (figure 3.19). The dunescape at Inch can be subdivided into two distinct zones, whereby zone A, the southern and mid-western zone (figure 3.17), is a zone of active dune activity and zone B, the northern and mid-eastern zone, is characterised by vegetated and stabilised dune and sand surfaces (Devoy, 2013). The duneforms in the north are circular or semi-circular blowouts backed by somewhat stabilised residual dunes. Those in the south are generally higher (often exceeding 15 metres; *e.g.* figure 3.20) and characterised by a series of elongated ridges orientated in the direction of the dominant wind. These are separated by deep, narrow troughs, which sometimes merge into parabolic dunes. Ephemeral embryo dunes form seaward of the foredunes along the length of the barrier, although most extensively at the southernmost tip of the barrier (figure 3.21). The dunes are comprised of well rounded, fine to medium sized quartz sands (c. 0.06-0.6 mm), with 3-10% of the material consisting of broken shell ($CaCO_3$) fragments (Devoy, 2013). The dominant dune grass is marram (*Ammophilia arenaria*). Inch supports five Annex I habitats protected under the EU Habitats Directive: annual vegetation of drift lines, embryonic dunes, shifting dunes along the shoreline with *Ammophilia arenaria*, fixed dune, and dunes with *Salix repens ssp. Argentea* (*Salicion arenaria*).

At Rossbehy, the dunescape is smaller in areal extent (approx. 0.8 km^2 compared to >4 km^2 at Inch). Rossbehy can similarly be divided into zones with characteristic duneforms. The largest zone, the southern and middle interior, is characterised by a series of transverse ridges, which, like at Inch, are oriented in the direction of the prevailing wind (figure 3.22). These are separated by intervening dune slacks (figure 3.23), which sometimes merge into parabolics. The innermost ridges are stabilised by marram, although in places vegetation cover is discontinuous. The dunes on the south-western section of the barrier are fronted by an extensive shingle bank, which runs along the length of the barrier for approx. 0.8 km. The foredunes here are mostly blowouts and are vulnerable to wave attack during storms. In the northern drift aligned section of the barrier (adjacent to the southern side of the breach), the foredunes are subject to extensive and on-going scarping and subsequent disintegration. Maximum rates of foredune recession here are on the order of 50-125 m/yr. Beyond the breach, where a barrier island is remnant of Rossbehy's pre-breach morphology, residual dune forms have remained stable, with discontinuous vegetation cover in the area adjacent to the northern margin of the breach.

Foredune development at Inch and Rossbehy is facilitated by a strong southwesterly wind regime (figure 3.24). Annual average wind speed at Valentia Observatory, located approximately 20-30 km southwest of Inch/Rossbehy, is 5.5 m/s. The number of days per year with gale force gusts regularly exceeds 100 (Met Eireann, 2011). It is interesting to note that due to its strong onshore wint climate, Inch has served as an ideal location for studying the predictive capacity of various Aeolian transport models (Sherman *et al.*, 1998; Sherman and Li, 2012).

3.2.3 Hydrodynamics

The tidal range along the coast of Dingle Bay is between 3.5 and 4.5 m and increases as the bay narrows. Inch is mesotidal, with a spring tidal range of 3.2 m (table 3.4) (Vial 2008). Due to their close proximity to one another, it can be assumed that tidal levels at Inch and Rossbehy are similar (Sala, 2010). The spring tidal range increases to 4.5 m at Cromane, and further up the estuary, to >5 m at Killorglin. This reflects an increase in the tidal prism moving up-estuary and is important because sediment supply to flanking systems is controlled by

the tidal prism (*e.g.* an increase in the tidal prism results in an increase in the delivery of sediment to the ebb-delta, which may be derived from flanking systems). The increase in cross-sectional area of the channel between Inch and Rossbehy and Cromane and Inch is a manifestation of the increasing tidal prism. The larger tidal prism between Cromane and Inch produces stronger (more hydraulically efficient) tidal currents and increased channel scour, resulting in a larger inlet cross-sectional area.

The main inlet channel generally follows an S-shape, abruptly curving southwards from the distal end of Cromane before again changing direction near the distal end of Inch. The inlet has been characterised as being between mixed wave/tide-dominanted and tide dominated (Sala, 2010). Tidal currents in the vicinity of the barriers are vigorous. Surface currents of 0.9 m/s were recorded by O'Shea and Murphy (2013) in the marginal flood channel fronting Rossbehy. The dominance of tidal currents over wave energy in this location is evidenced by the shore-normal orientation and elongate morphology of the ebb-tidal delta (Cooper *et al.*, 1995). The system is ebb-dominant (*e.g.* ebb currents are stronger than flood currents and rising tide is longer than falling tide).

In terms of wave climate, the coastal area in Dingle Bay is storm wave dominated with consistent incursions of swell waves approaching from the west and southwest. The mean offshore significant wave height is 2.8 m with a period of 7 s (Vial, 2008; Sala, 2010). Based on an analysis of 40 years (1958-1997) of wave hindcast data, Vial (2008) reported that a calm offshore wave climate (H_s<3m) dominates 70% of the time, with a storm wave climate (H_s>6 m) accounting for 4% of the time.

3.2.4 Storms

Storms can cause significant morphological impact at spatial and temporal mesoscales (10^1-10^2 km; 10^0-10^2 years) in Ireland (Cooper *et al.*, 2004). As such, a large body of work has been devoted to understanding these impacts on Irish coastal systems, including at Inch and Rossbehy (Cooper *et al.*, 1995; Duffy and Devoy, 1998; Orford *et al.*, 1999a; Orford *et al.*, 1999b; Lozano *et al.*, 2004; Cooper *et al.*, 2004; Vial, 2008; Vijaykumar *et al.* 2004; Williams *et al.*, 2015). This work is relevant to this study in that a primary aim of this research is to

understand the role of storms the evolution of the Inch-Rossbehy coastal system post-breaching.

The west coast of Ireland lies in the path of several common North Atlantic storm tracks, which approach the coast from the southwest. The majority of the storms that reach the coast are extratropical cyclones that originate as slow-moving depressions in the midlatitude westerly wind belt (Lozano *et al*., 2004). Lozano *et al.* (2004) found that the mean duration of these events (defined by a minimum wind speed of 15.3 m/s) is 4.5 days.

The surge component of these storms is dependent on their minimum central pressure. A general rule of thumb is surge height increases by 1 cm for every mb below 1015 mb (Murphy, *pers. comm*). According to Orford *et al.* (1999b, p. 1853), "the median event pressure [in the vicinity of the southwest coast] is 981 hPa. Storms with equal or less central pressure can be expected to appear c. 200 times a year…" This equates to a typical surge of approximately 34 cm.

In terms of storm wave climate, an analysis by Vial (2008), also referred to in the previous section, suggests that a storm wave climate (H_s>6 m) occurs just 4% of the time.

Extreme events reaching the coast can be defined in a variety of ways. One way is to calculate the water level maxima associated with an annual water level exceedence probability. A 2% annual water level exceedence probability has a 2% chance of occurring in a year, or once in every 50 years. Based on nearshore wave gauge data, Vial (2008) estimated the water level maxima for a 2% exceedence event at Inch (5.5 m nearshore H_{sig}) to be 1.75 m. Earlier work by Orford *et al*. (1999a; 1999b) defined storms at Inch based on their propensity for dune erosion.

Orford *et al.* (1999a; 1999b) investigated the relationship between extreme storms and dune erosion at Inch. They found that at the meso-scale (decade-century), dune erosion effectiveness was controlled by storm characteristics, particularly the surge component, severity (dimensions of central pressure and wind speed), storm track and associated wave direction, and coincidence with high tide. Notable findings from these studies are summarised as follows:

- Based on analyses of historic maps, aerial photographs, and field surveys, two periods of maximum shoreline retreat were identified over the period 1842-1993. Human activity, sea-level variability, and wind variability

were also analysed for the corresponding period and, based on lack of evidence, deemed unlikely to be the cause of the observed mesoscale temporal changes to foredunes.

- The incidences of extreme events were linked to the two observed phases of maximum shoreline retreat, which occurred sometime before 1842 (linked to the Night of the Big Wind) and between 1958 and 1967 (linked with Hurricane Debbie). The intervening periods were characterised by progressive foredune accretion. Based on these observations, Orford *et al*. (1999b) proposed a conceptual model (figure 3.25) whereby the meso-scale evolution of Inch is said to be controlled by episodic high-magnitude, low-frequency events triggering rapid erosion followed by gradual recovery. These events occur at approximately 30-50 year intervals. During the intervening calm periods, excess sediment in the nearshore is reworked and delivered to the beachface under fairweather wave conditions, facilitating dune regeneration. It should be cautioned that Orford *et al.*'s (1999a; 1999b) model was based on circumstantial evidence (due to the paucity of process data for the Irish coast).
- The importance of storm dimensions and timing was examined by looking at the characteristics of extreme storms dating back to 1839 and subsequent dune responses. The position of the storm centres relative to Dingle Bay was found by Orford *et al*. (1999a) to be important, as this determines the surge component. In addition, coincidence with high water was found to be a key element in determining storm effectiveness in terms of dune erosion. The coincidence of of these overlapping requirements such that they were sufficient to exceed the erosion threshold occurred only 3 times in the last 150 years.
- Orford *et al*. (1999b) modelled extreme wave conditions without a surge to determine the importance of the surge element. They looked at wave induced stress under modal and extreme conditions and found that a major surge increment (2-3 m) is required to result in significant dune erosion, otherwise refraction by the ebb-tidal delta means that energy is mostly dissipated by the delta front.

Cooper *et al*. (2004) later investigated the morphological response to storm induced wave forcing at Inch/Rossbehy and other sites along the west coast by simulating waves under modal and storm conditions. They found that during storms (H=6.6 m, T=13.6 s), the distal margins of Inch and Rossbehy remained sheltered from the impact of larger swell waves. Under modal swell (H=0.4 m and T=7 s), the maximum wave energy dissipation was concentrated midway along Inch and seaward of the area where the dune breach occurred at Rossbehy (figure 3.26). Looking at wave orbital velocities for modal swell conditions and high energy wind conditions similar to those that occurred during Hurricane Debbie, Cooper *et al*. (2004) observed an increase in velocities seaward of the area where the dune breach occurred at Rossbehy (figure 3.27). Based on these results, they concluded that an increase in swell size may not necessarily lead to a different wave energy dissipation pattern (*e.g.* storm waves are not necessarily always the most important control on dune erosion at Inch, and presumably Rossbehy, during storms).

Vial (2008) later simulated significant wave height patterns under storm conditions in Dingle Bay for varying water levels. His results (figure 3.28) suggest that the impact of storms is strongly dependent on water levels, with higher waves allowed to reach the area seaward of where breaching occurred (and other parts of the barriers) as water levels increase. This has significant implications, considering projections for future sea-level change.

Other storm attributes considered to be important at Inch/Rossbehy include peak period (T_p), storm duration and lag time. Williams *et al*. (2015) performed a series of model experiments for Rossbehy using XBeach, a 2D model for wave propagation, long waves and mean flow, sediment transport and morphological changes during storms. They found that, firstly, increasing T_p from 10 s to 16 s resulted in a 45% increase in dune recession. Second, they found that threshold H_{scrit} values for Sallenger's (2000) storm impact levels 2 (impact), 3 (overwash), and 4 (indundation) were *lower* for a 25 hour event than for a 12.5 hour event, demonstrating the importance of storm duration. They noted "storms of sustained duration or storms occurring in rapid succession with little time for shoreline recovery are important events driving morphological change" (Williams *et al.*, 2015, p. 12), although they did not include an evaluation of the importance of lag time in their study.

Finally, as result of climate warming, storms are projected to become more frequent and more intense (Kiely *et al.*, 2005; Beniston *et al.*, 2007; Mori *et al.*, 2010; Zappa *et al*., 2013), which would have negative implications for Inch and Rossbehy. The results of a multimodel assessment of future projections of North Atlantic and European extratropical cyclones indicate an increase in the number and intensity of cyclones associated with strong wind speeds over the British Isles (Zappa *et al*., 2013). Cooper *et al*. (1995) have speculated that a reduction in the return period of storms in the vicinity of Inch would be associated with a positive feedback mechanism on beach storage volume. As there is less time available between storms for beach storage replenishment, the volume of material available for dune building decreases. This means barrier dunes are more vulnerable to wind and wave action and thus more mobile at times of low beach storage. The need for further research into the impacts of storms on these systems under climate warming is, therefore, all the more important and urgent.

3.2.5 Barrier Breaching

Evidence for possible breaching events prior to the 2008 event at Rossbehy have been presented by Delaney *et al*. (2012) and O'Shea (2015). Delaney *et al*. (2012) interpreted an abrupt change from intertidal marsh sediments to channelised sands and silts in sediment cores as evidence of a pre-historic breaching event. Organic material from the underlying marsh sediments was dated at 2873 years BP. A transition from sand and silt into finer sediments was taken as evidence of barrier reformation (after c. 2,800 BP). If this were the breaching event responsible for the formation of the main inlet separating the two barriers (*e.g.* the second inlet formation scenario described in section 3.1.1), the transition to fine sediments could be interpreted as the time at which the inlet reached a stable configuration. This places the formation of the inlet much earlier than 1500 BP, as suggested in the scenario of Cooper *et al*. (1995) and calls into question the relative influence of sea-level on barrier evolution.

In the same study, a similar sequence (fine sediments overlain by sand) was interpreted by Delaney *et al.* (2012) as a possible second breaching event, this time dated at 800 BP. It should be cautioned, though, that it remains unclear whether or not these sequences represent barrier breaching at all, as they may simply represent overwashing of the barrier system (Delaney *et al*., 2012).

Other evidence of barrier breaching at Rossbehy has been presented by O'Shea (2015) in the form of historical barrier recurves identified from an aerial photograph (figure 3.29). These were interpreted as being either earlier northern limits of dune progression or southern limits to a historical breaching event. The age of the recurves is unknown.

Anecdotal evidence from Allanson-Winn (1899) and Guilcher *et al*. (1960) suggests the proximal neck of Rossbehy was breached around 1900 and again in the late 1950s. Similar accounts of breaching at Inch have not been reported, but barrier recurves similar to those at Rossbehy may represent historical breaching events.

Sala (2010), O'Shea *et al.,* (2011) and O'Shea and Murphy (2013) have described the circumstances leading up to the December 2008 breaching event at Rossbehy. From analyses of historical maps and aerial photographs, it was established that Rossbehy's distal neck began to narrow after 2000. Between 2000 and 2006, rates of dune recession north of the recurve point (where Rossbehy eventually breached) were on the order of 12 m/yr (an estimated removal of 52,000 m^3 of sediment per year). Also during this time, a straightening of the main inlet channel was observed from analysis of aerial photographs by O'Shea *et al.* (2011). This meant the hydraulic efficiency of the main inlet increased and sediment could be transported into deeper water, starving the ebb tidal bar and beach of sediment. As a result, the dunes landward of the ebb delta (in the zone north of the barrier's recurve point) became increasingly vulnerable to wave attack which eventually culminated in breaching. Between 2006-2009, the erosion rate in this area was estimated to be 530,000 m^3 / yr. Sala (2010) and O'Shea *et al.* (2011) have therefore argued that breaching was more likely due to a decline in beach volume rather than the impact of one or more storms, although there is evidence that the period between 2004-2009 had a higher than average concentration of winter storms (Sala, 2010).

Since breaching occurred, semi-diurnal tidal exchange has been ongoing and was observed as recently as June 2016. From an initial width of 500 m, the breach has widened to approximately 900 m (as of 2015) as the drift-aligned zone of the barrier continues to expand at the expense of the more stable swash-aligned section of the barrier. Preliminary work by Sala (2010) indicated some infilling

occurred immediately after breaching. Sala (2010) performed GPS surveys of the breach in 2009 and 2010. The depth of the breach fluctuated from -1.6 m to -1 m in the summer of 2009 to c. 0 m in the winter of 2010. Some infilling occurred between the winter of 2010 and the summer of 2010, but depths remained between -1 m to 0 m. By the winter of 2012, though, cross-breach GPS surveys conducted by O'Shea and Murphy (2013) showed that the breach had deepened to <-2 m. For reference, MHWS and MLWS are respectively equal to 1.46 m and -1.72 m (Sala, 2010).

Immediately following breaching, local concern as to how breaching might affect the estuary as a whole led to research ultimately published by O'Shea (2015). This concern centred on whether or not an increased tidal prism (as a result of breaching) would result in increased erosion and flooding of the back barrier estuary. Recall an increased tidal prism is associated with an increase in storage in the ebb tidal delta. This sediment needs to come from somewhere – *e.g.* sedimentary environments flanking the inlet channels. To investigate the potential for increased flooding in the area, O'Shea (2015) simulated waves and currents before (2000), just prior to (2006), and after (2009) breaching in a numerical model (MIKE21). He found that current speeds through the main channel increased from the pre- to post-breach bathymetries. Model results suggested that flood risk in low lying areas in the estuary is increasing with the continued erosion of Rossbehy, with the most vulnerable area identified as the lower Cromane area. Further discussion of the results of that research is provided in section 3.2.6.

3.2.6 Morphodynamic evolution

Orford *et al*. (1996) proposed a conceptual model describing control domains and morphological phases of gravel barrier evolution at multiple temporal scales. Despite being sand-dominated barriers, Inch and Rossbehy appear to fit within this model, as there are many parallels between their behaviour and the behaviour of the gravel dominated barriers in Nova Scotia, on which the model is based. Within this model, there are four domains of barrier development (not necessarily successive). These can be summarised as follows:

- **Growth** – This phase is characterised by an increase in sediment supply; the barrier is drift-aligned (orientated parallel or oblique to the incoming wave direction).

- **Consolidation** – The barrier switches from being drift-aligned to swash-aligned (barrier is orientated oblique or perpendicular to incoming waves); macro-scale cannibalisation is occurring. If the sequence of development described by Cooper *et al*. (1995) for Inch is correct, Inch would have displayed such behaviour between 1500 BP and 500 BP.

- **Breakdown** – Three phases occur: slow migration; fast migration; and dissolution; micro-scale cannibalisation occurs. Orford *et al*. (1996) believe that while the influence of long-term sea-level rise is likely muted at this stage, short-term (<10a) SLR may be an important control on the rate of barrier breakdown.

- **Reformation** – The barrier switches from being swash-aligned to drift-aligned.

These distinctive process domains are governed by the varying combinations and intensity of basic controls on barrier structure, including sediment supply, sea-level rise, wave climate, etc. The most important control on barrier alignment, they argue, is sediment supply. Sediment supply to Inch and Rossbehy from longshore transport is limited by the elongated orientation of the Dingle Bay shoreline. According to the model of Orford *et al*. (1996), this explains how the barriers came to be swash-aligned (*e.g.* there is to zero longshore transport potential at the bay-scale). The presence of minor drift aligned recurves at the barrier termini, however, indicates longshore transport does occur at the barrier-scale.

Sala (2010) invoked Orford *et al*. (1996)'s model to explain past and present changes in the shape of Rossbehy. She argued that a narrowing of Rossbehy's distal neck between 1842-1894 and also between 2003-2006 corresponds with Orford *et al.* (1996)'s 'slow migration' phase. This narrowing can be explained by a decrease in barrier-scale longshore transport, marking the onset of micro-scale cannibalisation. During cannibalisation, multiple sediment cells are created

along the length of the barrier. The cells are generally characterised by a down-drift zone of erosion and an up-drift zone of accretion. Sala (2010) delineated four cells at Rossbehy (figure 3.30). Cell 1 corresponds to the section of the barrier that remains swash-aligned and is not affected by cannibalisation. Cells 2, 3, and 4 represent the subdivision of what was, prior to the onset of cannibalisation, a second single cell. This single cell was subdivided as a result of a reduction in sediment supply. According to Sala (2010), breaching occurred as a result of this reduction and represents the final breakdown (dissolution) phase of Orford *et al*. (1996). Sala (2010) speculated that the reduction in littoral drift was related to tidal induced currents and wave processes

Building on this work, O'Shea (2015) investigated the potential medium-term (10^1) impacts of breaching on inner Dingle Bay using numerical modelling. He proposed a five-stage conceptual model of barrier evolution. This model can be summarised as follows (also see figure 3.30):

Stage 1

Stage 1 of the cycle represents the period prior to breaching, where the removal of the swash platform between 2004-2008 left the drift aligned section of Rossbehy vulnerable to wave attack. The straightening of the channel meant that sediment could be transported further offshore, starving the ebb bar of sediment.

Stage 2

Rossbehy is presently in stage 2 of its development. During this stage, a positive feedback is in operation whereby the growth of the ebb tidal bar is facilitating dune erosion in the drift aligned zone by causing waves to approach perpendicular to the foredunes. In addition, the establishment of a channel between the ebb bar and the barrier facilitates the removal of sediment from the system on the ebb flow.

Stage 3

During this stage, although the breach continues to widen, the ebb bar begins to migrate toward the drift aligned zone. This is as a result of channel infilling, with the sediment source being the material eroded from the dunes.

Stage 4

The bar welds onto the barrier (via channel infilling) and there is a slowdown in dune retreat. Embryo dunes develop on the breach.

Stage 5

Dune repair occurs. Due to the strong wind climate at Rossbehy, O'Shea (2015) argues that there is a high potential for dune regeneration at Rossbehy, provided storms don't destroy the embryo dunes.

The results of numerical modelling simulations suggest that the initiation of breach recovery is expected to occur by the year 2033, by which time the breach will have widened to 1400 m. O'Shea (2015) has argued that as the ebb tidal bar infills, it is expected to reduce the rate of widening. He estimates that it could be reasonably assumed that infilling would take place approximately 20 years from a base year of 2013.

In contrast to O'Shea (2015)'s conclusion with regard to the future of Rossbehy, Devoy (2015) has warned that under future climate warming, Rossbehy is more likely to disintegrate. The author of this thesis shares this view. It is argued in chapter 11 that a major weakness of O'Shea (2015)'s proposed model is its failure to take into account the impacts of storms (or a potential increase in their frequency and magnitude) or sea-level change. This research seeks to compensate for this.

The impacts of breaching at Rossbehy on Inch remain unclear. Devoy (2013, p. 5) speculated: "The break-up of Rossbehy, post 2008/9 with possible 'complete' disintegration if left to itself and without further human intervention, is likely now to alter the recorded quasi-periodicity in storm driven control of sediment movements in the active areas of Inch" (Devoy, 2013, p. 5). Further evidence from monitoring is required to assess the influence of breaching on Inch.

4 Sea-level change: past, present and future

According to the United Nations Atlas of the Oceans, approximately 44 % of the world's population lives within 150 kilometres of the sea (UN, 2010). Modern day sea-level change (SLC) places the livelihoods of millions of people at risk (Pilkey and Young, 2009), and given the scale of the issue, it is important to establish more specifically what this means for coastal populations. This is the impetus driving present-day sea-level change research. There is a vast amount of research that is relevant to SLC. An exhaustive review of this would be a formidable task and has largely already been performed by the IPCC (see Church *et al*., 2013). As such, this section concentrates on the following key areas of interest to this study: (1) past, present, and future changes in global and local mean sea-levels; (2) the impacts of SLC on soft coasts; and (3) the implications of future SLC for Ireland. The main areas of debate within this literature centre around how we project future SLC, how much sea-level will rise in the 21st century and beyond, and how soft (sediment-dominated) coasts will respond to SLC (including the potential impacts of more frequent and extreme storm events). An evaluation of the corresponding literature follows.

4.1 Paleoenvironmental work and sea-level change

The development of modern sequence stratigraphy in the mid-20th century was an important precursor to our current understanding of long-term global SLC. With the establishment of the global sea-level curve (figure 4.1) by researchers such as Peter Vail and his colleagues from Exxon in the late 1970s (Vail *et al.*, 1977) and subsequently others (Hallam, 1981; Haq *et al.*, 1987; Rohling *et al.*, 1998; Miller *et al.*, 2005) came an explosion of interest in paleoclimate research, which was gaining in importance as concerns about future anthropogenic climate warming grew in the 1980s and 1990s. This type of research is important because paleoclimates can serve as potential analogues of future climate and they represent invaluable opportunities to test and improve climate models. For example, during the mid-Pliocene (3.3-3.0 Ma), sea-levels were around 20 m higher than today (Miller *et al.*, 2012; Church *et al.*, 2013b). The significance of this period, and others like it (*e.g.* the mid-Miocene Climate Optimum), is that atmospheric CO_2 concentrations were likely similar to those of today (402 ppm as of February 2016 according to the US National Aeronautics and Space

Administration – see http://climate.nasa.gov/ for the most up-to-date figure). There are many unknowns, however, surrounding other environmental conditions at the time, so to extrapolate and say that sea-levels could rise by 20 m as a result of present-day warming would be inaccurate. Nonetheless, with cautious interpretation by experts in the field, past climates can provide some insight into how the Earth-atmosphere-ocean-ice-vegetation-climate system responds to changes in atmospheric CO_2 concentrations and thus help us to better understand the processes that result in changes in sea-level (Lee, 2015).

Paleoenvironmental research has substantially improved our understanding of the complex relationship between climate and sea-level. From the sea-level curves mentioned earlier, we know that global sea-levels have fluctuated by hundreds of metres in the geologic past. The curves themselves lend credence to Milankovitch theory, the widely accepted theory that Earth's climate is a function of variations in Earth's orbital position relative to the sun. This is sometimes referred to as orbital, or Croll-Milankovitch, (climate) forcing. During the Quaternary (2.6 Ma to present), sea-levels varied primarily as a result of this. A select group of researchers (Quaternary scientists) specialise in the study of environments shaped during this period. From this type of work, which is often interdisciplinary, it is known that sea-levels ranged from as much as 120 - 130 m lower than at present (Fairbanks, 1989; Bard *et al.*, 1990; Bard *et al.*, 1996; Rohling *et al.*, 1998) to 2-6 m above present (Chappell *et al.*, 1996; Church *et al.*, 2007; Murray-Wallace and Woodroffe, 2014). These figures are derived from sequence stratigraphy and, often, using carbon and other dating techniques, the general timing of sediment deposition can be inferred. Uncertainties with regard to paleo- sea-levels arise from differences in interpretations of stratigraphic sequences, uncertainties associated with dating techniques, and uncertainties associated with climate proxies.

Some Quaternary scientists specialise in Holocene (c. 10,000 years to present) SLC, which is easier to reconstruct given that there is less time between when sediments were laid down and the present, and therefore less chance of disturbance. From this type of work, it is known that rates of eustatic (or global) sea-level during the early- to mid- Holocene rose at up to 40 mm/yr before falling to <2 mm/yr from 3,000 BP to present day (Milne *et al.*, 2005; Bindoff *et*

al., 2007). In the literature, however, relative SLC (changes in the height of sea-level with respect to the land surface) is of more interest during this period. There has been much work done on constraining rates of glacio-isostatic adjustments (GIA), especially at high latitudes where ice loading has had (and, in many places, continues to have) a significant effect on the height of the land surface. In Ireland, major contributions to the study of relative SLC have been made by R.W.G. Carter (University of Ulster), Robert Devoy (University College Cork), and Julian Orford (Queens University of Belfast) (*e.g.* Devoy, 1983; Orford *et al.*, 1995; Carter *et al.*, 1989; Devoy *et al.*, 2006;

4.2 Recent sea-level changes

Over the past two centuries, eustatic (global) mean-sea level (GMSL) has been rising (Church *et al.*, 2013b). This is known from careful averaging of coastal tide gauge records from around the globe (Church and White, 2006). Figure 4.2 shows global mean SLC from 1870-2000 (CSIRO, 2014). During this period, GMSL has risen by approximately 2 cm at a rate of approximately 1.7 mm/yr (Church *et al.*, 2013b). More recently, a significant acceleration of SLR has been observed (Church and White, 2006). For the period between 1993-2014, the rate of GMSLR increased to 3.2 mm/yr (figure 4.3). Model results suggest that 87% of the observed SLR since 1970 was induced by human activity (Marcos and Amores, 2014).

There are significant regional variations in rates of mean SLC, with some areas experiencing a higher rate of rise than the global average (*e.g.* as a result of land subsidence) and some areas experiencing a sea-level fall (*e.g.* as a result of GIA). From tide gauge records we know that in Ireland, relative sea-level is rising at approximately 1 mm/yr at present, although, due to GIA, there are significant variations across the island (Devoy, 2008). For example, in Dublin, sea-level rose at a rate of 0.23 mm/yr for the period 1938-1996. At Malin Head (Donegal), sea-level rose at a rate of only 0.06 mm/yr for the period 1959-1997. In Belfast, sea-level fell at a rate of 0.99 mm/yr for the period 1957-1969 (Sweeney *et al.*, 2008). While the effects of climate warming on relative SLR in Ireland are not yet apparent (Devoy, 2008), a 2008 EPA study confirmed that "the Irish climate is experiencing changes which have been found to be

consistent with those occurring at a global scale and there now is growing confidence that these changes are largely attributable to global warming" (Sweeney *et al.*, 2008, p. 3). The most up-to-date figures suggest that sea-level rose at a rate of c. 0.23 mm/yr for the period 1901-2010 at Belfast (Murdy *et al.*, 2015).

4.3 Projecting SLR – process-based and semi-empirical models

There are two main approaches climate scientists use to make projections of future SLR – process-based modelling and semi-empirical modelling. The two approaches differ quite significantly. Process-based general circulation models (GCMs) and Regional Circulation Models (RCMs) are based on mathematical descriptions of the climate system over a 3-dimensional grid, the vertical layers representing a multi-layered ocean and atmosphere. The models work by summing the estimated contributions from various sources (thermal expansion, the Greenland and Antarctic ice sheets, mountain glaciers, and land water storage) to SLC. Esteemed climatologist and former head of the NASA Goddard Institute for Space Studies, James Hansen, was among the first to project future warming trends using the process-based approach in his seminal 1988 paper at the Proceedings of the first North American Conference on Preparing for Climate Change, *Prediction of near-term climate evolution: What can we tell decision-makers now* (Hansen *et al.*, 1988). It has since become the most commonly used approach in climate modelling and has been adopted by the IPCC to make projections about future SLC. In this thesis, projections of sea-level rise are absolute, unless stated otherwise, meaning they do not take into account relative changes in the height of the land surface.

Overall, there is generally high confidence in the process-based approach because modelled results agree well with observations (Church *et al.*, 2013b) (figure 4.4). In a comprehensive study by Church *et al*. (2013c), process-based models were able to explain about 80% of the observed eustatic SLR between 1900-2010, 85% of the rise between 1961-2010, and 90% of the observed rise between 1990 and 2010. Uncertainties in process-based model predictions arise, in part, from a lack of good-quality observational data from which modelled data can be validated against and also a lack of quality environmental boundary data

(Devoy, 2014). These data are important because they are used as input for forcing the model at its boundaries. Such data can be difficult to obtain, given the scale at which the models are run. One example is glaciological data. There are few glaciers for which mass budget observations are available - about 380 out of more than 170,000 glaciers on Earth (Cogley, 2009; Arendt *et al.*, 2012). As such, there is disagreement in the literature about the contribution of glaciers to SLC. In the IPCC's fourth assessment report, glaciers were estimated to account for 28% of eustatic SLR for the period 1961-2003 (Bindoff *et al.*, 2007). Work by Meier *et al*. (2007), however, suggests that glaciers and ice caps accounted for 60% of observed SLR in 2006. Similarly, Moore *et al.* (2010) estimated the contribution to be 58% for the period 1955-2003. More recent work by Gardner *et al.* (2013) suggests that for the period 2003 to 2009, glaciers accounted for only about 30% of SLR observed, closer to the 37% for the period 1972-2008 estimated by Church *et al.* (2011). One reason for these discrepancies is explained by Moore *et al*. (2013, p. 1):

> *"Mountain glaciers, numbering hundreds of thousands, must be modelled by extensive statistical extrapolation from a much smaller calibration data set. Rugged topography creates problems in process-based mass balance simulations forced by regional climate models with resolutions 10–100 times larger than the glaciers."*

Glacier losses are also highly variable on annual to decadal timescales (Church *et al.*, 2013b), so the different periods over which the studies were conducted likely accounts for the discrepancies and is testament to the complexity of estimating short- to medium-term contributions from glaciers. And this is just the 'tip of the iceberg' – there are many other reasons why GCMs cannot explain 100% of observed SLR. Like all models, they are approximations of reality. Overall, however, there is generally a high degree confidence in the ability of the current generation of GCMs to reproduce observed changes in global mean sea-level with a reasonable degree of accuracy, although there is less confidence in the ability of GCMs to model the contributions of some individual components of the system. This is where the process-based modelling community is concentrating much of its efforts.

Significant progress has been made in process-based modelling in recent years, for example, with the adoption of the 'ensemble' approach by the IPCC. In the past, process-based models were developed by different research groups and run independently (Devoy, 2014). Now, future projections are the result of a collaborative effort. The Coupled Model Intercomparison Project (CMIP) is a collaborative climate modelling process coordinated by the World Climate Research Programme (WCRP) that provides a framework for coordinated climate model experiments. The CMIP archive is constantly being updated, "with modelling groups eager to contribute their best available data to the research community" (IPCC, 2014, para. 2). This approach has been shown to generally outperform individual models (Weigel *et al.*, 2008; Weigel *et al.*, 2010). The approach is useful in that it provides crosschecks and allows for greater homogeneity in spatial and temporal model outputs (Devoy, 2014). CMIP data can be downloaded freely from their website (http://cmip-pcmdi.llnl.gov/).

In addition to the adoption of the ensemble approach, significant progress in process-based modelling has been made in the way ice sheets are modelled, although, the results of these models remain somewhat controversial (Hansen *et al.*, 2015). In the IPCC's fourth assessment report, dynamic ice sheet processes were excluded altogether due to an inadequate understanding of ice sheet dynamics at the time. The decision to omit these processes was met with much criticism. Considerable progress has been made since then, though, which paved the way for the inclusion of these processes in the fifth assessment report (Church *et al.*, 2013). Dynamical ice sheet models still remain incomplete, though. According to Moore (2013 p. 4), this is due to "sparse observational data on grounding-line migration, the lack of realistic calving models, the largely unknown subshelf melting/aggregation distribution, and the poorly constrained basal drag and its spatiotemporal variability." As a result, at present, it is not possible to robustly quantify the probability of a potentially rapid increase in ice sheet outflow (*e.g.* due to the potential collapse of marine-based sectors of the Greenland and Antarctic ice sheets), and, as such, the IPCC did not include such a scenario in AR5. James Hansen *et al.* (2015), in a controversial discussion paper for the journal *Atmospheric Chemistry and Physics*, warn that their model

results suggest that such a scenario is possible for a 2 °C global warming and is highly dangerous. Kevin Trenberth of the National Center for Atmospheric Research, however, has criticised the study, arguing the experiments were "unrealistic" and that too many assumptions and extrapolations had been made for it "to be taken seriously other than to promote further studies" (Mooney, 2015, para. 25). While some may take issues with certain aspects of Hansen *et al.*'s recent work, most scientists, including the authors of the IPCC's fifth assessment report, agree that the IPCC estimates are conservative (Church *et al.*, 2013a).

It is relevant to stress that the experiments run using numerical models to project changes in future global sea-level are often time bounded and run over relatively short time periods – the blink of an eye, so to speak, in geologic time. The IPCC projections are provided at decadal to century scales - beyond that, confidence in model projections decreases due to uncertainties associated with processes that dominate over longer temporal scales (beyond 10^2-10^3 years). Hiatus periods in short- to medium-term processes (operating on the scale of decades to centuries) may also affect the ability of the current generation of GCMs to predict global sea-level changes because they are often calibrated using data collected over a similar scale.

It is also relevant to stress that the probabilistic nature of process-based models is a limitation of this approach to modelling in and of itself. It is assumed that when two models are fed the same inputs, or indeed even one model is run twice with the same inputs, the outputs would not be significantly different. This is usually, but not necessarily always, the case with a probabilistic approach. This applies not only to GCMs, but also to hydrodynamic and morphodynamic models such as those described in this thesis.

The second approach to making future projections of SLR is the semi-empirical approach. This approach was developed by Rahmstorf (2007) in response to the under-prediction of process-based models in the IPCCs third and fourth assessment reports, and was later refined by Vermeer and Rahmstorf (2009), Grinsted *et al.* (2010), Jevrejeva *et al.* (2012) and others. The semi-empirical approach is based on statistical relationships between observed global MSL and

surface air temperature (SAT) or total radiative forcing (RF). Instead of attributing SLR to individual components, semi-empirical models "regard a change in sea-level as an integrated response of the entire climate system" (Church *et al.*, 2013b; p. 1182).

The original Rahmstorf model is based on the assumption that the rate of SLR at any given time is proportional to the deviation from some global equilibrium air temperature at that time. This can be expressed as:

$$\frac{dH}{dt} = a(T - T_0) \tag{7}$$

Where H = sea-level; dH/dt = rate of SLR; T is the near surface air temperature; T_o = the equilibrium temperature; and a = a proportionality constant. T_o and a can be derived from past observational data based on this relationship. This equation can be integrated to give:

$$H(t) = a \int_{t_0}^{t} (T(t') - T_0)\, dt' \tag{8}$$

Where t' refers to the time variable. By substituting 21st century temperature rise scenarios into this equation, H(2100) can be calculated. This approximation is only valid in the short term (Rahmstorf posits "for a few centuries"), as beyond that, the relationship will tend toward a new equilibrium. Subsequent modifications include the addition of a rapid response term (Vermeer and Rahmstorf, 2009), the use of pre-historic data (dating back 2,000 years) to construct a more advanced model based on multiple parameters (Grinsted *et al.*, 2010), and a modification of Grinsted's model whereby sea-level was related to radiative forcing rather than temperature (Jevrejeva *et al.*, 2009).

Early estimates of SLR from semi-empirical studies were considerably higher than those of process-based studies. For example, Rahmstorf (2007) projected up to 140 cm of rise by 2100. More recently, though, estimates using both methods have converged. This was found to be largely due to (1) an upward shift in estimates from process-based models and (2) large uncertainties in process-based model estimates of ice sheets' mass loss (Moore *et al.*, 2013). The semi-empirical approach, however, has been widely criticised for its simplicity.

The presence of a linear relationship between global mean surface temperature and the rate of global mean SLC has been contested (Holgate *et al.*, 2007). Also, given a catastrophic event, such as the collapse of an ice sheet, the linear approach fails. At present, there is no scientific consensus with regard to the reliability of semi-empirical model projections, though they do offer an upper extreme estimate (Nerem, 2014).

Aside from the limitations associated with how we project SLC, whether it be using simple semi-empirical models or more complex process-based GCMs, there are other more general issues with future projections of SLR. One is that the magnitude of SLR over the next century is largely dependent upon GHG emissions (Church *et al*., 2013b). While some headway has recently been made in terms of attempting to limit global GHG emissions in the form of the Paris Agreement (COP21), no detailed timetable or country-specific goals for emissions have been laid out, therefore it remains difficult to make future projections with a reasonable degree of certainty. In response to such uncertainties, the IPCC's GCM projections of SLR are given according to possible climate futures (scenarios) that each depends on emissions in the years to come. In the fifth assessment report, these future scenarios are called Representative Concentration Pathways (RCPs) and supersede the earlier Special Report on Emissions Scenarios (SRES) trajectories published in 2000 (Nakicenovic *et al*., 2000). The categories, RCP2.6, RCP4.5, RCP6, and RCP8.5, reflect radiative forcing values in the year 2100 relative to pre-industrial values (see van Vuuren *et al*. (2011) for a comprehensive overview). While these scenarios are useful for setting a standard for communicating and comparing model results, they serve only as a guideline and cannot necessarily be taken to represent the complexity of humanity's possible future emissions. In addition, the potential amount of methane emissions from melting permafrost is unclear, further complicating future climate projections (Schuur *et al*., 2015).

4.3.1 21st century projections

As of its fifth assesment report (AR5), the IPCC projects a rise in GMSL of between 0.26 and 0.98 m by 2100 (Church *et al.*, 2013b), although, many experts consider these estimates to be somewhat conservative (Hansen, 2007; Horton *et*

al., 2014; Jevrejeva *et al.*, 2014). For each of the emissions scenarios outlined in AR5, for the period 2081-2100 compared to 1986-2005, process-based projections of GMSLR are as follows:

- 0.26 to 0.55 m for RCP2.6
- 0.32 to 0.63 m for RCP4.5
- 0.33 to 0.63 m for RCP6.0
- 0.45 to 0.82 m for RCP8.5, with the rise by 2100 projected to be 0.52 to 0.98 m at a rate of 16 mm per year

As stated earlier, these figures do not take into account the potential collapse of land-based sectors of the Antarctic ice sheet. The IPCC states with medium confidence (defined as about a 1 in 2 chance) that the magnitude of SLR induced by such a scenario would not exceed several tens of cm during the 21st century (Church *et al*., 2013). However, there is a lack of scientific consensus on the probability of its collapse.

Regional scale process-based modelling has been undertaken to project SLR for Britain and Ireland (Lowe *et al.*, 2009). This type of modelling is based on the downscaling of GCMs (usually the multi-model ensembles used by the IPCC). Prior to presenting regional-scale climate model projections, it is important to stress that GCM estimates of SLR are *absolute* - they do not take into account relative changes in the height of the land surface due to, for example, glacial isostatic adjustment. In Ireland, this is especially important because there is significant spatial variation in rates of RSLR across the island due to these crustal movements, thus vertical land movement must be taken into account when projecting SLC at the regional or local level[6].

The UK Climate Projections scientific report (CP09) commissioned by the UK Meteorological Office Hadley Centre (Lowe *et al.*, 2009) is, at present, the primary source of regional SLR estimates for Britain and Ireland. The work undertaken in this report took into account absolute SLR around the British Isles, originating from IPCC AR4 GCM projections, together with estimates of vertical land movement to calculate relative SLC. Figure 4.5 shows derived projected

[6] For more on relative sea-level changes in Ireland, see Lambeck and Chappell (2001) and Devoy *et al.* (2006).

RSL increases under the IPCC AR4 medium emissions scenario for the year 2095. These results suggest that for this scenario, the southwest coast of Ireland can expect a relative SLR of roughly 45-50 cm, approximately 20 cm more than the coast of Northern Ireland. Low and high emissions scenarios were also simulated. The results of all three scenarios have been made available to the public via an online interface (http://ukclimateprojections-ui.metoffice.gov.uk/ui/start/start.php), although RSL projections for the Republic of Ireland are not available. It should be noted that the UKCP09 projections were based on the IPCC fourth assessment report projections of absolute SLC, which were made prior to the inclusion of dynamic ice sheet processes and are thus associated with a higher degree of uncertainty than those reported in the 5th assessment report. An update based on these new figures is due in 2018.

More recently, Grinsted *et al*. (2015) published projections for 21st century relative sea-level rise in Northern Europe, including for the British Isles. They calculated relative sea-level rise using a probabilistic method based on the IPCCs RCP 8.5 (high emissions) scenario. The median relative sea-level rise projection, which the authors consider their "best guess," for Dublin (nearest location to the study area) was 0.69 m.

Given the high level of uncertainty with regard to projections of sea-level rise, it is difficult to say with certainty what can be expected for Inch and Rossbehy. Based on the UKCP09 projections, it appears that relative sea-level in the region could reasonably be on the order of 45-50 cm higher by the end of the 21st century. This figure was therefore used as a guide in the development of the numerical modelling scenarios presented in chapter 10. The Grinsted *et al*. (2015) study was published after the modelling was completed, and was not taken into account in the planning of the scenarios.

4.4 Impacts of SLR and vulnerability of soft coasts

Climate change induced SLR represents a threat to both natural and human systems along the world's coasts in the next century and beyond, especially on soft coasts (Hinkel *et al.*, 2013; Wong *et al.*, 2014). These include beaches, barriers, sand dunes, wetlands, estuaries and lagoons, whose hinterland is often low lying and susceptible to submergence and flooding. In addition, changes in

sea-level will have an impact on sediment distribution and thus patterns of erosion. The IPCC states with high confidence that beaches, sand dunes, and cliffs currently eroding will continue to do so under increasing sea-level, although beach response to SLR may be more complex than a simple retreat (Irish *et al.*, 2010; Wong *et al.*, 2014).

SLR increases the risk of coastal inundation, which usually occurs during extreme events when elevated water levels, due to storm surge and increased wave heights, result in the overtopping of barriers and inundation of dry land surface. The loss of property and life that occurred during extreme events in the last decade, such as Hurricane Katrina in the Gulf of Mexico (2005), Hurricane Sandy along the Eastern US seaboard (2012), and Typhoon Haiyan in southeast Asia (2013), underscore the vulnerability of coastal regions to storm surges and flooding (FitzGerald *et al.*, 2008; Halverson and Rabenhorst, 2013; Tajima *et al.*, 2014). In the absence of adaptation, soft coasts may become less resilient to changes in sea-level and coastal flooding is likely to become more frequent (Wong *et al.*, 2014).

In addition to flooding during extreme events, many coasts are becoming increasingly threatened by tidal flooding. Tidal flooding, sometimes called 'nuisance flooding,' is flooding that occurs during extreme high tides. In the US, tidal flooding occurs more frequently and for longer periods today than ever before (Ezer and Atkinson, 2014; Spanger-Siegfried *et al.*, 2014; Sweet *et al.*, 2014). In a study on changes in nuisance flood frequency around the East, Gulf and West coasts of the United States, Sweet *et al*. (2014) found that the frequency of nuisance flooding increased from about once every one to five years in the 1950s to once every three months by 2012. During this period, the rate of eustatic SLR rose from approximately 1.7 mm/yr to 3.2 mm/yr (Church *et al.*, 2013b).

Tidal flooding is a problem in Ireland too, where it is often exacerbated by high rainfall, and is especially a problem for cities like Cork, Dublin, and Galway, which are located at the mouths of rivers. In February 2014, several businesses along one of Cork city centre's main shopping streets were flooded due to tidal flooding (Roche, 2014). No similar work to that done in the USA is known to

have been carried out on examining whether or not there has been a change in the frequency or duration of tidal flooding in Ireland.

Relative sea-level rise due to dynamic adjustments of the land surface (*e.g.* land subsidence) is also important in terms of understanding the risk of coastal inundation at local to regional scales. Changes in relative sea-level due to such adjustments may not be related to climate warming at all. For example, the intensive pumping of groundwater or hydrocarbons is common in many coastal regions and can noticeably affect relative sea-levels on relatively short time scales (Church *et al*., 2013). In New Orleans, LA, between 1951-1955 the ground surface sank at a rate of 5 mm per year (Burkett *et al*., 2002). This subsidence is, in part, to blame for increasing the overall vulnerability of the coastal area to hurricanes, such as Hurricane Katrina in 2005 (Colten *et al*., 2008).

Dry land loss due to coastal erosion is another major concern associated with SLR (de la Vega-Leinert and Nicholls, 2008; Devoy, 2008; Williams, 2013). In a major study on the global effects of climate-induced SLR on the erosion of sandy beaches, Hinkel *et al.* (2013) found that in the absence of adaptation, 6,000-17,000 km^2 of land may be lost during the 21st century. This could potentially lead to the forced migration of 1.6 – 5.3 million people (Hinkel *et al.*, 2013). These numbers, however, should be treated cautiously, as they are not intended for coastal management purposes, which require more complex morphodynamic methods, but to help quantify the impacts of climate change on a global level.

To understand the impacts of SLR on coastal erosion at regional and local scales, either numerical morphodynamic models, such as Delft3D or MIKE21, or simple geometric profile relationships, such as the Bruun rule (1962), are often used. There is disagreement in the literature on the value of each of these techniques. Process-based numerical models, which are based on mathematical descriptions of physical processes, are widely used to study hydrodynamic and sediment transport problems and have been shown to simulate flows and transports reasonably well (Roelvink *et al.*, 2009; Remya *et al.*, 2012). However, the accuracy of these models depends on the quality of the data available to constrain

the model structure and parameters, which, unfortunately, is often lacking, so process-based model results must be treated with caution (Hanson *et al.*, 2003; Hutton, 2012).

On the other hand, simple deterministic Bruun-based models are commonly used as an alternative, although they have received a considerable degree of criticism. Critics of deterministic Bruun-based models argue that many of the assumptions behind the Bruun Rule are known to be false (Cooper and Pilkey, 2004). For example, it assumes no net longshore transport, which almost never occurs in nature. Ranasinghe *et al*. (2012) have further argued that the relatively low quantitative accuracy and robustness associated with the Bruun Rule render it unsuitable for local scale assessments in which reliable estimates are required. For example, Ranasinghe and Stive (2009) have shown that when uncertainties associated with the individual terms of the Bruun Rule are taken into account, there was a potential variability of up to 4,000% in predictions. Others (Rosati *et al.*, 2013), however, have claimed that modifications to the original Bruun Rule can extend its usefulness, although they also admit that the Bruun Rule does not apply to all coasts.

Further complicating our understanding of the vulnerability of the coastal zone to flooding and erosion is the impact of human pressures on coastal systems. Human pressures have resulted in the loss or degradation of many natural protective barriers, such as coastal sand dunes (Pilkey and Young, 2009). The protective qualities of soft coasts are well established (Hanson *et al.*, 2002; Cooper and Pilkey, 2012). For example, during Hurricane Sandy, the presence of dunes significantly reduced the likelihood of damage to the New Jersey coast (Ozbas *et al.*, 2013). Also, after the 2004 Indian Ocean tsunami, mangrove forests and vegetated dunes were found to have mitigated damage associated with the tsunami in Sri Lanka, Thailand, and India (Kathiresan and Rajendran, 2005; Tanaka *et al.*, 2007; Mascarenhas and Jayakumar, 2008). Despite this, many of the world's dune systems and mangrove forests suffer from degradation and/or complete removal at the expense of coastal development, putting the coastal hinterland at further risk of SLR.

Finally, potential changes in the frequency and intensity of storms, including that of surge and storm waves, is predicted to further exacerbate the impacts of relative SLR (Wong *et al.*, 2014). Unfortunately, at present, there is still a high degree of uncertainty with regard to potential changes in storm intensity and frequency. This is because of the difficulty associated with modelling the complex three-dimensional features of atmospheric circulation associated with cyclones (Ulbrich *et al.*, 2009). At present, studies seem to suggest that there is a likely poleward movement of extra-tropical cyclones (Wang *et al*., 2008). Most models suggest that under climate warming, the number of intense cyclones over the north Atlantic and British Isles will increase, but this appears to be dependent upon core pressure (Ulbrich *et al.*, 2009).

In summary, understanding the impacts of SLR on soft coasts is not straightforward due to the myriad of uncertainties associated with predictive models and human behaviour. However, as new datasets become available and models improve, so will our understanding of the response of coastal systems to SLR, and thus our ability to adapt and manage the risks related to SLR. For a comprehensive review of the impacts of SLR on coasts and low-lying areas, see Wong *et al.* (2014).

4.5 Future implications for Ireland

Relative to other European countries, Ireland as a whole is seen as having an overall low vulnerability to the impacts of SLR due to its relatively small population and the presence of natural protective biogeophysical features at the coast such as resistant rocky cliffs (Devoy, 2008). That said, at local scales, the situation is more complex. Urban centres, including the major cities of Cork, Limerick, Dublin and Galway, are likely to suffer the greatest economic losses as a result of SLC (Irish Committee on Climate Change, 2004). Ireland's wetlands are also particularly vulnerable - it is estimated that a 1 m SLR could result in a loss of about 30% of Irish wetlands (Devoy, 2008). Devoy (2008) provided an appraisal of the main susceptibility-resilience factors to SLR in Ireland. These are summarized (bold) and commented upon as follows:

- **"Natural" Influences**
    - **Projected changes in storm frequency and intensity will increase Ireland's exposure to high wave energy, thus increasing the risk of erosion and flooding.** In an EPA funded study on future Irish climate conditions for the period 2021-2060, McGrath *et al*. (2005) found that the frequency of intense extratropical cyclones (core pressure >950 hPa) over the North Atlantic in the vicinity of Ireland will increase by 15%, with greater increases in winter and spring. This is potentially worrying, as erosion and flood damage sustained by lands managed by the Office of Public Works (OPW) after the extreme events associated with the winter 2013-2014 storms alone cost upwards of €110 million (National Directorate for Fire and Emergency Management, 2014).
    - **SLR will exacerbate the effects of freshwater flooding at estuary mouths, particularly where river floods coincide with marine surges.** Freshwater flooding occurs when the ground becomes saturated due to heavy precipitation. Sweeny *et al.* (2008) posit that changes in precipitation may well be the most important aspect of future climate for Ireland. Their regional climate modelling work suggests that Ireland can expect a 10% increase in winter rainfall by the 2050s, with a 12-17% reduction in summer rainfall. Plugging this into a catchment model, they found that changes in stream flow mean that what is presently a 10-year flood reduces to a 3- to 7- year event in most catchments by the 2050s. More alarmingly, the 50-year flood became a 6- to 35- year event in all but one catchment in the study. These results, however, should be interpreted cautiously, as the regional climate model used to make the projections of changes in precipitation was shown to be unable to predict past trends with a reasonable degree of accuracy (Sweeny *et al.*, 2008). Other studies though (*e.g.* McGrath *et al.*, 2005) do support the general conclusion that an increase in winter precipitation will result

in more frequent and intense discharge episodes at the mouths of estuaries in Ireland, raising the risk of future flooding.

- **Projected changes in important boundary controls on coastal systems (*e.g.* winds, storm magnitude, and rainfall) will affect patterns of sediment distribution, although it is unclear what net effect this will have on Ireland's coastline, which is already somewhat tuned to natural extremes.** The 2008 breaching and subsequent erosion of Rossbehy is testament to the degree to which changes in sediment distribution can affect the shape of the local coastline in the short-term. Less is known about the long-term impact of changes in sediment distribution on the shape of the coast as a result of future SLR.
- **Sediment deficit occurring as a result of both natural and human influences limits onshore coastal movements and will be exacerbated under SLR. This means that in the absence of a strategic coastal zone management (CZM) policy, there is a high potential for the complete loss of many sedimentary systems.** There are many examples, both in Ireland and abroad, of how shoreline stabilization and other human influences have resulted in the partial or complete loss of adjacent beaches (Pilkey and Wright, 1988; Cooper and Pilkey, 2012). One example in Northern Ireland is Portballintrae, Co. Antrim. Once a popular seaside destination, Portballintrae has suffered from progressive sediment loss over the last 116 years. This was likely due to the installation of a pier in its western section, which interrupted natural sediment cycling and resulted in beach material being lost to deeper water (Jackson, 2012). The subsequent narrowing of the beach is illustrated in figure 4.6.
- **The natural resilience of marshes, which, at present, remain in good condition, to inundation may be reduced as a result of SLR together with human pressures on coastal land use.** Marshes are susceptible to SLR because their vertical accretion rates are limited and, therefore, they are at risk of drowning (FitzGerald *et al.*, 2008). Understanding marsh loss is important because it may promote

hydrodynamic and sedimentological change, which can affect entire coastal systems. Loss of biodiversity is also a concern.

- **"Human Influences"**
  - **Population pressures in the form of land reclamation, built coastal structures, agriculture and degradation of coastal dune and other biogeophysical coastal systems have affected sedimentation patterns along many parts of the Irish coast. However, the lack of negative impact of Ireland's earlier larger population and usage of the coast suggests that the Irish coast has a high natural potential for absorption of human pressure and recovery from change. Also, large parts of Ireland's coast remain sparsely populated, potentially increasing Ireland's ability to cope with human pressures under SLR.** Despite this, the Irish coastal zone is still economically vulnerable to relatively small SLR and associated increases in storm surge activity. For example, Sweeney *et al.* (2013) reported that c. 350 km^2 of the Irish coast is exposed under a 1 m rise, with potential property insurance losses due to flooding in at-risk areas potentially amounting to over €1 billion**.**
  - **Poor governance of Ireland's coastal zone (*e.g.* overlapping and inefficient administrative structures) limits Ireland's adaptive capacity for dealing with the impacts of SLR.** The current absence of any national coastal management policy in Ireland has resulted in the promotion of reactive rather than proactive responses to coastal issues, such as erosion, flooding, and damages related to extreme events. These issues are likely to intensify under SLR, further highlighting the need for better governance. Many (*e.g.* Devoy, 2008; Falaleeva *et al.*, 2011; Gault *et al.*, 2011; Kopke and O'Mahony, 2011) have called for the implementation of integrated coastal zone management (ICZM) policies in Ireland to more effectively deal with coastal issues such as erosion and flooding.
  - **A limited awareness of issues related to coastal vulnerability hampers society's ability to cope with the impacts of SLR. Ireland's reactive rather than proactive policies in response to**

erosion management, for example, persist because coastal authorities feel pressured to respond to the public's demand for action, even when those responses are inappropriate. Coastal researchers in Ireland generally call for more strategic planning based on a scientifically rigorous approach rather than public demand (*e.g.* Gault *et al*., 2011). Researchers in UCCs Coastal and Marine Research Centre (CMRC, now MaREI), for example, have made recommendations to relevant authorities on numerous occasions to this effect (*e.g.* Falaleeva *et al.*, 2011).

- **Existing coastal defenses in Ireland, which protect less than 4% of the Irish coast, are, for the most part, insufficient to cope with future SLR. Relevant authorities tasked with maintaining these structures are presently under-resourced and underfunded.** This essentially highlights issues related to the development and maintenance of coastal defenses in Ireland. Given the present inability of authorities to cope with many coastal issues in Ireland (O'Connor *et al*., 2010), a reassessment of the appropriateness of building and maintaining current and future coastal defenses is essential under future SLR.

The overall picture here is that the impacts of SLR in Ireland will be exacerbated by other natural, climate change and weather related phenomena, and our ability to adapt to these changes is largely dependent on the attitudes of society towards integrated coastal zone management (ICZM) (Devoy, 2008). That said, a better understanding of the potential impacts of SLR on Irish coasts (gained from studies such as this one) is ultimately required to inform sound management policy.

5 Storms

Storms are important drivers of morphological change in morpho-sedimentary systems, and as such, understanding their significance in this context warrants attention (Stone and Orford, 2004). This chapter will firstly provide quantitative definitions of "storms" as defined in the published literature to demonstrate the complexities associated with defining/characterising storm events, each of which is unique. Section 5.2 examines the ways in which we assess the impacts of storms on coastal barriers, using as examples the theoretical and methodological contributions made by Sallenger (2000) and Cooper *et al.* (2004). This section provides a critical review of the various approaches used to investigate the morphological and hydrodynamic processes in operation during storm events, citing examples where relevant. Finally, the results of recent climatological studies of past and potential future changes in storminess in the North Atlantic that are of particular relevance to this study are examined in section 5.3.

5.1 Definitions

No one definition can describe a "storm" or "extreme event" and therefore, several metrics are used in the published literature, each carefully chosen according to its relevance to a given study. Storm definitions are usually based on wind speeds, minimum sea-level pressure, event duration, water levels (wave heights and/or surge heights), or some combination of these. According to the Beaufort Scale, storms occur when mean wind speeds reach 24.5 m/s. This definition, however, is often too generic for scientific research purposes and is therefore not usually used to define events in the coastal literature.

Sometimes events are defined based on the return period of extreme winds (*e.g.* Abild *et al.*, 1992) or water levels (*e.g.* D'Onofrio *et al.*, 1999). Return periods, however, can be misleading, in that often there is an insufficiently long record over which data are available to reasonably estimate the likelihood of the recurrence of the event (Stedinger *et al.*, 1993). Also, it must be assumed that the probability of tevent occurrence does not vary over time, which may be invalid, especially given future projections of climate change.

Boccotti (2000) defines storms (more specifically, "sea storms") based on wave heights. According to this definition, events are defined as a sequence of sea

states in which significant wave height, $H_s(t)$, exceeds a fixed threshold, h_{crit}, and does not fall below this threshold for a continuous time interval greater than 12 hours. The fixed threshold is equal to $1.5*(H_{sig}(t))$, or 1.5 times the mean annual significant wave height. This definition may be useful when water levels are important, although it does not take into account tidal state or surge height (a function of atmospheric and sea-level pressure).

MacClenahan *et al.* (2001) took an empirical approach to the classification of storms. They defined events on the basis of wind speed and duration thresholds extracted from coastal weather station data around Ireland. Seven types of extreme storms were classified on this basis:

- storms with mean wind speeds of 50 knots (25.7 m/s) lasting for at least one hour;
- storms with mean wind speeds of 60 knots (30.9 m/s) lasting for at least one hour;
- storms with mean wind speeds of 30 knots (15.4 m/s) lasting for at least 48 hours;
- storms with mean wind speeds of 40 knots (20.6 m/s) lasting for at least 24 hours;
- storms with mean wind speeds of 40 knots (20.6 m/s) lasting for at least 10 hours;
- storms with mean wind speeds of 40 knots (20.6 m/s) lasting for at least 5 hours; and
- storms with mean wind speeds of 34 knots (17.5 m/s) lasting for at least 24 hours.

Again this approach is limited in that tidal state and surge height are not accounted for.

Carter and Stone (1989) defined storms in geomorphological terms - on the basis of their potential to cause foredune erosion. They identified three types of dune eroding events from specific combinations of wave, wind and tidal conditions in Northern Ireland. They found that a wind speed threshold of 25 knots (12.9 m/s) was required to qualify as an 'erosive storm.' This study was specific to sand

dune cliffs at Magilligan, Northern Ireland, and as such, the published 'threshold' may be considered site specific, although their methodology may be applicable elsewhere.

The examples highlighted here represent a small subset of the various ways of describing and characterising storm events in the coastal literature and reflect the need for careful consideration of how storms are defined in a given study where storms are of relevance. This is especially important to consider when designing laboratory and empirical studies on the impacts of storms on beach/barrier morphodynamics and in mathematical model simulations of same.

5.2 Assessing and predicting the morphological impacts of storms on coastal barriers

Two approaches of potential relevance to this research for assessing the impacts of storms on coastal barriers are that of Sallenger (2000) and of Cooper *et al.* (2004). Sallenger (2000) developed a scale for categorizing storm impact level on barrier islands based on the observed response of two coastal barriers – Duck, North Carolina, USA and Isles Dernieres, Louisiana, USA – to tropical and extratropical storms. The scale consists of four regimes, each representing a different level of morphological impact. Impact level one is the swash regime. This occurs when storm run-up is confined to the foreshore and thus there is no net change in dune morphology. Impact level two is the collision regime. This is when wave run-up exceeds the base of the foredune ridge and there is a net erosion of the dune. Impact level three is the overwash regime, which occurs when wave run-up overtops the berm or foredune ridge, resulting in a net landward migration of the barrier island. The most severe regime, impact level four, is the inundation regime. This occurs when there is complete submergence of the barrier island, resulting in a net landward transport of sand across the barrier island. While overall this model is conceptual, each regime can be described mathematically in terms of the high (R_{high}) and low (R_{low}) elevations of the landward margin of swash (relative to a fixed vertical datum) and the base (D_{low}) and crest (D_{high}) of the foredune ridge[7]. The Sallenger Scale is widely used in the coastal literature and serves as a useful guide for the systematic

[7] See Sallenger (2000) for mathematical descriptions of storm impact regimes.

documentation of storms on barrier islands (Shroder *et al.*, 2014). In some cases, however, the Sallenger Scale can be considered *too* generic. For example, it does not address the fact that barrier islands may (and often do) respond uniquely to events, especially those at the extreme end of the spectrum. In their assessment of storm impacts on the high-energy west coast of Ireland, Cooper *et al*. (2004) found that beach and dune systems respond variably to storm forcing at the decadal scale and also that in many cases, storm susceptibility is site-specific. Using a combination of historical records, meteorological records, field observations, and wave modelling, Cooper *et al*. (2004) attempted to relate event characteristics to storm response at both instantaneous and historical timescales at several locations on the western Irish seaboard. This approach differed from Sallenger's in that it highlighted the variability of storm response over a relatively short coastline. Sallenger attempted to control for spatial variability in his model by choosing barrier islands that represented "end members of relief", but how representative these are of the barriers in between is questionable. Had the response of other types of barriers been considered, the conceptual framework of the Sallenger Scale might have been improved upon.

Predicting the morphological response of barriers to storms is a topic that receives far more attention in the coastal geomorphology and engineering literature than simply assessing storm impacts. Just searching the phrase "predicting beach response to storms" in GoogleScholar produces in excess of 31,600 results[8]. Given the breadth of the subject matter, the following discussion provides a general overview of the most widely used set of approaches to making such predictions. Specifically, it describes and critiques the various types of models used to investigate the morphological and hydrodynamic processes in operation during storm events.

Three general types of models are used to investigate coastal morphodynamics – conceptual, empirical, and mathematical models. Conceptual models are mostly descriptive in nature and often based on observations (conceptual empirical models). A succinct definition is given by Pilkey and Pilkey-Jarvis (2007, p. 27):

[8] As of 25 August 2015.

> *"A conceptual model is a qualitative one in which the description or prediction can be expressed as written or spoken words or by technical drawings or even cartoons. The model provides an explanation for how something works—the rules behind some process."*

Orford *et al*.'s (1999) model of the meso-scale evolution of Inch, which is said to be controlled by episodic high-magnitude, low-frequency events triggering rapid erosion followed by gradual recovery, is an example of a conceptual morphodynamic model. Another example is Sallenger's storm impact scale (described previously), which, given that storm characteristics and antecedent conditions are known, may be used to forecast the potential impacts of events as storms approach the coast. The conceptual approach to modelling (as opposed to the mathematical approach) has been championed by Cooper and Pilkey (2004) and Pilkey and Pilkey-Jarvis (2007). They argue that field observations, qualitative modelling, and past experience can be sufficient predictors of the future behaviour of beaches. This common-sense approach is often abandoned for practical engineering applications in favour of more complex quantitative mathematical models, which, in many cases, have been shown to perform poorly and/or have been applied inappropriately. The experience of a beach nourishment project at Folly Beach, South Carolina, USA is one example. In 1993, the US Army Corps of Engineers spent approximately $12 million on nourishing five miles of Folly Beach. The beach nourishment interval predicted by the mathematical models of the Corps was eight years. Without even the passage of a significant storm, most of the dry nourished beach had disappeared by 1995, only two years later (Bush *et al.*, 1996). According to Pilkey and Pilkey-Jarvis (2007), an overreliance on mathematical models to predict the outcome of natural and beach barrier-island processes by the Corps has consistently led to many similar costly mistakes. Cooper and Pilkey (2004) do not contest that mathematical models are useful. On the contrary, they suggest that a composite approach involving the use of a combination of model and field observations "could provide the most reliable, non-quantitative answers for engineering

purposes" (Cooper and Pilkey, 2004, p. 643). While coastal geomorphologists generally favour this approach, it has not been widely adopted by engineers.

A second approach to modelling storm impacts is the empirical approach. This approach is purely experimental and based on field and/or laboratory observations. While some empirical models may be conceptual, not all are. For example, the Bruun Rule (chapter 2) is an example of an empirical mathematical model. Empirical models of storm response are usually built on observed relationships between morphology and storm characteristics and/or antecedent morphology. One example is that of Pepper and Stone (2004), who attempted to relate sediment transport on subtidal shoals to storm characteristics during cold front passages on the inner shelf of the northern Gulf of Mexico. They measured sediment transport and direction and found qualitative linkages between observed transport patterns and measured waves and currents. Another example is that of Castelle *et al*. (2015), who used a combination of satellite imagery, beach surveys, and wave data to link patterns of beach and dune erosion to antecedent morphology and storm wave characteristics (period and angle of incidence). Other examples of empirical models of storm response include those of Houser *et al*. (2008), Forbes *et al*. (2004), Regnauld *et al.* (2004) and Spencer *et al*. (2015). Often the relationships between morphological observations and storm characteristics are too complex and cannot be explained empirically. In this case, further questions can be addressed using mathematical modelling, assuming boundary parameters are known.

Mathematical models are based on mathematical descriptions of one or more components of the coastal systems and represent one of the most widely used sets of techniques for investigating coastal processes. These models are useful for studying hydrodynamic (waves and currents) and sediment transport processes and their interaction with bottom topography changes (morphodynamics) at various space and time scales (de Vriend, 1991). They can be especially useful when observational data are unavailable or unobtainable in a practical sense, thus extending inquiry beyond observation alone and allowing for the exploration of landscape dynamics over a range of spatial and temporal scales (Hutton, 2012). Two of the most important considerations when using this approach are, firstly, the model predictions should have been tested against

observations (model validation) to ensure model reliability and second, model input should be accurate and of sufficient resolution with respect to the processes under investigation (Bird, 2011).

Mathematical models often form the basis for public policy decisions, especially in the US and the Netherlands. Some (Thieler *et al.*, 2000; Cooper and Pilkey, 2004; Pilkey and Pilkey-Jarvis, 2007) argue that mathematical models are not reliable enough for such applications. According to Thieler *et al.*, (2000), this is because: (1) many of the assumptions used in these models are invalid in the natural world; (2) many relationships used in mathematical relationships are of questionable validity (*e.g.* modelled data show poor agreement with observations); (3) model calibration and verification are often used incorrectly as an assertion of model veracity; (4) project monitoring, hind sighting and objective review of engineering projects reliant on these models are often absent, inhibiting an evaluation of the predictive success of the models; and (5) the degree of uncertainty associated with many mathematical models is insufficient for the quantitative prediction of coastal evolution at engineering time and space scales. While Thieler *et al*. (2000) cite many examples of poor mathematical modelling experiments (poor data, poor by design, *etc.*), there are many examples in the coastal literature of good quality, rigorous mathematical modelling studies which appropriately acknowledge the shortcomings of their approach (*e.g.* Roelvink *et al.*, 2009). These models are constantly being improved upon and represent, at the very least, a worthwhile avenue for investigating the dynamics of complex coastal systems.

Two general approaches to mathematical modelling are the deterministic approach and the probabilistic approach. The deterministic approach is based on the principles of fluid mechanics. Deterministic models do not include information on the uncertainty of model predictions. They are often used in conjunction with laboratory experiments, which allow selected parameters to be held constant while one can be varied at a time (Goudie, 2004). One example is the CERC formula for predicting longshore sediment transport on sandy beaches (US Army Corps of Engineers, 1984). The formula links total longshore transport in the surf zone to wave breaking angle and sediment porosity. The CERC formula is deterministic because it reduces sediment transport to a single

number that can be determined by a single set of inputs. The deterministic approach is inherently limited in that it lacks the ability to make stochastic predictions. The probabilistic approach to mathematical modelling, on the other hand, includes statistical features, which means uncertainties associated with predictions can be quantified. Unlike deterministic models, given the same set of inputs, repeated probabilistic simulations will yield different outputs – *e.g.* an element of randomness is built in to these models. The probabilistic approach has been shown to perform well in simulating beach erosion due to extreme storms. For example, Callaghan *et al.* (2008) showed that this approach was capable of simulating wave climate and extreme beach erosion at Narrabeen Beach, Australia for return periods of up to 10 years. Others have similarly found that this approach is well-suited to the forward modelling of coastal beach and cliff erosion, with up to 90% agreement between observed and modelled data (Ruggiero *et al.*, 1996; Hapke and Plant, 2010).

Sometimes hybrid modelling approaches (semi-probabilistic or semi-deterministic) are employed in mathematical modelling. This is where a set of deterministic equations is used in combination with input from probability distribution functions such that the variability of the system response can be evaluated (Schwartz, 2006). Dai (2011) employed such an approach in his evaluation of long-term barrier island responses to storms and SLR. Given the potentially endless combinations of storm variability and SLR scenarios in the next century, it is difficult to make meaningful projections of long-term barrier island response to these forcings. So Dai (2011) used the probabilistic Monte-Carlo simulation approach in an attempt to quantify the uncertainty associated with such projections. He simulated 1000 realizations of hurricane magnitude, frequency and track and assessed scenario uncertainty by considering five plausible rates of SLR. Uncertainties were found to increase with time and rate of SLR. Unfortunately, however, as Larson (2006) points out, hybrid approaches like those of Dai (2011), are underexploited in coastal applications, with forecasts of coastal evolution all too often being made without any attempt to assess the uncertainty in the predictions.

Another type of mathematical model commonly used to study coastal systems is the process-based model. Process-based models are based on a reductionist

approach, by which the complex coastal system is seen as the sum of its individual constituents. These models simulate forcing by waves and/or currents and resultant sediment transports and morphology. Process-based models of storm impacts on barriers (*e.g.* XBEACH) have been shown to perform reasonably well (Roelvink *et al.*, 2009; Vousdoukas *et al.*, 2011; Terlouw, 2013; Williams *et al.*, 2015). Over longer timescales, though, process-based results are subject to a considerable degree of uncertainty because there are still major gaps in our understanding of long-term coastal behaviour (Southgate and Brampton, 2001; Hanson *et al.*, 2003).

Mathematical models can be numerical or analytical. Numerical models involve iterative calculations at specified time steps, while analytical models have a closed form solution (Euler's method). Numerical models are advantageous over analytical models in that they are often better able to deal with systems of greater complexity (*e.g.* natural systems), whereas analytical solutions to equations describing such systems can often become fairly complicated. Popular numerical models of coastal dynamics include XBEACH (Roelvink *et al.*, 2010), Delft 3D (Deltares, 2011a; Deltares, 2011b), and MIKE21 (DHI, 2005). Although analytical models have also been shown to perform reasonably well in some cases (*e.g.* Larson *et al.*, 2004), most coastal modelling is presently performed using (usually process-based) numerical models.

5.3 Past and future changes in storminess in the North Atlantic

The identification of changes in patterns of storminess is difficult due to a lack of good-quality long-term records and the inherent variability in meteorological data, making it hard to identify trends against noise. For such studies, statistical techniques are usually employed. Future changes in storminess are forecast using regional scale GCMs. Good-quality records are also required for these studies (*e.g.* for hindcasting studies, which are essential for model validation). This section presents substantive findings of studies on past and future changes in storminess in the North Atlantic that make use of these techniques. Because the aim of this study is to evaluate the potential impacts of storms under 21st century SLR, a brief explanation of these findings is relevant here.

The IPCC, in its 5th assessment report, presented an updated assessment of past trends in storminess. Globally, it is likely that a poleward shift in extratropical cyclones has occurred, but there is overall low confidence in intensity changes (Hartman *et al.*, 2013). Uncertainties in the strength of these trends arise from inconsistencies between studies and a lack of good quality long-term observational data. There is, however, very strong evidence that regional storm activity has increased in the North Atlantic - in fact, the IPCC considers it "virtually certain that the frequency and intensity of the strongest tropical cyclones in the North Atlantic has increased since the 1970s" (Hartman *et al.*, 2013, p. 162). There is less certainty with regard to extratropical cyclone activity, although many studies suggest that wintertime cyclones in the high latitude Atlantic have become more frequent and intense during the last 60 years (Schneidereit *et al.*, 2007; Raible *et al.*, 2008; Vilibić and Šepić, 2010).

There has been considerable attention in the literature recently paid to changes in the North Atlantic Oscillation (NAO) and its influence on the Atlantic Meridional Overturning Circulation (AMOC). The NAO is an important control on storm track in the North Atlantic. During a positive NAO index phase, there is a stronger than usual subtropical high pressure centre and a deeper than normal Icelandic low. This phase is associated with more northerly storm tracks, *e.g.* tracks with Ireland in their path. During a negative NAO phase, there is a weak subtropical high and a weak Icelandic Low. This results in weaker and less frequent winter storms with more west-east tracks (Bell and Visbeck, 2003). Changes in the NAO therefore affect Ireland and, in some cases, they may be correlated with changes in regional storminess (Sweeney *et al*., 2008). Due to the extreme variability of the NAO, it is extremely difficult to forecast. It is uncertain, at present, how climate change will affect the NAO.

At the regional scale, changes in storminess along the Atlantic coastlines of Europe have been evaluated by Lozano *et al.* (2004). They examined high-resolution (hourly to 6 hourly) coastal meteorological data from eight sites along the Atlantic coast of Europe. From 1940-1998, analysis shows the occurrence of more stormy winters and calmer summers for Ireland, which may be linked to a northward displacement in the main North Atlantic cyclone track. In addition, a positive trend in the winter wind climate for all three Irish meteorological

observatories included in the study was observed. Similarly, analysis by Keim *et al*. (2004) suggests that although there was an overall decline in the frequency of extratropical cyclones, there was an increase in the frequency of very powerful storms in the North Atlantic Basin (NAB), especially at higher latitudes, over the past 50-100 years. More recent analysis by Matthews *et al.* (2014) shows that this trend has continued. In their assessment of multi-decadal variations in storminess, which considered cyclone frequency and intensity together, Matthews *et al.* (2014) showed that over the past 143 years, storm intensity had increased over the North Atlantic. In addition, during the winter 2013/2014, Ireland and the UK experienced the stormiest winter on record for at least 143 years. It is as yet unclear whether or not these shifts are related to climate change or natural variability.

Potential future changes in storminess under climate warming for the eastern North Atlantic have been examined by Lozano *et al.* (2004), McGrath *et al.* (2005), and others. Generally speaking, regional GCM results from these studies suggest Ireland can expect fewer, but more intense, extratropical cyclones by 2060. While the frequency of extratropical cyclones may decrease, results from McGrath *et al.* (2005) suggest that there may be an increase in the frequency of intense (core pressures <950 mb) cyclones, with the strongest increases likely to occur in winter and spring. In terms of storm surges, findings by Wang *et al.* (2008) suggest a significant increase in the frequency and height of winter storm surges across most of Ireland, except along the south Irish coast (not including Dingle Bay) for the period 2031-2060. It is relevant to note that inputs to these models are often derived from a limited number of meteorological stations, which may not fully represent the spatial variability in the wind climate across the island.

To conclude, as a result of climate warming, Ireland can expect an increase in the probability of occurrence of extreme events (*e.g.* those with core pressures <950 mb). Surges and extreme waves associated with such events can result in significant erosion of soft coasts. When such events become more frequent, there is less time for beaches and dunes to recover. An understanding of the relationship between storms and sand barriers is essential for effective

management. It is an aim of this study to improve our understanding of this relationship.

6 Terrestrial laser scanning: a new approach to monitoring beach morphodynamics

Quantitative information on coastal elevation is often an essential component of geomorphic inquiry. In many cases, this information is obtained using a theodolite, Electronic Distance Meter (EDM), precise Global Navigation Satellite Systems (GNSS), or other similar instruments. To obtain elevation data with these tools, usually single regularly or irregularly spaced elevation measurements are taken along a cross-shore or longshore transect. This approach to surveying is limited, though, in that in practice, only a small number of data points can be captured in the time it takes to complete a typical field survey. Advances associated with Light Detection and Ranging (LIDAR) technology in the late 20th century, however, have made it possible to collect many elevation measurements (on the order of millions) over a large area in a very short period of time. This technology has since been adopted by geomorphologists to map coastal features and has led to improved knowledge of coastal geomorphic processes (Brock and Purkis, 2009). Some examples are presented in section 6.1.

LiDAR is an active remote sensing technology that uses either a reflected laser pulse or, less commonly, differences in phase from a continuous beam, to measure the distance to a surface. Pulse-based sensors sweep millions of laser pulses across a surface and use the time it takes for those pulses to be reflected back to the instrument to measure the distance to the surface. A rotating optical mirror directs the pulses over the area to be surveyed at spatial intervals specified by the user.

There are two general types of LiDAR systems – airborne (figure 6.1) and ground based (figure 6.2). Airborne systems are flown on an aircraft, and thus are capable of capturing data over a relatively wide area. They consist of three main parts: the sensor, the inertial measurement unit (IMU), and the GNSS, which work together to produce georeferenced topographic data. Ground-based LiDAR systems, or Terrestrial Laser Scanners (TLS), capture data from one or more fixed positions on the ground. Georeferencing is usually established through the use of a known benchmark, although newer models may have a built-in GNSS and altimeter. The result of a LiDAR survey, airborne or ground-

based, is millions of densely-packed 3-D points, each with a unique xyz coordinate, collectively known as a point cloud. Figure 6.3 shows a visualisation of this data. Additionally, information on the intensity of each return is collected.

Although ground-based LiDAR systems are limited in terms of coverage area compared to airborne systems, the operational costs associated with these systems are much lower and the instruments can more easily be deployed on demand and at short notice. Ground-based laser scanners are therefore often better suited to the study of beach response resulting from short-term forcings (*e.g.* storms, aeolian processes, and seasonal changes in wave climate), particularly when data from multiple surveys are required. TLS systems are also advantageous over aerial systems in the study of micro-scale morphologies because they can capture higher resolution data, including of near vertical surfaces. Terrestrial laser scanners are capable of capturing point densities on the order of 3 orders of magnitude greater than airborne systems, with many instruments able to capture sub-centimetre resolutions.

In terms of field data collection, with ground-based surveying, usually multiple scans are obtained at different angles over a survey area to minimise shadowing. Shadow zones are regions of missing data located behind some obstruction to the scanner's field of view, such as low hummocks (*e.g.* figure 6.4). Point clouds from multiple scans from different positions over a surveyed area can be registered to a common coordinate system using survey targets (figure 6.5). They can also be georeferenced using the GNSS coordinates of targets.

TLS instruments are rapidly evolving and data capture is becoming ever faster. One of the first terrestrial laser scanners was introduced in 1998 by Leica Geosystems and could capture 100 points per second (Heritage and Large, 2009). The Faro Focus 3D, introduced in 2012, can capture almost 1,000,000 points per second. The scanners used in this PhD research, the Leica ScanStation and the Leica C10, are capable of capturing 4,000 and 50,000 points per second, respectively. Rates of data capture associated with some older models can limit scan resolution and coverage area due to practicalities associated with the time it takes to complete a survey in the field. For example, a three hour scan with the

C10 may produce tens to hundreds of millions more points than a three hour scan with the ScanStation.

While such large datasets may be useful for capturing a high degree of spatial heterogeneity, their size represents a challenge in terms of data management. Because the data can be collected relatively quickly, huge amounts of data can pile up quickly and storage may become an issue, especially when subsequent analysis requires manipulation of the data and multiple backups/versions of the data begin accumulating. In terms of data analysis, computer hardware and software sometimes cannot handle such large datasets, therefore they may need to be subdivided. For an excellent of these and other issues specific to the use of TLS in the environmental sciences, see Heritage and Large (2009).

The following section will present a selected subset of studies that illustrate practical applications of TLS surveying in the published coastal literature and evaluate the methodological contributions of these studies in terms of field data collection and post-processing of the data. Section 6.2 describes various techniques for filtering and classifying point clouds in natural environments and also provide an evaluation of the methodological contributions of this group of studies. One objective of this PhD is to assess the viability of TLS as a monitoring technique in vegetated coastal dune environments, and, as such, an evaluation of past work in this and similar contexts is relevant here.

6.1 TLS in coastal environments

Despite its advantages over aerial LiDAR, the use of TLS as a tool for the study of coastal morphodynamics is heavily underrepresented in the literature. Only a handful of studies in this area have been published in the last decade, many of which have been published only in the last few years. These studies both illustrate the practical applications of TLS in coastal environments, and elucidate instructive methodologies for collecting and analysing TLS data, with a particular focus on scan registration, georeferencing, the generation of digital elevation models (DEMs) and the quantification of morphological change. Unfortunately, there exists a lack of standardised approaches to these procedures in the published literature. As such, these methodological issues are examined

here, with a particular emphasis on vegetation filtration and TLS scene classification presented in section 6.2.

Terrestrial laser scanning has been successfully employed in the study of geomorphological processes occurring on sandy and gravel beaches, coastal dunes and hard rock sea cliffs. For example, Pietro *et al.* (2008) and Hoffmeister *et al.* (2012) have demonstrated the effectiveness of TLS as a tool for studying changes in beach volume on the east coast of the USA and in western Greece, respectively. Pietro *et al.* (2008) used TLS data to monitor beach nourishment performance and Hoffmeister *et al.* (2012) used TLS data to estimate beach volume changes in various littoral settings. These studies firstly demonstrate the relative efficiency of TLS field data collection – single surveys by Pietro *et al.* (2008) covered an area of 500 m x 70 m at 0.20 m resolution. Secondly, they demonstrate the ability to produce high resolution DEMs from TLS data, which can be used to precisely assess short-term elevation (*e.g.* figure 6.6) and volumetric change. Similar studies exemplify the usefulness of TLS in dune ecosystems. For example, Feagin *et al.* (2014) quantified changes in dune morphology after Hurricane Ike on the East Matagora Peninsula, Texas. Seasonal surveys of a 100 m x 100 m plot of beach were carried out at 1 cm resolution from September 2008 to October 2009. Volumetric change analyses from these surveys helped to quantify storm impacts and subsequent recovery.

Another group of studies explores the usefulness of TLS as a tool for the study of hard rock cliff erosion. Lim *et al.* (2005), Rosser *et al.* (2005), Poulton *et al.* (2006), and Olsen *et al.* (2009) used TLS to model changes in sea cliff morphology. These studies demonstrate the utility of TLS data over traditional cliff survey data (*e.g.* derived from aerial photos or historic maps) and even aerial-LiDAR data. Datasets derived from these sources are all limited in their ability to resolve vertical change due to the fact that the data are collected in plan-form. Because laser scanners can be orientated to collect data in front of the cliff face, cliff morphologies can be modelled more precisely. Rosser *et al.* (2005) showed that volumetric changes of <0.001 m^3 can be resolved using TLS survey data collected at 3 cm resolution. Over the past decade, terrestrial laser scanned datasets have led to significant improvements in our understanding of the activity patterns of coastal cliffs. TLS datasets have allowed researchers to

identify direct mechanisms of cliff failure that were previously unidentifiable using traditional cliff survey techniques (*e.g.* Lim *et al.*, 2005) and, for the first time, they helped demonstrate the ineffectiveness of these established techniques for cliff monitoring (*e.g.* Lim *et al.*, 2005; Rosser *et al.*, 2005), which has serious implications for current management practices.

Other studies involving the use of TLS in coastal environments have described innovative experiments in the study of aeolian transport – one, for example, relying on TLS to study deformation regimes during harsh weather conditions (Lindenbergh *et al.*, 2011) and another to study aeolian saltation clouds on a drying beach (Nield and Wiggs, 2011). Lindenbergh *et al.* (2011) erected a large screen orientated perpendicular to the wind on a beach at Vlugtenburg, the Netherlands and obtained multiple 1 mm resolution TLS surveys of the beach surrounding the screen at semi-regular intervals over a period of 88 hours. Despite harsh weather conditions throughout (with maximum gusts of up to 20 m/s), an analysis of distances between targets through time demonstrated that it was possible to identify local morphodynamic changes at the millimetre level. Lindenbergh *et al.* (2011) were able to link these sand volume changes (derived from the TLS data) to meteorological and saltiphone data obtained during the experiment and were able to use these data to identify different deformation regimes linked to wind speed and direction. This study demonstrates firstly the practicability of TLS for the study of aeolian deformation and secondly the ability of TLS sensors to operate successfully in harsh weather conditions. Nield and Wiggs (2011) demonstrated how TLS could be used to characterise aeolian saltation cloud dimensions. They collected multiple 5 mm resolution TLS scans over a drying beach at Ynyslas, Wales. Non-surface points were used to characterise saltation cloud intensity, which was then linked to changes in surface moisture and roughness. This study demonstrated another innovative use of TLS data in a coastal setting, taking advantage of data that would ordinarily be filtered out of the point cloud.

Finally, TLS has been used in the study of the evolution of distinctive coastal features, such as beach cusps (van Gaalen *et al.*, 2011), embryo dunes (Montreuil *et al.*, 2013a; Montreuil *et al.*, 2013b), aeolian sand strips/protodunes (Nield *et al.*, 2011) and an anthropogenic beach berm (Schubert *et al.*, 2014). In the past,

it would have been difficult, even impossible, to capture precise information about the dimensions of these relatively small features, irrespective of how they might change over time scales on the order of hours. These studies, though, highlight the value of TLS in this sense.

Within the studies described in this section, various methodological issues have been explored. These can be divided into matters related to:

1. Scan registration and georeferencing;
2. Filtering and classification;
3. Generation of DEMs; and
4. Quantification of morphological change.

These are discussed as follows, with filtering and classification techniques treated separately in section 6.2.

Scan Registration and Georeferencing

Scan registration is usually achieved using survey targets (figure 6.5), permanent markers (*e.g.* built structures), or differential or real-time kinematic (RTK) GNSSs. Deciding on which is most appropriate depends primarily on the survey accuracy required and whether or not multi-temporal registration is required. Survey targets can be used to register overlapping scans comprising a single survey. For surveys on sandy beaches ranging from 1-10 cm resolution, accuracies of registration when using survey targets are typically on the order of a few millimetres (*e.g.* Hoffmeister *et al.*, 2012; Montreuil *et al.*, 2013b; Schubert *et al.*, 2014). Survey targets, however, are inappropriate for registration of multi-temporal surveys due to their impermanence, therefore either permanent markers in the field or differential or RTK GNSS are used. van Gaalen *et al.* (2011) reported using a set of stairs to register multi-temporal scans and for 3.7 cm resolution scans, a registration error of 15 cm was reported. The use of fixed control points, however, is sometimes not possible or practical in dynamic environments, such as on beaches. Feagin *et al.* (2014) illustrate this problem in their study based on the Gulf coast of Texas, USA. As there was a lack of permanent structures on the beach they were working on, Feagin *et al.* (2014) installed semi-permanent reference stakes for subsequent registration. These, however, were eroded during Hurricane Ike. They, therefore, had to resort to

using control points they thought were in a similar position after the storm - the base of a fence post, a drift log, and a sign - for registration. They reported standard errors ranging from 6.4 cm to 22.9 cm for 1 cm resolution scans, which can be considered relatively poor compared to other studies like that of van Gaalen *et al.* (2011). To avoid this issue, Montreuil *et al.* (2013b) and Olsen *et al.* (2009) georeferenced each survey using RTK GPS (±10 mm horizontal and ±20 mm vertical accuracy) such that all surveys would be collected in the same coordinate system. For 9.4 cm resolution scans, Olsen *et al.* (2009) reported a Root Mean Square (RMS) error of registration between multi-temporal scans of 7.9 cm.

Generation of DEMs

TLS data often form the basis of surface models, such as raster-based DEMs (figure 6.7) or vector-based triangular irregular networks (TINs) (figure 6.8). DEMs can easily be generated using GIS or computer aided design (CAD) platforms and other stand-alone software packages. Considerations in the generation of DEMs include the optimal resolution of raster cells and the method of interpolation. Feagin *et al.* (2014) found that for 1 cm resolution scans in a coastal dune environment, the optimal raster cell size was 0.5 m x 0.5 m. With this cell size, 'bumps' related to shadow zones were minimised while the natural contours of the dune were preserved. This resolution, however, should only serve as a guide, given that other considerations, such as method of vegetation filtering, must be taken into account when deciding on an appropriate raster resolution.

There are various interpolation methods available for creating DEMs from point clouds, with no single one favoured in the literature for the generation of DEMs from TLS data. The most common methods include natural neighbour, inverse distance weighting (IDW), spline, and kriging. They all require some set of inputs (*e.g.* actual point measurements) and can each be summarised as follows:

Natural neighbour (see figure 6.9)

In the example shown in figure 6.9, the black dots represent the inputs and the red star represents an unknown value or "query point". With Natural Neighbour,

a Voronoi diagram is first constructed based on the inputs. A Voronoi diagram is a plane in which polygons are partitioned based on the distance between points in a subset of the plane (Aurenhammer, 1991). These polygons are called Voronoi (or thiessen) polygons and are represented by the green shapes in figure 6.9. A new Voronoi polygon is created around the query point (beige polygon in figure 6.9). The proportions of overlap between this new polygon and the original polygons are used as weights. It is from these weights that a value is interpolated for the query point. This procedure is repeated for each query point within the area of interest.

Because it adapts to the structure of the input data, natural neighbour often works best when there are large volumes of input points. This summary is based on an explanation in the ESRI ArcGIS 9.2 Desktop Help pages, available at http://webhelp.esri.com/arcgisdesktop/9.2/index.cfm?TopicName=Natural%20Neighbor%20Interpolation.

Inverse Distance Weighting (see figure 6.10)

Inverse distance weighting is based on the idea that input points closer to a query point are more likely to have similar values than input points further away. In the example shown in figure 6.10, the black and red dots are the inputs and the yellow dot represents the query point. The value for the query point is estimated by averaging the value of points in its neighbourhood (the area within the yellow circle), with nearer points given more weight and further points given less weight in this averaging process. This is different from natural neighbour in that it can take into account inputs that may extend beyond those around the same query point using natural neighbour. Again, the procedure is repeated for each query point within the area of interest.

Like natural neighbour, IDW is a useful technique when there is a relatively large volume of input points. Because the procedure can take into account many more inputs for each query point, IDW may require more processing time than the natural neighbour interpolation for the same size dataset. This summary is based on an explanation in the ESRI ArcGIS 9.2 Desktop Help pages, available at

http://webhelp.esri.com/arcgisdesktop/9.2/index.cfm?TopicName=Implementing_Inverse_Distance_Weighted_(IDW).

Spline

The spline interpolation estimates query points by fitting a minimum-curvature surface to the input data. The surface passes exactly through all input points. There are two types of spline – regularised and tension. Regularised spline produces a smooth surface, but estimated values may fall well outside the range of inputs. Tension spline produces a less smooth surface, but conforms more closely to the input values.

Spline can be useful when there is a relatively small volume of input points (*e.g.* when IDW and Natural Neighbour are unsuitable). This summary is based on an explanation in the ESRI ArcGIS 9.2 Desktop Help pages, available at http://webhelp.esri.com/arcgisdesktop/9.2/index.cfm?TopicName=How%20Spline%20works .

Kriging

Kriging is a probabilistic interpolation technique based on statistical models. It is more complicated to employ than the other techniques because multiple assumptions must be fulfilled before it can be rendered suitable and multiple iterations are usually required to obtain a satisfactory model. Kriging is based around the semivariogram – a statistical model of spatial variation within the dataset. This is first computed from a sample population from the input point dataset. The values of query points are then calculated based on this model. Because only a sample population of inputs was used, the semivariogram can be used to check the actual variance of the dataset versus the predicted variance of the dataset. It is therefore possible to quantify prediction errors, which is a particular strength of this method.

Before kriging can be employed, an exploratory spatial data analysis is required. Kriging is appropriate only for data that (1) follows a normal distribution; (2) is stationary (*e.g.* has constant variance); and (3) does not contain trends (systematic changes in the values of the data across the study area – *e.g.* as in a beach profile that extends from the lower intertidal zone to a dune crest). If all of

these requirements are fulfilled (or the data can be manipulated to such an extent that they become fulfilled), kriging may be employed.

If kriging is found to be unsuitable, it is usually possible to employ empirical Bayesian kriging (EBK), which is not as sensitive to the requirements outlined above. EBK works by dividing the data into subsets and creating multiple semivariograms, one for each subset. In this way, stationarity and trends are less of an issue. It also automatically fits the predicted and modelled semivariograms to produce optimal results. As a result, EBK is more efficient and more accurate than other kriging methods. Processing, however, is slower and customisation is limited. This summary is based on a talk by ESRI staff, which is available at http://video.esri.com/watch/1796/concepts-and-applications-of-kriging

Many studies fail to defend their method of interpolation choice with an error assessment. An exception is Montreuil *et al.* (2013b), who used inverse distance weighting (IDW) to interpolate TLS data from surveys of embryo dunes (1 cm resolution) in the UK. They compared z values from the original survey point clouds to those of the interpolated DEMs and reported RMS errors ranging from 0.057 to 0.068 m.

For modelling near vertical cliff faces, TINs can be advantageous over DEMs because they are better able to represent areas of variable complexity (*e.g.* not limited to a single resolution). For example, Rosser *et al.* (2005) used a 2 ½ dimension view dependent triangulation algorithm to generate a triangular mesh of a hard rock cliff face (figure 6.8). This method is unique in that it allows for an oblique, rather than aerial, viewing angle of the cliff face, highlighting changes in the morphology of the face of the cliff rather than simply recession of the cliff top. Development of such an algorithm, however, requires expert programming skills.

Quantification of Morphological Change

Finally, sequential DEMs derived from TLS data are being used to monitor coastal elevation and volume change. Elevation change maps can be produced by subtracting multi-temporal DEMs to produce DEMs of Difference (DODs). Figure 6.11 shows examples of DODs generated from TLS data by Montreuil *et*

al. (2013b) showing seasonal changes in the elevation of embryo dunes for three periods between October 2009 and October 2010. These can be easily created, for example, in ESRI ArcGIS using the raster calculator tool. From these DODs, elevation changes on the order of tens of cm are detectable.

Volume changes can also be calculated using GIS software. Montreuil *et al.* (2013b) calculated monthly rate of change in volume of sand (R_{vs}; in $m^3\ m^{-2}\ month^{-1}$) following the method of Young and Ashford (2006), where:

$$R_{vs} = \frac{V_s}{A \times T} \tag{9}$$

and

V_s = total volume of change (m^3)

A = area of the scan survey (m^2)

T = time between surveys (months)

The total volume of change and area of the scan survey were calculated from the DEMs using ESRI ArcView (now ArcGIS for Desktop Basic).

Olsen *et al.* (2012) have further automated these processes through the development of a user-friendly tool for comparing sequential DEMs and statistically analyzing alongshore-topographic change in ArcGIS. The tool is called TOPCAT. While it was developed with sea cliffs in mind, it can be used in the study of other elongated morphological features, such as coastal dunes, as was demonstrated in a case study at North Cape Hatteras National Seashore, North Carolina, USA (Olsen *et al.*, 2012). The tool requires two overlapping DEMs (before and after) as inputs. Alongshore compartments whose widths are specified by the user are then generated along the shoreline. For each compartment, TOPCAT calculates mean cliff face retreat and volumetric change per unit length per year by differencing the two input DEMs, such that alongshore changes can be evaluated. Figure 6.12 shows an example of TOPCAT outputs for a case study using DEMs generated from TLS data from Dog Beach, Del Mar, California. Average cliff face retreat rate (centre) and volumetric change (bottom) calculated for each compartment along the length of

the cliff (top) are shown. These data allow the end user to evaluate alongshore change, rather than simply interpret a DEM of difference or single net volume change calculation for the whole study area.

6.2 Filtering and classification techniques

An issue of particular importance in this research is filtering and classification of TLS point clouds. This is because the surveys to be undertaken are in a densely vegetated dune environment, and vegetation classification and subsequent filtration is essential for reasonably accurate DEMs to be generated from these data. Given that there are numerous approaches to filtering and classifying LiDAR point clouds in natural environments, this section examines a few of particular potential relevance to this research, first providing some background on feature extraction to put these into context.

The development of techniques for feature extraction from point clouds has been the subject of a wide body of literature over the last decade or so (Sithole and Vosselman, 2004; Brodu and Lague, 2012). Feature extraction is important for applications such as generating bare-Earth DEMs (Raber *et al.*, 2002), estimating biomass volume (Watt and Donoghue, 2005), and differentiating between natural and artificial objects (Straub *et al.*, 2009). Depending on the features of interest, automated classification and/or filtering[9] may be required. The choice of an appropriate filtering or classification technique is largely dependent upon the application, but may also be influenced by point density and whether or not information on multiple returns is available. For example, bare-Earth DEMs generated from aerial LiDAR data are often derived using information from the first and last return of single laser pulses. Many different algorithms that make use of this have been developed (for an excellent review, see Sithole and Vosselman, 2004). However, presently, most commercially available terrestrial laser scanners only record the first return signal, therefore this technique is not applicable for data obtained using these instruments. As such, the issue of generating bare-Earth DEMs from TLS point clouds has become a topic for

[9] For clarification, classification involves differentiating between different elements within a scene, and filtering involves the removal of unwanted points from a cloud, as in the generation of a bare-Earth digital elevation model (DEM).

discussion in the literature. Brodu and Lague (2012) describe four issues associated with classifying TLS point cloud data in natural environments:

1. The fully 3D nature of TLS point cloud data inhibits the use of pre-existing data analysis procedures developed for 2D data (such as raster data) or even 2.5D data (such as aerial LiDAR).
2. Shadow effects (*e.g.* as a result of obstructions to the scanner's field of view) and missing data (*e.g.* from standing water) result in datasets of variable resolution.
3. The complex geometries of natural surfaces are difficult to define.
4. Large, dense datasets, such as those associated with TLS data, establish the need for fast and precise automated processing procedures.

Working in the context of these issues, various authors have developed methods for filtering and classifying TLS point clouds. The most common methods rely on:

(1) laser scanned intensity;
(2) spectral reflectance signatures;
(3) elevation values within a specified grid cell; or
(4) the geometrical properties of points in the cloud.

Point cloud classification using laser-scanned intensity

It is sometimes possible to differentiate between classes within a point cloud on the basis of differences in reflected laser intensity. Intensity is the ratio of the strength of reflected light to that of emitted light and is proportional to the reflectance of the target at the wavelength of the incident laser. In TLS point clouds, more reflective surfaces (such as metal) generally have higher intensities than less reflective surfaces (such as concrete). Provided there is little overlap in reflectance characteristics between two objects, it may be possible to use laser-scanned intensity to differentiate between them. Franceschi *et al.* (2009), for example, were successful in using this technique to differentiate between clay mineral layers in a rock outcrop. Significant differences in the distributions of intensity values for limestones and marls meant that the two could easily be discriminated from one another. One problem with the use of laser-scanned

intensity to differentiate between classes is that intensity values vary depending on geometry of acquisition (distance from scanner and angle of incidence), surface moisture content, atmospheric conditions, and even the instrument used (Kaasalainen *et al.*, 2011). In some cases, it is possible to correct for some of these issues. Franceschi *et al.* (2009), for example, had to derive a correction for distance of acquisition and apply it to their data before discrimination between classes could be achieved. Derivations of such corrections can be complicated, especially for scenes that are spread over a large area and have complex geometries (Kaasalainen *et al.*, 2011). Also, if the intensity distributions of two classes have significant overlap, it may not be possible to successfully differentiate between the classes anyway. For these reasons, few studies have made use of laser-scanned intensity data as a discriminator.

Point cloud classification using spectral reflectance signatures

Some laser scanners are capable of capturing spectral information, which can be highly effective for classification purposes. Classification of data captured from terrestrial laser scanners with multispectral capabilities may be achieved on the basis of a feature's unique spectral reflectance signature. Unfortunately, most commercially available scanners don't have multispectral capabilities of high enough spectral resolution for the precise classification of complex scenes. Lichti (2005), however, showed that it is possible to perform relatively successful simple classifications using RGB values from an integrated digital camera, something many commercially available scanners *do* have. Using a thematic classification algorithm that exploits the four-channel (blue, green, red and near infrared) multispectral imaging capability of his TLSs in-built camera, Lichti (2005) attempted classification of two TLS datasets. The datasets included one from a vegetated bush land in Western Australia with three classes (sand, grass, and blue fringe vegetation) and one at a university campus with six classes (grass, trees, red brick, orange tile, concrete, and steel). Overall classification accuracies were high, with 87% achieved for the bush land and 82% for the university campus. Inaccuracies associated with this technique arise from the fact that RGB values are influenced by shadow projections, variations in image exposure, and surface humidity. Given that these vary with time, it is not a particularly useful technique for the classification of multi-temporal scans.

Non-ground point filtering using elevation values within specified grid cells

Sometimes, only the ground surface is of interest (*e.g.* for the generation of DEMs). In this case, the aim is to simply filter unwanted non-ground points (*e.g.* vegetation, buildings, etc.) so that only points that represent the true ground surface are left. A simple way to achieve this is to use lowest points analysis (LPA) – also sometimes called least return filtering. With this method, the scanned area is divided into a 2D grid. Within each grid cell, all points except those with the lowest z-value are filtered out, and the remaining points on the grid are meant to represent the bare-ground surface, *e.g.* the analysis assumes that for each grid cell, the point with the lowest z-value must represent the ground surface elevation. This, however, may not always be the case and is subject to both errors of omission (points excluded that *are* representative of the ground surface) and errors of commission (points included that *are not* representative of the ground surface). Depending on the application and the required resolution of the end-product, these errors may or may not be acceptable (Coveney and Fotheringham, 2011). Many software packages contain tools based on this technique, including Quick Terrain Modeler (the "Above Ground Level" tool) and Leica Cyclone II Topo (the "Find the Ground" tool) (Applied Imagery, 2009; Leica Geosystems, 2011). These tools usually give the user the option to specify or vary the ground finding parameters to meet specific requirements if necessary. For example, Cyclone II Topo allows users to specify largest or smallest vertical step (acceptable difference in height between neighbouring points), maximum slope, and grid cell size. While LPA may be useful for segregating buildings, cars, and trees from the ground surface in urban environments (Land Surveyors United, 2013), it has been found to be of limited use in areas where dense ground vegetation predominates. Coveney and Fotheringham (2011) performed an evaluation of the effectiveness of lowest points analysis in such an environment - an open, relatively flat, densely vegetated coastal saltmarsh on the west coast of Ireland. They found that for scans with an initial resolution of 6 cm and a filtering grid made up of 1 m cells, the 95% elevation error introduced as a result of applying a local lowest-point filter was 98 cm. This result was 16 times larger than the single largest error from the other error sources they looked at, which included registration to a

common coordinate system, GPS validation data error, GPS error, georeferencing error, and target position definition error. Also, a significant proportion of the cloud, 99.7%, was lost as a result of filtering at this resolution. Montreuil *et al.* (2013b), though, had some success with this technique for filtering vegetation from embryo dunes. For 1 cm resolution surveys, all but the lowest points within a 5 cm grid cell were removed. The results of a ground truthing analysis suggested that the overall magnitude of changes in dune height and volume exceeded errors associated with filtering and subsequent DEM generation. It should be noted that the success of the technique in this case was likely to be related to the fact that the vegetation cover was sparse. Others (Wester, 2011; Feagin *et al.*, 2012) have used this technique for filtering vegetation from coastal dunes, but a formal evaluation of errors introduced as a result of LPA filtering (*ie.* through ground truthing) was not reported in either of those studies.

Point cloud classification using the geometrical properties of points in the cloud

A final set of techniques for vegetation filtering and TLS point cloud classification is based on the geometric characteristics of lines or surfaces bounded by neighbouring points within the cloud. One example is that of Brodu and Lague (2012), who developed an innovative technique for classifying complex natural scenes using a multi-scale dimensionality criterion. This technique uses the 3D geometrical properties of scene elements across multiple scales to differentiate between them. The technique is based on the idea that at different scales, different elements within a 3D scene often have different dimensionalities. For example, in the context of a vegetated dune environment, at very small scales, vegetation may appear more one or two dimensional (*e.g.* as stems and leaves), but at a larger scale, it will start to appear more 3-dimensional (*e.g.* as a bush or tufts of grass). The ground surface may appear 3 dimensional at a very small scale (*e.g.* ripples in the sand), but 2-dimensional at a larger scale (*e.g.* a beach). By exploiting these differences in dimensionality at different scales, it is possible to build unique signatures for identifying different categories of objects or elements within a scene. The technique uses Principal Component Analysis (PCA), a statistical technique for finding patterns in data of high dimension, to characterise "local dimensionality" around each point at a given

scale. The user can then use training sets to decide which combination of scales should be used to maximise the separability between classes and build the classifier based on this information. Brodu and Lague (2012) developed an algorithm called CANUPO (CAractérisation de NUages de POints) for building classifiers and applying them to TLS point clouds. For the two test sites described in their paper, a tidal marsh and a steep, mountain river bed (figure 6.13), classification accuracies for separating vegetation from ground and other classes was reported to be greater than 98%. CANUPO is now built into the user-friendly, publicly available 3D point cloud and mesh processing software, CloudCompare (http://www.danielgm.net/cc/).

It is clear from the above discussions that TLS surveying has great potential in the study of coastal geomorphology. The studies described here provide justification for the further use of TLS to model change in coastal environments. Presently, however, the lack of standardized approaches to post-processing the data mean best-practice guidance related to the collection and analysis of TLS data in coastal environments is limited. This study aims to take a practical, and potentially operational, approach based on the guidance provided by the studies described in this section, specifically with regard to data collection and analysis. This is further addressed in chapter seven.

7 Observations of morphdynamic behaviour under the influence of storms

A major morphological monitoring campaign was undertaken as part of this PhD research from May 2012 to July 2014. The purpose of this campaign was to collect information about how the Inch and Rossbehy foredunes responded to storm events in an effort to evaluate the importance of the storm driver. The data was also used to evaluate how effective the numerical model described in chapter 10 was at simulating dune volume changes near the breach at Rossbehy.

Terrestrial laser scanning (TLS) was chosen as the basis for data collection. This technique was chosen for a number of reasons. Firstly, with TLS, topographic data can be captured on demand. This is a requirement of this research in that to understand the response of the dunes to storms, surveys must be completed in the direct aftermath of storms. Similar technology, such as aerial LiDAR, was deemed impractical for this research in that surveys can rarely be completed on demand, and even to perform a single survey would have been out of the budget of this study. Second, TLS technology is better able to capture the spatial heterogeneity of surveyed topography than more widely used techniques, such as dGPS or electronic distance meter (EDM) surveying because millions of measurements can be captured relatively quickly (as opposed to tens or hundreds with dGPS or EDM surveying). This means (sediment) volume changes can be calculated with more precision, and smaller changes in volume are detectable. As prior to the start of the monitoring campaign it was unclear how much volume change would be observed, it was reasoned that the use of TLS would allow for the best chance of detecting volume change. Finally, the technology was freely available for this research and kindly provided by the Geography Department at UCC.

Various issues related to the use of TLS in this research had to be addressed, including registration of multi-temporal scans, vegetation filtration, generation of DEMs, and quantification of topographic and volumetric change. This chapter describes in detail the methods used to (1) collect and analyse the TLS data, (2) identify storm events and characteristics and (3) search for relationships between

observed morphologic change and storms. The results of these analyses are presented in this chapter.

Two field sites were selected for analysis (figure 7.1). These each covered an area of approximately 100 m x 50 m. The Inch site lies adjacent to the main inlet between the two barriers. Scans here covered a large section of the face of a high foredune (>20 m ODM), an extensive ephemeral embryo dune field and part of the upper beach (figures 7.2-7.3). Figure 7.4 shows an aerial view of the site. This site was chosen for its relative remoteness (lack of human interference) and proximity to the main inlet, where current speeds are high and capable of transporting eroded material between the ebb delta and inner bay/estuary.

The Rossbehy site is located at the terminus of the section of the Rossbehy barrier still attached to the mainland, adjacent to the new tidal inlet formed in 2008 (figure 7.5). Scans here mainly covered the upper beach and foredune scarp on the seaward side of the barrier (figure 7.6), although some scans covered part of the vegetated dune field behind the scarp and the upper back barrier beach fronting these dunes (figure 7.7).

The locations of the two field sites, at the distal ends of the dune barriers, are important in the context of this study because previous research (*e.g.* Sala, 2010 and Vial, 2008) suggests that this is where sediment fluxes are appreciable over timescales on par with this research.

7.1 TLS data collection

Topographic field surveys at each site were completed using a Leica ScanStation (Leica Geosystems AG, Heerbrugg, St. Gallen, Switzerland, Scan Station), a pulsed laser scanner with a positional accuracy of ±6 mm up to a range of 50 m (Leica Geosystems, 2006). Features of the ScanStation include a pulsed proprietary microchip laser (Class 3R, IEC 60825-1), an optomechanical mirror system with a full 360° horizontal and 270° vertical field of view, and an integrated high-resolution digital camera (Leica Geosystems, 2006). Figure 7.8 shows a diagram of the individual system components. The setup primarily consists of the scanner itself, a tripod, a battery pack, and a laptop from which the instrument is operated. During the course of the monitoring period, problems

related to battery performance necessitated the use of a generator, rather than a battery pack, to power the instrument. A Leica C10, a newer and more efficient model than the ScanStation[10], was used for some of the surveys due to technical difficulties with the ScanStation. Figure 7.9 shows the ScanStation setup at the Rossbehy field site.

As neither field site was accessible by car, alternative arrangements for transporting the equipment to and from each site had to be made. At Rossbehy, field equipment was transported on a trolley by foot from the nearby car park to the field site, a distance of approximately 2.5 km. At Inch, this distance was considerably further – approx. 5.5 km – so arrangements were made with local farmers to tow the equipment using a quad bike or tractor (figure 7.10). Both sites were often inaccessible at and around high tide (MHWS=+3.76 m ODM); therefore careful planning (and common sense) was required to ensure safe passage on entry and return. Sometimes it was necessary to compromise on the planned coverage area, number of scans or scan resolution to reduce scan time in the field, particularly on days when there was inclement weather and/or any concern that it would be possible to return safely.

Field surveys were typically carried out as follows. On arrival, the general location of scan stations[11] would be decided upon. Approximately 15 minutes was required for instrument set up (levelling, powering on, setting up laptop) at each scan station. Leica high-definition surveying (HDS) registration targets (figure 7.11) were set up strategically such that they would each be visible from each station for subsequent registration.

The laser scanner itself was controlled from a laptop using Leica Cyclone v. 8.1 (Leica Geosystems, 2013), Leica's 3D point cloud processing software. In Cyclone, the scanner would first be directed to take photos in 360° space around the scanner to aid in selecting the area to be scanned. From the resulting photo mosaic, the three HDS targets, along with semi-permanent targets for multi-temporal scan registration, could be identified (figure 7.12). Once visually

[10] The C10 also has a positional accuracy of ±6 mm up to a range of 50 m, but can capture data up to 12.5 times faster than the ScanStation.
[11] Scan station = location of instrument set-up – often multiple scans from different positions would be obtained to increase coverage area and decrease shadowing in zones of overlap.

identified, the registration markers are 'fenced.' This means their locations in 360° space are specified to the software by drawing a perimeter around each one in the photo mosaic. The targets were then scanned in high resolution (1 mm), providing precisely located markers required for later scan registration. This was always performed first, because if for any reason the main survey was interrupted (*e.g.* if it started to rain heavily), the data collected up to that point would contain information about the location of these markers relative to the main scan. Once this was complete, the area to be scanned was fenced. Once the resolution and range were specified, the scanner could be directed to begin scanning. A typical scan for a 180° scene with 2.5 cm resolution over a 30 m range would take approximately 1.5 hours with the Leica ScanStation. Most scans were obtained at resolutions of 1-3 cm, although the December 2013 and January 2014 surveys at Rossbehy were obtained at resolutions of 10 and 15 cm, respectively, to minimise time spent in the field in poor weather conditions. The desired optimal scan resolution of 1 cm was based on previous studies that used TLS to study seasonal variations in dune morphology (*e.g.* Montreuil *et al*., 2013 and Feagin *et al*., 2014). In practice, it was difficult to obtain 1 cm resolution scans for this study due to limitations associated with the efficiency of the instrumentation and tidal and weather conditions. In terms of time intervals between surveys, an approximately 1-2 month interval was desirable such that morphological changes could be attributed to particular events. Again, in practice, this was difficult to accomplish.

Overall, 22 surveys were completed in total – 9 at Inch and 13 at Rossbehy. Tables 7.1 and 7.2 summarise information about the individual field visits for which surveys were successfully completed for Rossbehy and Inch, respectively. The August 2013 survey at Rossbehy covered only a small part of the upper beach that did not sufficiently overlap with other surveys and was therefore not used in any subsequent analysis.

The data captured by the scanner for each point included the x, y, and z coordinates, laser scanned intensity values, and red, green and blue (RGB) values (obtained from photographs taken with the inbuilt digital camera). The x, y, and z coordinates were collected relative to the position of the scanner in an arbitrary grid reference system. In this system, the scanner is located at the (0,0,0)

coordinate. Each point in the cloud is represented by its distance in the x, y, and z directions (in metres) from the scanner. Laser scanned intensity is also supplied by the scanner and represented in Cyclone on a scale ranging from -2048 to + 2048. Because each individual scanner can have different intensity characteristics (minimum, maximum, response curve, etc.), the values supplied by the scanner to Cyclone are scaled to this range. RGB values are given from 0-255 as per the RGB colour model. The data are stored in the Cyclone database as project (.imp) files and can be exported in a variety of formats, including as ASCII text.

7.2 Post-processing

7.2.1 Scan Registration

The reference datum for each TLS point cloud is the position of the scanner; therefore all clouds are collected in arbitrary coordinate systems and must be registered to one another. Two general types of scan registration were performed for this research – same-date scan registration and different date scan registration. Scans obtained from two or more stations on same-date field surveys were each registered to a single arbitrary common coordinate system using the Leica HDS registration targets. This process is outlined as follows and illustrated in figure 7.13, with the Inch field site shown as an example.

A minimum of three Leica HDS targets was set up in the field such that they were visible to the scanner from all stations. The position of the targets is important, as the RMS error associated with registration can only be guaranteed for the area enclosed by the targets, so the wider the area enclosed by the targets, the better. The targets were manually defined in Cyclone in the field (*e.g.* named t1, t2, t3) and a fine scan (1 mm) of each target was completed at each station. Corresponding target names were used to register the cloud obtained at station 2 to the cloud obtained at station one. The resulting registered (or "unified") cloud is therefore in the coordinate system of the cloud obtained at station 1. This was performed for each of the same-date scans in Cyclone.

Cyclone provides a report on the overall accuracy of the registration. This includes the error in the x, y, and z directions and the Root Mean Square (RMS) error for each target constraint. RMS errors between same-date scans were

typically <1 cm. Histograms showing the distributions of registration errors for all of the same-date scans are shown in figures 7.14 for Inch and 7.15 for Rossbehy. The mean RMS error associated with same-date registration for Inch was 0.004 m and for Rossbehy 0.005 m. All registration errors at Inch were <0.01 m, while at Rossbehy, five coordinates (12% of the total population) had registration errors >0.01 m. The maximum error of registration was 0.025 m.

Different-date scans also had to be registered to one another such that volume change between them could be compared. As this research is concerned with small-scale morphologic change, it was preferable to identify a way of registering scans that would allow for maximum multi-temporal registration accuracy. The use of differential GPS was initially deemed inappropriate, as accuracies on the order of mm to cm could not be guaranteed - the accuracy associated with the instrument available, a Trimble ProXH, was typically about 20 cm (Trimble, 2005). Stationary, semi-permanent markers installed in the field, however, can act as registration markers that would allow for such accuracies to be achieved (van Gaalen *et al*., 2011; Feagin *et al*., 2013). Initially, five wooden posts with steel nail heads protruding approximately 5 cm out of the tops of the posts were installed at each field site (figure 7.16).

The nail heads acted as registration markers and were easily identifiable within the scans. While these registration markers stayed in place for the duration of the study at the Inch field site, they were removed on three occasions from the Rossbehy field site due to rapid erosion and possible theft, and thus new posts had to be installed[12]. Because the GPS coordinates of the posts were taken each time with the dGPS as a precautionary measure, scans lacking common coordinates from semi-permanent targets could be compared using Irish National Grid, the coordinate system in which the dGPS coordinates were collected in. In future cases, however, the Irish Transverse Mercator coordinate system should be used to collect GPS data in Ireland, as it was specifically introduced to be GPS compatible.

[12] The first set of posts disappeared between the November 2012 and January 2013 surveys, the second between February 2013 and April 2013, and the third between June 2013 and August 2013.

Considering the fact that the Rossbehy field site was eroding rapidly throughout the study (for example, up to 54 m of dune toe recession was observed between December 2013 and January 2014), the relatively small errors associated with registration using the dGPS coordinates (on the order of 10s of cm) were no longer deemed a major issue. The November 2012 to January 2013, February 2013 to April 2013, June 2013 to December 2013, and all subsequent scans were registered in this way.

To register the scans in Cyclone using the semi-permanent targets, the targets were identified and a point common to both scans obtained at t_1 and t_2 (*e.g.* on the nail head) was selected and tagged as a "constraint." Each constraint was given a unique name, which was shared in the t_1 and t_2 scans. For example, figure 4.16 shows an example of one of these constraints on a semi-permanent target (called "Post 3") used to register the May 2012 and August 2012 scans at Inch. A minimum of three constraints is required for registration. For the three constraints from the semi-permanent wooden posts used to register the May 2012 to August 2012 scans at Inch, RMS error was between 1 and 3 mm (figure 7.17). Histograms showing the distributions of registration errors for all of the scans registered using the semi-permanent targets are shown in figures 7.18 for Inch and 7.19 for Rossbehy. The mean RMS errors for scans registered in this way were 0.003 m for Inch and 0.012 m for Rossbehy. All registration errors at Inch were <0.01 m, while at Rossbehy, 6 coordinates (40%) had registration errors >0.01 m. The maximum error of registration was 0.045 m. The larger registration errors associated with the Rossbehy data may be due to small changes in the position of the posts in the field. This was an initial concern, considering the dynamic nature of the dunes in which they were placed. For this reason, dGPS coordinates of each of the posts were obtained as a backup. Even so, mean registration errors were lower for those scans registered using the semi-permanent targets than those using the dGPS coordinates (discussed in the following paragraph).

For the scans that did not share common registration markers from the semi-permanent targets, scans were registered using the dGPS coordinates of markers scanned in the field (either HDS targets or the semi-permanent markers). The dGPS coordinates of each of these markers were obtained on-site in Irish

National Grid using a Trimble Pro-XH differential GPS. These data were downloaded and imported into Cyclone. The corresponding target locations were identified in the scanned data and then used to register the cloud to the imported dGPS data. As none of the semi-permanent markers were removed for the duration of the study at Inch, only scans at Rossbehy had to be registered in this way. A histogram showing the distribution of registration errors for all of the scans registered using the dGPS coordinates is shown in figure 7.20. The mean RMS errors for scans registered in this way was 0.109 m, considerably higher than for those registered using semi-permanent targets (0.012 m). This is likely to be due to the limited positional accuracy of the dGPS or the use of the Irish National Grid coordinate system rather than ITM. While the maximum registration error associated with this type of registration was 0.546 m, most errors (29/32 or 91%) were under 0.2 m.

7.2.2 Vegetation filtration

The dune surface at both field sites was obscured by dense marram (*Ammophila arenaria*) cover. Given the height of the vegetation (approx. 0.3-0.5 m) relative to the resolution of the scans (0.01 to 0.1 m), DEMs generated from raw point cloud data would not be representative of the true ground surface. As such, the vegetation was filtered from the scans. While it is acknowledged that vegetation plays a key role in dune morphodynamics, its role in the morphodynamic evolution of the dunes at the field sites was not examined, as this was not a key objective of this study.

Given the number of scans obtained during this research, an efficient automated approach to vegetation filtration was most desirable. Various approaches were considered and tested on subsets of data. These included (1) lowest points analysis; (2) the use of reflected laser intensity distributions; and (3) the use of the geometrical properties of points in the cloud to differentiate between the ground and vegetation. Ultimately, the latter was chosen, but a short explanation of why the first two were deemed inappropriate is presented as follows.

Lowest Points Analysis

Lowest points analysis is a commonly used technique for separating ground and non-ground points within LiDAR point clouds. In their study on changes in sediment and vegetation volumes after Hurricane Ike, Feagin *et al*. (2012) used this analysis to separate vegetation from the ground surface (see chapter 6). With this method, the scanned surface is divided into a grid. Within each grid cell, all points but that with the lowest z-value are removed, and the remaining points on the grid are interpolated. Essentially, the analysis assumes that the point with the lowest z-value must be the ground. This, however, may not always be the case and is subject to both errors of omission (points excluded that are representative of the ground surface) and errors of commission (points included that are not representative of the ground surface). To minimise such errors, it is important to choose a suitable grid size, *e.g.* one that is sufficiently large that it can reasonably be assumed that a ground point has been included, but one that is small enough to capture the spatial heterogeneity of the bare surface.

This technique was tested on a small patch data using Cyclone II Topo's "Find the Ground" tool (Leica Geosystems, 2011). The data was obtained at the Inch field site at 1 cm resolution in May 2012. Five grid cell sizes were considered: 5 cm, 10 cm, 20 cm, 50 cm, and 1 m. Based on visual inspection of the data, the 10 cm grid spacing appeared to yield the best results. The smaller grid size resulted in poor classification, while the larger grid sizes meant the cloud density was considerably reduced. Even with the 10 cm grid spacing, 90% of the points in the cloud were lost. Figure 7.21 shows the cloud before (top) and after (bottom) filtering using the 10 cm grid spacing, with a cross section through the centre of the cloud shown right. While much of the vegetation appeared to have been removed, the results of this test reveal the technique is not robust for steep slopes. This is to be expected, given that more points would be likely to be filtered in a 10 cm x 10 cm grid cell on a slope than on a flat surface. Dune scarps are prevalent features at both sites, particularly at Rossbehy, therefore this method was regarded as not particularly suitable for vegetation filtration at these locations.

Vegetation Filtering using Reflected Laser Intensity

In some cases, it may be possible to differentiate between land cover types (including bare ground and vegetation) using reflected laser intensity (Franceschi *et al.*, 2009; Guarnieri *et al.*, 2009). Intensity values represent the ratio of the strength of reflected light to that of emitted light and are proportional to the reflectance of the target at the specific wavelength of the incident laser. Different materials return different intensity values, and thus intensity may be used to differentiate between land cover types over a scanned area (Chust *et al.*, 2008). There are, however, many factors which influence intensity, including surface roughness, geometry of acquisition (*e.g.* distance from scanner, angle of incidence), surface moisture content, atmospheric dispersion, and even the instrument used (Lichti and Harvey, 2002; Jensen, 2009; Kaasalainen *et al.*, 2011). For the ScanStation used in this study, Cyclone scales intensity values to a range of -2048 to + 2048. The values are unique to the laser scanner used and based on its minimum and maximum capabilities.

Some studies have attempted to quantify the importance some of the factors that affect laser scanned intensity. For example, Pesci and Teza (2008) set up an experiment to examine the role of surface roughness on intensity data. They found that (1) on a flat surface, intensity varies with angle of incidence, and (2) on an irregular surface, intensity remains almost constant, regardless of variations in angle of incidence. This implies that the technique may be more useful in geological/geophysical surveying (as opposed to architectural/cultural heritage applications) because the surfaces under investigation are usually irregular. Earlier work by Bellian *et al*. (2005) supports this assertion, as reflected laser intensity was successfully used to classify geological features on a canyon wall.

While the use of intensity values to classify complex scenes (*e.g.* those with several different land cover types / classes) is generally regarded as unsuitable, it may be possible to use this technique in vegetated dune environments, where there are usually only two land cover types: bare sand and vegetation. It was hypothesized that if the two return different intensities, a bimodal intensity distribution would be expected, with one peak representing intensity values

associated with vegetation and the other with the bare ground surface. On initial inspection of the data from the same patch on which lowest points analysis was tested (from May 2012 at Inch; shown in figure 7.22), a somewhat bimodal intensity distribution was observed (figure 7.23). The light blue peak appeared to correspond to primarily ground points, while the green peak appeared to represent patches of vegetation. Points on the lower tail of the blue peak, however, also appeared to represent vegetation. When points with intensities outside those associated with the main blue peak (from values of -233 to -156) were filtered, the result was an extremely poor classification (figure 7.24). Various ranges were tested, with none generating satisfactory output.

In a second test, it was attempted to identify whether or not particular groups of intensity values were associated with each class. Four-hundred sample points, consisting of 200 points belonging to the ground class and 200 points belonging to the vegetation class, were manually selected in the cloud and their intensity values were recorded. Figure 7.25 shows a histogram illustrating the distribution of intensities associated with each of these classes for the 400 sample points. For the most part, the two distributions were found to overlap. On further investigation, it was found that some points from each class shared exactly the same intensity values, which meant it would not be possible to use intensity values in their raw form to differentiate between the two classes.

It's known that laser scanned intensity is a function of geometry of acquisition, primarily distance from the scanner and angle of incidence, and it's possible that this is a cause of the observed overlap between the intensity values of the two classes. Some studies (*e.g.* Pesci and Teza, 2008; Kaasalainen *et al.*, 2011) have demonstrated that it may be possible to work out an empirical correction for site-specific datasets. Such a correction, when applied to the entire dataset, may assist with the differentiation between classes. To see if a correction could be determined for the Inch subset to correct for distance, mean intensity values were plotted for the sample points within the ground (figure 7.26) and vegetation (figure 6.27) classes at intervals of 5 m from the scanner up to a distance of 50 m. It was found that the intensity values of ground and vegetation points vary with distance, but not in the same way, therefore different empirical corrections would be required for each class. This meant that for this dataset, it would not be

possible to apply a single correction for distance to all points in the cloud for the purpose of discrimination between classes. Given the difficulties associated with using laser scanned intensity for this purpose and the significant amount of time devoted to unsuccessfully attempting to identify and correct for these issues, this approach was abandoned in favour of a more promising one described in a paper by Brodu and Lague (2012).

Vegetation Filtering using CAractérisation de NUages de POints (CANUPO)

Brodu and Lague (2012) developed a technique for classifying TLS point clouds in complex natural environments. This technique uses the 3D geometrical properties of scene elements across multiple scales to differentiate between them. The technique is based on the idea that at different scales, different elements within a 3D scene often have different dimensionalities. Dimensionality is defined conceptually by Brodu and Lague (2012, p. 123) as "how the cloud geometrically looks like at a given location and a given scale: whether it is more like a line (1D), a plane surface (2D), or whether points are distributed in the whole volume around the considered location (3D)." The quantitative measure of dimensionality for each point is defined by the eigenvalues resulting from a principle component analysis (PCA) on the points.

To give an example in the context of a vegetated dune environment, at very small scales, vegetation may appear more one or two dimensional (*e.g.* as stems and leaves), but at a larger scale, it will start to appear more 3-dimensional (*e.g.* as a bush or tufts of grass). On the other hand, the ground surface may be more 3-dimensional at a very small scale (*e.g.* ripples in the sand), but more 2-dimensional at a larger scale (*e.g.* a beach). By exploiting these differences in dimensionality at different scales, it is possible to build unique signatures for identifying different categories of objects or elements within a scene.

Brodu and Lague (2012) developed this idea in the form of an algorithm called CAractérisation de NUages de POints (CANUPO). CANUPO can be used to build site-specific classifiers, which can then be used to classify TLS point clouds. For this PhD research, classifiers were built using the built-in CANUPO plugin for CloudCompare, an open source 3D point cloud and mesh processing software freely available from http://www.danielgm.net/cc/. Figure 7.28 shows

the workflow used in this research for classifier construction, which is described as follows.

The first step was to prepare training sets, or "examples", of each category – in this case, vegetation and ground (examples from the May 2012 test dataset at Inch are shown under "Step 1" in fig. 7.28). Training sets should be as representative as possible of each class and can include as many samples as necessary. Next, a relevant set of scale intervals must be specified for which dimensionality between classes sufficiently differs. Local dimensionality is quantiatively defined in CANUPO using Principal Component Analysis (PCA), a statistical technique for finding patterns in data of high dimension. Initially, a "best-guess" based on knowledge of the scene elements can be performed to aid in decision-making with regard to the identification of appropriate scale intervals. Further refinement can be achieved based upon visual analysis of density plots, triangular plots which aid in visualisation of the dimensionality, at each scale. Each corner of a density plot represents the tendency of the cloud to be 1D (lower left), 2D (lower right), or 3D (top). The plots are generated from the eigenvalue ratios calculated during PCA. Density plots for four scales are shown for the vegetation and ground classes of the Inch test dataset in fig. 7.28 (under "Step 2"). At all of these scales, the ground surface remains mostly 2-dimensional, while the vegetation tends to become more three dimensional as the scales increase.

Based upon the dimensionality of the training sets at the specified scales, the algorithm then generates a probabilistic classifier by projecting the data in a plane of maximum separability between classes and then separating the classes in the plane. Fig. 7.28 (under "Step 3") shows an example of a proposed classifier. Classified points lie in the multiscale featurespace (red / blue) and the decision boundary (line separating the two) is generated using Linear Discriminant Analysis (Theodoridis and Koutroumbas, 2008). The classifier can also be generated using Support Vector Machines, although both produce almost identical results (Brodu and Lague, 2012). It is also possible to manually shift or tune the position of the decision boundary. Once the classifier is validated ("Step 4") it can be applied to the entire scene ("Step 5"). The classes can be viewed, separated, and exported in CloudCompare.

The process of generating a successful classifier is often an iterative one. A good way to improve classifier performance is to identify false positives (*e.g.* ground classified as vegetation or vice-versa), include these as new training sets, and build the classifier again (Brodu, *pers comm.*). In this research, the choice of best classifier was determined by testing whether or not the difference in residual error (see section 7.3.1) between the cloud filtered using the improved classifier was statistically significantly different from the cloud filtered using the previous classifier. As the error distributions for these clouds were not normal, Mann-Whitney U tests were performed from one iteration to the next until the differences were no longer significant[13].

Several classifiers were built for Inch and Rossbehy. At Rossbehy, separate classifiers were built to deal with vegetation on the scarp (*e.g.* exposed roots) and on the beach (*e.g.* slump blocks and pioneer plants). For the scarp, scale intervals of 0.1 m from 0.7 m to 1.5 m were found to result in the best classifier performance. For the beach, scale intervals of 0.5 m from 0.5 m to 2.5 m were found to result in the best performance. For the lower resolution scans from December 2013 (10 cm) and January 2014 (15 cm), scale intervals of 1 m from 5 m to 20 m were found to result in best classifier performance for the beach. The classifier used on the scarps of the higher resolution scans performed well on these lower resolution scans so no new classifier was built for these. At Inch, scale intervals of 0.1 m from 0.1 m to 1 m resulted in the best classification.

Prior to applying these classifications to and filtering each dataset, the raw scanned data had to be prepared. The first step was to remove erroneous data from the raw point clouds. Erroneous data can be generated from people or animals walking in front of the scanner, suspended sand or dust particles, or interference with direct sunlight. Erroneous data removal is fairly straightforward and can be done in either Cyclone (Leica Geosystems, 2013) or CloudCompare (danielgm, 2013). Such points can either be selected individually or in groups, at which point they can simply be deleted manually. Zones of poor or irregular resolution, usually at the far edges of the scans, and registration markers were

[13] This could only be performed on the datasets for which ground truthing data (from an EDM survey, presented in section 7.3.1) were available. These datasets were the 27 February 2013 data for Inch and 28 February 2013 data for Rossbehy.

also removed. At Rossbehy, scanned data were divided into beach and scarp for separate analyses. Also at Rossbehy, due to the fact that two different types of coordinate systems were used (the arbitrary ones and Irish National Grid), some datasets had to be prepared twice – once in the original coordinate system (for, for example, comparing to the previous survey) and once in Irish National Grid (for example, for comparing to the subsequent survey). This was because the semi-permanent posts had been eroded or removed on multiple occasions, so registration had to be performed using the Irish National Grid coordinate system (*e.g.* the only one that was common to all clouds). All analyses were not performed using Irish National Grid because clouds registered to Irish National Grid were associated with higher registration errors. Therefore, where possible, clouds registered to one another using the semi-permanent targets were used.

Once the data were cleaned up, the classifiers were run on each dataset, the vegetation was removed, and the bare-ground points were saved as ASCII text files for subsequent analysis.

Many surveys of the foredune at Rossbehy did not overlap in plan form. This was because the foredune had receded considerable distances landward on multiple occasions. An example is illustrated in figure 7.29. The November 2012 cloud is shown in the foreground and the January 2013 cloud in the background. The barrier terminus had receded landward by 44 m (the distance represented by the red line). Because of this, volumetric change analysis was performed in the horizontal rather than the vertical dimension (see also section 7.3). To perform such an analysis in ArcGIS, the coordinate systems of scans of the scarp at Rossbehy had to be translated in a subsequent step in CloudCompare. This process is illustrated in figure 7.30. Two foredune point clouds from Rossbehy are shown, one captured at time t_1 (red) and another captured at a later date, t_2 (blue). These are shown in plan view – e.g. looking down from above (top). Using CloudCompare software, it is possible to rotate the clouds along a rotation axis using the rotate/translate tool. An oblique view of the clouds captured as they were being rotated is shown (middle). The clouds were rotated 90 degrees about this axis, such that their final orientation was as shown (bottom). As a result of this translation, when the data are imported into ArcGIS, foredune elevation is represented along the y-axis.

Figures 7.31 and 7.32 show typical examples of classified and filtered clouds at Inch (figure 7.31) and Rossbehy (figure 7.32). An assessment of error attributable to vegetation filtration was carried out using the February 2013 survey datasets, as corresponding ground-truthing data were obtained at that time. That assessment is described and results are presented in section 7.3.1.

7.2.2 Generation of DEMs

Point clouds processed in CloudCompare were exported as ASCII text files and imported into ArcGIS v. 10.2 as xy data. The xy data were then exported to point shapefiles (along with corresponding z values) for further analysis. Digital elevation models (DEMs) were generated from the point shapefiles using inverse distance weighting (IDW) with a variable search radius set to include 12 points (default). This method was chosen based on an evaluation of its performance against the empirical Bayesian kriging (EBK) and natural neighbour (NN) interpolations. Raster DEMs were exported at a resolution of 0.1 m for elevation and volumetric change analysis.

7.3 Chronotopographic and volumetric change analysis of TLS data

Chronotopographic (elevation change) and volumetric change analysis were also performed in ArcGIS. While a standard method of subtracting elevation (z) values at T_1 from T_2 could be applied to the Inch data (shown diagrammatically in figure 7.33), this was problematic at Rossbehy, where the scarp had shifted landward so much during the survey period that few scans actually overlapped in plan view. For example, during the short time between November 2012 and January 2013, the scarp had receded by more than 40 m. To address this issue, change analysis on the scarp at Rossbehy was performed in the horizontal, facing the scarp (shown diagrammatically in figure 7.34). The coordinate systems of the scarp clouds were translated in CloudCompare prior to the generation of DEMs so that chronotopographic and volumetric change analysis could be performed in ArcGIS. In some cases, parts of the beach at Rossbehy did overlap in plan form. In these cases, chronotopographic and volumetric change analysis was performed in the same way as on the Inch data.

The GIS workflow for generating elevation/distance change maps and calculating volumetric change between data from T_1 to T_2 is illustrated in a cartographic model (figure 7.35) and outlined as follows:

1. Use minimum bounding geometry tool (set to convex hull) to create a perimeter around each raw TLS point shapefile. Edit as necessary to ensure that no gaps are present.
2. Use the intersect tool to create a new polygon, inside which DEMs from T_1 and T_2 will overlap.
3. Use the raster clip tool on exported raster DEMs with the polygon created in step 2.
4. Use the raster calculator to subtract z values at T_1 from z values at T_2.
5. Format layout and export as cliff face / elevation change map.
6. Use the cut and fill tool to extract information about volumetric change (net gains and losses).

Rates of volume change were calculated for each period using the formula of Young and Ashford (2006):

$$R_{vs} = \frac{V_s}{A \times T} \tag{10}$$

where:

R_{vs} = rate of volumetric change (m^3 per m^2 per day)

V_s = volume change (m^3)

A = areal extent of analysis (m^2)

T = time between surveys (days)

The outputs of this analysis included elevation / scarp distance change maps; maps showing the location of net volumetric gains and losses; and information about mean elevation / scarp distance change, net volumetric gains and losses, and rates of volumetric change between surveys. The results of this analysis are presented in sections 7.4 and 7.5.

7.3.1 Error quantification and propagation

In February 2013, a ground-truthing initiative was undertaken in an effort to quantify error and uncertainty associated with vegetation filtration and DEM generation. Error, in this context, is defined as the difference between the measured elevation (z) values (*e.g.* elevations derived from TLS data) and the true elevation values (*e.g.* elevations obtained from EDM data). Uncertainty is the quantification of the doubt that exists about the elevations for the DEMs where true values don't exist. The aims of this assessment were to find out:

1. Are errors ($z_{TLS} - z_{EDM}$[14]) significantly lower for filtered vs. unfiltered clouds?
2. Which interpolation technique results in lowest errors?
3. Are errors associated with filtering *and* DEM generation significantly lower for filtered vs. unfiltered clouds?
4. What is the minimum level of change detection possible for the Inch and Rossbehy data?

To answer these questions, TLS surveys were carried out simultaneously with electronic distance meter (EDM) surveys at each field site such that the elevations of known ground points (EDM measurements) could be compared with those from the unfiltered and filtered TLS data. EDM surveys were carried out using a Trimble S8 Total Station, an EDM accurate to within 1 mm at a range of up to 5,000 m (Trimble, 2007). Seventy-six EDM measurements were obtained at Inch and seventy-eight at Rossbehy. The EDM data were registered to the coordinate system of the TLS point cloud using the coordinates of the HDS targets set up in the field, which were surveyed with both instruments. Figure 7.36 shows the distributions of the surveyed EDM points relative to the unfiltered DEMs generated from the corresponding TLS survey data.

To quantify the error associated with vegetation filtering alone, residuals (z_{TLS}-z_{EDM}) were calculated for both the unfiltered and filtered clouds. This was completed in ArcGIS using a spatial join, whereby information about the elevation of the corresponding TLS elevations was output to the EDM layer. The distances between each point were also output to this layer. Figures 7.37

[14] z_{TLS} = elevation of TLS coordinate; z_{EDM} = elevation of corresponding EDM coordinate

and 7.38 show histograms of these residuals for both the unfiltered (top) and filtered (bottom) clouds for Inch and Rossbehy, respectively. The mean error associated with the unfiltered cloud at Inch was 0.32 m. After filtering, the mean residual was reduced to 0.06 m. At Rossbehy, errors were higher. The mean error associated with the unfiltered cloud here was 0.43 m. After filtering, the mean error was reduced by only 0.1 m to 0.33 m.

Independent sample t-tests were performed to check if the differences between the unfiltered and filtered errors were statistically significant for the Inch and Rossbehy datasets. The tests were performed in the IBM SPSS Statistics v. 21 software package. At Inch, there was a significant difference in the errors for the unfiltered (mean=0.32, SD=0.64) and filtered (mean=0.06, SD=0.32) clouds; $t(111.876)=3.192$, $p<0.001$. Similarly at Rossbehy, there was a significant difference in the errors for the unfiltered (mean=0.43, SD=0.18) and filtered (mean=0.33, SD=0.25) clouds, but only at a lower significance; $t(139.915)=3.005$, $p<0.005$. These results confirm that there is a significant difference between the errors associated filtered vs. unfiltered clouds. This means that the filtered clouds are significantly more accurate representations of the ground surface. As such, changes in sediment volume can be more accurately quantified from the filtered clouds than from the unfiltered clouds.

An assessment of error associated with DEM generation was also carried out. Given that the differences for the filtered and unfiltered clouds were found to be statistically significant for both the Inch and Rossbehy datasets, this was only performed on the Inch dataset. DEMs were generated for both the filtered and unfiltered TLS point clouds using three common interpolation techniques (natural neighbour (NN), inverse distance weighting (IDW), and Empirical Bayesian Kriging (EBK)). Ordinary kriging was deemed inappropriate for the sample data as the data were found to be non-stationary (the variance of the data was not constant). This was checked qualitatively, as recommended by Krivoruchko and Krause (2012), by checking the Voronoi map (figure 7.39). When symbolized by entropy or standard deviation, there should be randomness in the symbolized thiessen polygons for the data to be considered stationary. This was not the case for the Inch data. Because of the way kriging works, the statistical relationship between any pair of points must be similar because the

same model (semivariogram) must work on all points. Kriging was therefore not used for this research. Krivoruchko and Krause (2012) recommended Empirical Bayesian Kriging as an alternative to kriging. With this technique, multiple semivariograms are automatically fitted to the data, so stationarity is not an issue. As such, EBK was deemed suitable for these data.

Residual errors ($z_{DEM} - z_{EDM}$) were calculated for DEMs generated from unfiltered and filtered point clouds. The lowest mean residual for DEMs generated from filtered clouds was achieved using the IDW interpolation and was -0.037 m. The mean residual associated with the NN interpolation was 0.183 m and with EBK was 0.147 m. These differences are likely due to the different ways in which query points are calculated for each method (described in chapter 6). The RMS error associated with each interpolation method was calculated to account for differences in the distributions of the residuals and was found to similar for all three techniques, with RMSE for NN=0.222, for IDW=0.285, and for EBK=0.244.

The final test was to determine whether or not errors associated with filtering *and* DEM generation were still significantly lower for filtered vs. unfiltered clouds. Mean residual errors for unfiltered and filtered clouds associated with each interpolation technique are shown in table 7.3. Paired t-tests were performed in SPSS to determine whether or not the differences in the distributions of the residuals for the DEMs generated from unfiltered and filtered clouds were statistically significantly different from one another. The result for all three techniques was that the lower residuals associated with the DEMs generated from the filtered clouds were statistically significantly different from the residuals associated with the unfiltered clouds (test statistics reported in table 7.3).

An assessment of uncertainty was also carried out for both the Inch and Rossbehy data. As a result of the propagation of error, there is a minimum level of change that it is possible to detect between surveys at t_1 and t_2. This assessment took into account errors associated with registration, filtering, and DEM generation calculated previously. The following error propagation equations were adapted from Wheaton *et al.* (2010). The first (equation 11)

describes propagated error as a result of registration, filtering, and DEM generation:

$$\delta(z)_{within} = \sqrt{(\delta(z)_{RMSE\ Reg})^2 + (\delta(z)_{filt+DEM})^2} \quad (11)$$

Where:

$\delta(z)_{within}$ = propagated error as a result of registration, filtering, and DEM generation

$\delta(z)_{RMSE\ Reg}$ = RMSE of registration

$\delta(z)_{filt+DEM}$ = residuals between EDM elevations and corresponding elevations from DEMs generated from filtered TLS point clouds

The second (equation 12) describes propagated error between DEMs (*e.g.* propagated error for DEMs of difference):

$$\delta(z) = \sqrt{(\delta(z)_{t1})^2 + (\delta(z)_{t2})^2} \quad (12)$$

Where:

$\delta(z)$ = propagated error between DEMs

$\delta(z)_{t1}$ = propagated error of DEM at t_1 (calculated from previous equation)

$\delta(z)_{t2}$ = propagated error of DEM at t_2

For these analyses, mean RMS errors of registration between DEMs and mean residuals associated with DEMs generated from filtered clouds were used. Propagated error between DEMs was calculated under the assumption that the error at t_1 was equal to that at t_2. At Rossbehy, propagated error was calculated for both DEMs generated from clouds registered using semi-permanent targets and for DEMs generated from clouds registered using dGPS coordinates.

At Inch, propagated error as a result of cloud registration and DEM generation was 0.04 m and propagated error between DEMs was 0.05 m. At Rossbehy, for clouds registered using semi-permanent targets, propagated error as a result of cloud registration and DEM generation was 0.29 m and propagated error between DEMs was 0.41 m, significantly higher than for Inch. For clouds

registered using dGPS coordinates, propagated error within DEMs was found to be 0.31 m and propagated error between DEMs was 0.44 m.

As well as linear error margins (for assessing minimum level of detectable elevation change), volumetric error margins were identified (for assessing minimum level of detectable volume change). First, volume error per grid cell was calculated based on the formula:

$$Volume\ error = A\delta_z \tag{13}$$

Where:

A = area of grid cell (in all cases, 0.1 m x 0.1 m) in m^2

δ_z = error in the z direction in metres

This was then converted to volume error per square m (m^3/m^2) and multiplied by the area over which volumes were calculated and is reported in the results section of this chapter.

The results of this uncertainty analysis indicate that:

- The minimum level of elevation change detection possible at Inch is on the order of 0.05 m (0.05 m^3/m^2);
- The minimum level of elevation change detection possible at Rossbehy for DEMs generated from clouds registered using semi-permanent targets is on the order of 0.41 m (0.41 m^3/m^2);
- The minimum level of elevation change detection possible at Rossbehy for DEMs generated from clouds registered using dGPS coordinates is on the order of 0.44 m (0.44 m^3/m^2).

7.4 Results - Rossbehy

Elevation and distance change maps (DEMs of Difference or DODs) were created separately for the beach and foredune at Rossbehy and are presented in this section. For illustrative purposes, an areal photograph showing the general location of the surveyed area is shown in figure 7.40. Locations A and B correspond to the DOD maps presented in figures 7.41 to 7.58 (to help illustrate map orientation) and represent the dune barrier terminus (A) and the southern

periphery of the surveyed area (B) at the time of the corresponding survey. Because not all surveys overlap (due to rapid foredune recession), the precise locations of A and B for DODs from previous/subsequent survey periods are not always the same. To address this, polygons showing the area enclosed by the DOD for the previous period of analysis are included on each of the beach elevation change DODs. In addition, a single coordinate common to multiple DODs is marked on each of the beach elevation change DODs where overlap occurs (depicted as either a triangle, circle, or square). The elevation at t_1 (relative to MSL) is included on the maps for reference.

Beach elevation change and foredune distance change maps are presented sequentially. For the beach elevation change maps, beach lowering (erosion) is shown in varying shades of red and accretion is shown in varying shades of blue. Elevation change below the level of detectable change is shown in gray.

The DODs of the foredune represent distance change in the horizontal (facing the dune scarp). Elevation above mean sea level (MSL) is shown on the y-axis. Mean sea level is equal to +2.3 m ODM. The spring tidal range at the site is 3.2 m. Dune recession in shown in varying shades of red and dune advance in varying shades of blue. Distance change below the level of detectable change is shown in gray.

Tables 7.4 and 7.5 summarise the mean distance/elevation changes and volumetric changes between the beach and foredune, respectively. These results are presented as follows:

Beach and foredune change between 2012-06-28 and 2012-08-05 (figures 7.41 and 7.42)

Between 28 June 2012 and 5 August 2012, there was a net volume loss to the foredune of 322.5±127 m^3. Volume losses within the surveyed area were concentrated at the dune barrier terminus (northern area of surveyed area), where a low beach elevation at the dune toe probably facilitated wave attack. Further south, some volume gains were present at the dune toe. In this area, beach elevations were higher (> 4 m), so accumulated sand was less likely to be vulnerable to wave action. Net beach volume change in the surveyed area was

well below the detectable level (149±2983 m^3). It was therefore unclear as to where the eroded foredune material went during this time.

Beach elevation and volume change between 2012-08-05 and 2012-10-07 (figure 7.43)

Due to technical issues with the laser scanner, only a partial scan was completed in October 2012, which included the beach only. For this reason, only a DOD for the beach was generated. Between 5 August 2012 and 7 October 2012, there was an overall increase in beach volume (+2870.9±1468 m^3). This was the largest increase in beach volume observed over the entire duration of the study period. The majority of volume gains across the surveyed area were concentrated seaward of the dune scarp. The magnitude of elevation/volume change increased with decreasing beach elevation. Given the relative stability in the position of the dune toe (see section on shoreline change, presented later in this section), this material likely came from offshore.

Foredune Distance and volume change between 2012-08-05 and 2012-11-15 (figure 7.44)

Between 5 August 2012 and 15 November 2012, there was little overlap between the scans covering the upper beach, so elevation and volumetric change analyses were only performed on the foredune. During this time, there was a net volume loss of 2059.8±99 m^3. This was the third largest loss observed over the duration of the study period. Similar to what was observed previously (2012-06-28 to 2012-08-05), volume losses were concentrated at the scarp terminus, where elevations above MSL were low, while gains were concentrated in the south (at the dune toe), where elevations were >4 m above MSL.

Foredune distance and volume change between 2012-11-15 and 2013-01-30 (figure 7.45)

Between 15 November 2012 and 30 January 2013, a major foredune recession occurred. This was the first time the semi-permanent markers disappeared from the field site. Initially, it was thought that they may have been stolen, but when the scans were later registered, it was clear that the dunes in which they were placed were completely destroyed. For the first time, no part of the scans overlapped in plan view due to the magnitude of scarp retreat.

The mean distance between the two DEMs was -28.1±0.44 m, with up to 45 m of scarp recession observed. There was a net volume loss of 9469.4±153 m^3. Volume losses were highest at the scarp terminus and decreased southwards. Only the following winter (during the period 2013-12-11 to 2014-01-16) was net volume loss greater than during this period.

Beach and foredune change between 2013-01-30 and 2013-02-28 (figures 7.46 and 7.47)

Foredune volume loss continued between 30 January 2013 and 28 February 2013 (net volume loss = 364.1±137 m^3). Similar to the pattern observed in previous periods (2012-06-28 to 2012-08-05 and 2012-08-05 to 2012-11-15), the majority of losses were concentrated at the barrier terminus, with gains observed in a small area of the southern section of the survey area along the dune toe. However, these gains were at a lower elevation (<4 m) than observed previously. These gains extended onto the beach, decreasing seawards. This pattern, along with field observations, suggests the observed gains may have been due to slumping of the foredune, possibly due to wave undercutting.

Beach and foredune change between 2013-02-28 and 2013-04-19 (figures 7.48 and 7.49)

Between 28 February 2013 and 19 April 2013, volume losses extended across both the entire surveyed beach (net loss = 1369.6±412 m^3) and the foredune (net loss = 836.1±104 m^3). This was the greatest loss of beach material observed over the duration of the study. The pattern of elevation/distance change across both the surveyed beach and scarp was relatively uniform.

Beach and foredune change between 2013-04-19 and 2013-06-05 (figures 7.50 and 7.51)

During this period, relatively little volume change occurred. Across both the surveyed beach and scarp, most of the elevation / distance change was below the detectable level (±0.41 m).

Beach and foredune Change between 2013-06-05 and 2013-12-11 (figures 7.52 and 7.53)

Between 5 June 2013 and 11 December 2013, foredune volume losses were concentrated near the scarp terminus at elevations >3 m. Gains near the dune toe

suggest some slumping may have occurred here. Foredune volume gains in the southern part of the survey area dominate, even at higher elevations. This may suggest dune building as a result of wind-blown sand accumulation occurred here. Foredune volume gains (333.9±11.2m^3) nearly cancelled out volume losses (322.6±11.2m^3).

Foredune distance and volume change between 2013-12-11 and 2014-01-16 (figure 7.54)

During this period, the scarp experienced the highest magnitude recession observed over the entire study period. Distance change across the scarp ranged from -54.3±0.44 m to -33.06±0.44 m, with the greatest recession at the dune barrier terminus. There was a net volume loss to the foredune of 15,337.3±179 m^3. As with the previous winter (period from 2012-11-15 to 2013-01-30), no part of the beach scans overlapped due to the high magnitude of scarp retreat.

Beach and foredune Change between 2014-01-16 and 2014-05-04 (figures 7.55 and 7.56)

Between 16 January 2014 and 4 May 2014, the foredune continued to recede across the surveyed area. There was a net volume loss of 661.9±126 m^3. During this period, the southern periphery of the surveyed area experienced more recession than the barrier terminus. This was the first instance in which this was the case, as, on several previous occasions, the barrier terminus was subject to greater magnitude recession than the area further south. Accretion on the beach in the area behind the foredune suggests overtopping may have occurred here during this period and is interpreted as a washover deposit.

Beach and foredune Change between 2014-05-04 and 2014-07-29

During the final survey period, relatively little change occurred at the site. Elevation and distance change across the majority of the surveyed beach and foredune were below the level of detectable error. Dune toe advance (above the level of detectable change) was observed at the barrier terminus and signals some recovery occurred during this period.

Shoreline change

To better illustrate the observed morphological changes, shoreline positions were extracted from the TLS survey data and mapped against a reference shoreline from March 2012, digitised from an aerial photograph. Shorelines were defined as the boundary between the base of the scarp and the flat beach and are shown in figure 7.59. As outlined previously, major dune toe recessions occurred between 2012-11-15 and 2013-01-30 and between 2013-12-11 and 2014-01-06. Overall, the barrier terminus receded by approximately 100 m.

Summary (Rossbehy)

Tables 7.4 and 7.5 summarise distance/elevation and volume changes for the foredune scarp and beach, respectively. The periods experiencing the highest volumetric losses across the foredune scarp were 2013-11-15 to 2013-01-30 (-9469.4±153 m^3 with scarp recession of up to 45±0.44 m) and 2013-12-11 to 2014-01-16 (-15337.3±179 m^3 with scarp recession of up to 54.3±0.44 m). On three occasions the volumetric error margin exceeded the rate of volume change (2013-06-19 to 2013-06-05, 2013-06-05 to 2013-12-11, and 2014-05-04 to 2014-07-29). Overall beach volume changes, on the other hand, tended to be positive, but in most cases, the volumetric error margin exceeded the rate of volume change. Total net volume changes for the overall duration of the monitoring campaign are summarised as follows:

- **TOTAL NET BEACH VOLUME CHANGE** = +5,306±10,232 m^3
- **TOTAL NET FOREDUNE VOLUME CHANGE** = -28,990±1,345 m^3
- **TOTAL (BEACH+FOREDUNE) NET VOLUME CHANGE** = -23,684±11,577 m^3

7.5 Results - Inch

To assist with interpretation of the elevation change maps for the Inch field site, annotated areal photographs of the site are shown in figure 7.60. The area enclosed by the green polygon is the area over which all surveys overlap and is marked on each of the elevation change maps (figures 7.61 to 7.68). Contours at t_1 have been superimposed on the difference maps for reference. The foredune is represented by the closely spaced contours in the upper part of the DODs. The

grey polygon represents the area over which all the DEMs overlap (herein called "the area of overall overlap"). Mean elevation changes and volume gains and losses, which are reported in table 7.6, were calculated for the data within this area only for direct comparison. These results are summarised as follows:

Elevation and Volume change between 2012-05-24 and 2012-08-06 (figure 7.61)

During the period 24 May 2012 to 6 August 2012, volume losses exceeded gains in the area of overall overlap (net volume loss = 407.2±124 m^3). Most of the losses occurred in the alongshore-trending trough between the embryo dunes and main dune which cuts across the lower part of the study site (see figure 7.60). Because this trough lacks vegetation, it may be more vulnerable to deflation.

Most of the net gains were on the foredune itself. Given the relatively high elevation of the foredune, these gains were likely due to wind-blown sand accumulation, which would have been facilitated by the presence of dense vegetation cover.

Outside of the area of overall overlap (where only the May and August surveys overlap), there are lengthwise alternating bands of gains and losses in the alongshore direction. These bands are in the embryo dune field, where vegetation is probably responsible for trapping wind-blown sediment.

During the nearest corresponding survey period for Rossbehy (28 June 2012 to 5 August 2012), there was a much greater net loss of sediment to the foredune than for the longer survey period at Inch (1.04 m^3 / m^2 at Rossbehy versus 0.16 m^3 / m^2 at Inch). This may be because the elevation of the dune toe at Inch was approximately 3 m higher than at Rossbehy and, therefore, less vulnerable to wave attack. In addition, the presence of an embryo dune field at Inch protects the dune toe from extreme water levels.

Elevation and Volume change between 2012-08-06 and 2012-10-06 (figure 7.62)

Between 6 August 2012 and 6 October 2012, volume losses again exceeded gains in the area of overall overlap (net volume loss = 718.9±124 m^3). This was the highest net volume loss for Inch observed over the duration of the monitoring

campaign. Similar to the previous period, most losses were concentrated in the trough fronting the foredune and most gains occurred on upper part of the foredune. Outside of the area of overall overlap (where only the August and October surveys overlap), the embryo dune field experienced extensive lowering.

During the nearest corresponding survey period for Rossbehy (5 August 2012 to 15 November 2012), which was just over 1 month longer than that of Inch, volume losses to Inch were still considerably lower than those to the Rossbehy foredune (losses were equal to 0.29 m^3 / m^2 for Inch versus 8.49 m^3 / m^2 for the Rossbehy foredune).

Elevation and Volume change between 2012-10-06 and 2013-01-09 (figure 7.63)

During this period, there was relatively little volumetric change (-46.5±124 m^3) in the area of overall overlap. Volume changes across much of the foredune were below the level of detectable change. This was in stark contrast to the volume changes for the nearest corresponding period observed at Rossbehy (15 November 2012 to 30 January 2013), which were the second highest observed over the entire duration of the monitoring campaign. The dune toe at Rossbehy during this period was lower in elevation than at Inch (3 m above MSL versus 4 m above MSL), which, again, probably made the (Rossbehy) foredune more vulnerable to wave attack. While the net volume change for Inch during this period was below the level of detectable change (±124 m^3), the net volume change for Rossbehy was well above this (9469.4±153 m^3).

Elevation and Volume change between 2013-01-09 and 2013-02-27 (figure 7.64)

Between 9 January 2013 and 27 February 2013, the net volume change across the study area was below the level of detectable volume change (+70.7 ±124 m^3). Where elevation changes were above the level of detectable change, erosion occurred primarily on the upper foredune in the eastern section of the surveyed area. Accretion was dominant in the embryo dune field.

At Rossbehy, the nearest corresponding monitoring period was between 30 January 2013 and 28 February 2013. Unlike at Inch, there was a net loss to the

foredune (net volume change = -364.1±137 m^3), although this was relatively low compared to the other survey periods.

Elevation and Volume change between 2013-02-27 and 2013-05-02 (figure 7.65)

During this period, little volume change occurred. The net volume change across the study area was only slightly above the level of detectable volume change (-129.6±124 m^3). Patterns of erosion and accretion were similar to those observed during the previous monitoring period. At Rossbehy, there was also a net loss to the foredune during the nearest corresponding monitoring period (2013-02-28 to 2013-04-19), although it was much greater (volume change = -3.54 m^3 / m^2 for Rossbehy versus -0.05 m^3 / m^2 for Inch).

Elevation and Volume change between 2013-05-02 and 2013-06-20 (figure 7.66)

From 2 May 2013 to 20 June 2013, the net volume change across the study area was again below the level of detectable volume change (-81.0 ±124 m^3). Where elevation change was above the level of detectable change, patterns of erosion and accretion were similar to those observed during the previous three monitoring periods. At Rossbehy, for the nearest corresponding monitoring period (4 April 2013 to 6 June 2013) net volume change for the foredune was also below the level of detectable change.

Elevation and Volume change between 2013-06-20 and 2014-03-12 (figure 7.67)

The 2013-2014 winter storms had a major impact on Rossbehy and many other west coast beaches, yet this was not evident at the Inch study site. In fact, during the period between 20 June 2013 and 12 March 2014, there was a net volume gain of 277.7±124 m^3 in the area of overall overlap. This might be explained by the fact that this monitoring period was considerably longer than that of previous periods, which meant there was more time for sand accumulation to occur. However, even if the net volume changes for each of the previous periods were added together, this figure far exceeds any gains observed at the site previously. This result is counterintuitive. However, much of the beach seaward of the area of overall overlap (the area covered by both the June 2013 and March 2014

surveys, but not the others) had lowered considerably (up to 3.56 m). In fact, between 20 June 2013 and 12 March 2014, the embryo dune field had all but disappeared (figure 6.69). This may help to explain the observed volume gain in the study area in that the embryo dune field may have shielded the foredune from extreme waves during the winter 2013/2014 storms, resulting in minimal volume losses to the surveyed area. As such, it demonstrates the potential significance of embryo dunes in protecting the foredune.

Elevation and Volume change between 2014-03-12 and 2014-08-28 (figure 7.68)

During this, the final survey period, the net volume change (803.2 ±124 m^3) was highest of all the surveyed periods. Widespread volume gains occurred across the embryo dune field, in the trough and across the foredune. This material was likely brought back onshore following the destruction of the embryo dunes during the winter 2013/2014. No similar recovery to the foredune was observed at Rossbehy.

Summary (Inch)

Table 7.6 summarises elevation and volume changes observed at Inch. Rates of volume change varied from -0.0048 ±0.0008 m^3 m^2 day to 0.0019±0.0003 m^3 m^2 day in the area of overlap. The greatest net loss (-718.9±124 m^3) occurred between 6 August 2012 and 6 October 2012 (also the period with the lowest rate of volume change). The greatest net gain (803.2 ±124 m^3) occurred between 12 March 2014 and 28 August 2014 (also the period with the highest rate of volume change). The total net volume change for the overall duration of the monitoring campaign was below the level of detectable change (-231.6±992 m^3).

Figure 7.70 shows how observed rates of volume change at Inch compare with Rossbehy. Overall rates of volume change at Inch (average = -0.0007 m^3 per m^2 per day) were relatively low compared to rates of beach volume change at Rossbehy (average = 0.0025 m^3 per m^2 per day) and the relatively large rates of scarp volume change at Rossbehy (average = -0.16 m^3 per m^2 per day). Net volume change over the duration of the monitoring period was also very low (-231 m^3) compared to Rossbehy beach (+5,306 m^3) and Rossbehy scarp (-28,990 m^3). There was no obvious (qualitative) relationship between rates of volume

change observed over the course of the overall monitoring period at Inch and Rossbehy. A statistical assessment of this is hindered by the fact that the individual monitoring periods at each site did not span the same duration.

8 Relationships between observed morphologic change and storms

An assessment of storm events that occurred during the study period was undertaken in an effort to link storm characteristics with observed morphodynamic behavior. This chapter outlines how storm events were defined and extracted from local weather station data and goes on to present the results of statistical analyses between storm characteristics and the observations of morphological change reported in the previous chapter.

8.1 Storm events and their characteristics during the study period

For this research, storm events were defined based on modelled nearshore wave heights. According to the definition given by Boccotti (2000), storm events can be defined as a sequence of sea states in which significant wave height, $H_s(t)$, exceeds a fixed threshold, h_{crit}, and does not fall below this threshold for a continuous time interval of greater than 12 hours. The fixed threshold is equal to $1.5*(H_s(t))$, or 1.5 times the mean annual significant wave height. Storms were defined in this way because often storm waves are responsible for dune erosion. The meteorological characteristics for each event identified in this way were obtained from local weather station data (provided by Mr. Chris Byrne, http://ventryweather.com/).

8.1.1 WAM Data

In order to identify storms, wave data were required. Nearshore wave heights near Inch and Rossbehy were hindcast using the WAve prediction Model (WAM) (The WAMDI Group, 1988). The WAM is a third generation numerical wave prediction model which predicts the propagation of offshore waves across a spherical grid using wind data as input. The model is unique in that it solves the wave transport equation explicitly without constraints on spectral shape. Key parameters represented in the WAM model include the local rate of change of wave energy density, propagation in geographical space, shifting of frequency and refraction due to the spatial variation of the depth and current, and the effects of generation and dissipation of the waves including wind input, white capping dissipation, non-linear quadruplet wave–wave interactions and bottom friction dissipation (Monbaliu *et al*., 2000).

The WAM for Dingle Bay was set up and validated by O'Shea (2015) using hourly wind data obtained from the Marine Institute's M3 offshore wave buoy, located approximately 30 nautical miles (56km) southwest of Mizen Head (51.2166°N 10.5500°W). This was the nearest buoy to the study site. The bathymetric data for the model set up were derived by Sala (2010) from a combination of the Marine Institute INFOMAR survey data, an1850s admiralty chart, and recent areal photographs.

The model was run at hourly intervals from 2011-01-01 00:00 to 2014-04-30 04:00 by Dr. Jimmy Murphy, of UCCs (former) Hydraulics and Maritime Research Centre (HMRC) (now part of MaREI). Figure 8.1 shows the model domain and flexible mesh, whose nearshore triangular resolution can resolve elements on the bed of up to 5,000 m^2 (Sala, 2010). Outputs of time, significant wave height (H_s), wave period (T_z) and mean wave direction were extracted for the five points shown in figure 8.2 and exported to a Microsoft Excel spread sheet. Events were identified using data for the middle coordinate (third from the top), as it was nearest the Rossbehy study area.

The following bullet points summarise the steps taken to extract information about the events based on the definition from above from the raw data.

- The mean annual significant wave height (H_s) was calculated to be 0.69 m, although it should be noted that this is based on the short record over which data was available (January 2011-April 2014). There is some evidence to suggest, however, that this is a reasonable figure. For the corresponding period, the offshore mean annual significant wave height recorded at the M3 wave buoy was 3.04 m, 77% higher than that of the simulated H_s at the nearshore coordinate. This difference is in line with previous work published by Sala (2010), which was based on observational data obtained from a nearshore wave gauge deployed 1.5 km seaward of Rossbehy in 2007. Sala (2010) found that swell waves were lower on average in the nearshore zone than at the offshore M3 wave buoy. For one particular storm event, wave heights recorded at the nearshore gauge were 63% lower than offshore. While that study was limited in that it was based on only three months of wave gauge data, it suggests that nearshore significant wave heights are

considerably lower than offshore significant wave heights. Such a difference is exemplified in the simulated data and the corresponding M3 buoy data. Longer-term observational data, however, would be required to more precisely quantify the nearshore H_s.

- Following the definition proposed by Boccotti (2000), the critical threshold (h_{crit}) for an event to be considered a storm was calculated to be 1.03 m (1.5 times H_s). Data were filtered in Excel such that only information associated with times when H_s exceeded h_{crit} was left (these were considered "potential events").
- Time differences between each row in the database were calculated. If the time difference between the rows was one hour, the rows were considered to be part of a single "potential event" and given an ID (from 1 to n).
- The duration of each of these potential events was then calculated and potential events lasting <12 hours were discarded. New event IDs were generated for the remaining discrete events.

These remaining events therefore represented times when H_s exceeded h_{crit} for a minimum duration of 12 hours. Information contained in the spreadsheet about these events included event ID, start date, end date, event duration, lag time (time between events), mean H_s, maximum H_s, peak period, and mean wave direction. Table 8.1 summarises this information for the duration of the morphologic monitoring period only (May 2012-April 2014).

The storm criterion (h_{crit}=1.03 m) was sufficient to raise water levels to the level at which storm waves would come into contact with the surveyed areas at Inch and Rossbehy. The definition was further deemed satisfactory in that all major weather events recorded at Met Eireann's Valentia weather station between May 2012-April 2014 were represented in the modeled dataset. Met Eireann (2014) reported that storm force winds occurred at Valentia on 5 different days – 18th, 26th, and 27th December 2013 and 1st and 12th February 2013. All five events were represented in the modeled datasets.

Between 1 January 2011 and 30 April 2014 (the entire duration spanned by the WAM dataset), a total of 127 events occurred, 72 of which occurred between May 2012 and April 2014 (the period when the morphologic surveys were

carried out). Figure 8.3 illustrates the distribution of storm event duration amongst the 72 events that occurred during the morphological monitoring period. Many (28) events lasted <1 day, with longer duration events occurring less frequently. Six events lasted longer than six days. These were likely to be the result of multiple back-to-back storms. The longest modelled single event occurred between 2013-12-13 06:00:00 and 2014-01-18 15:00:00. That particular winter was "severely affected by an exceptional run of winter storms" (Met Eireann, 2014). As such, this 633-hour-long-duration event obviously represented multiple storms resulting in the exceedance of H_{crit} for an extended period of time.

Figure 8.4 shows modelled event frequency by month for the morphologic monitoring period. The highest frequencies of events tended to occur in the winter months (December, January, February), although because the 633-hour-long-duration event of December 2013 was recorded as one single event, it somewhat misleadingly appears that there is a relatively low frequency of events in December 2013.

Various Excel formulae, including arrays, were written to extract information about events that occurred between morphological surveys. These are summarised for the WAM data in figure 8.5 and for local weather station data in figure 8.6. Event characteristics during these periods are examined in the following section, along with local weather station data, in relation to observed morphological change.

8.1.2 Local weather station data

A weather station with a HOBO Micro Station Logger [H21-002] was initially set up at the Inch field site (figure 8.7) to record local weather conditions over the duration of the study. The station was set up to record wind speed, gust speed, wind direction, and rainfall at 5-minute intervals. Due to numerous technical difficulties, only a small, discontinuous record was obtained from this instrument. This record spanned from 6 August 2012 to 5 September 2012 (a duration of 29 days); 15 October 2012 to 26 October 2012 (a duration of 11 days); and 21 August 2013 to 3 September 2013 (a duration of 13 days). A continuous record, however, was available from a weather station maintained by

a local weather enthusiast in Ventry, approximately 25 km west of Inch. Data from this station were kindly provided free of charge for this research. The Ventry data were recorded on a sheeva plug (a miniature computer) and uploaded to a designated web server in real time using Meteohub, a miniaturized weather server that records and uploads weather station data via file transfer protocol (ftp) to a web server. The web server can be accessed from any computer. Various parameters, including sea level pressure (SLP), wind speed, gust speed, and wind direction, were recorded at irregular intervals of on the order of a few seconds. Due to the sheer volume of available data (as the temporal resolution of the data was so high), data were extracted from the raw datasets at roughly half hourly intervals.

The Ventry data were compared to the available Inch data to check if similar weather conditions prevailed at the two sites. Figures 8.8, 8.9, and 8.10 show wind speeds and wind roses for Inch and Ventry from the first, second, and third periods, respectively, in which data were recorded at Inch. The wind speeds used to construct the graphs in these figures are from instantaneous wind speeds averaged at half hourly (or approximately half hourly) intervals. Running means (with 48 hour periods) have been superimposed on this data for visual clarity.

Overall, wind speed patterns generally mimic one another at both sites, although not in all instances. To more robustly assess the similarity between the datasets, a Pearson product-moment correlation coefficient was computed. There was a moderate positive correlation between the two datasets, with $r=0.58$, $n=2568$, and $p<0.001$. This indicates that the wind speeds recorded at Ventry are broadly similar to those recorded at Inch over the periods in which data were available for both sites.

The direction of prevailing winds differed somewhat between the two sites. Between 6 August and 5 September 2012, the prevailing winds at Inch tended to come from the northeast, with northeasterlies dominating 45% of the time. Southwesterly winds prevailed at Ventry, with southwesterlies dominating between 35 and 40% of the time. Between 15 October and 26 October 2012, southerly winds prevailed Inch, while northeasterly winds were dominant at Ventry. The percent of time during which southerly winds were prevalent was

similar at both sites (30% at Inch and 35% at Ventry). During the period 21 August and 3 September 2013, westerlies dominated at both sites, but for less of the time at Ventry (55%) than at Inch (almost 90%). The elevation and aspect of the two sites, as well as the position of the weather stations in relation to built structures, may play a role in the directional discrepancies. The Inch weather station was located at an elevation of <10 m ODM (it was adjacent to the beach), while the Ventry weather station was located approximately 50 m ODM and almost 2 km inland. While the Ventry weather station was located in a relatively open field, the Inch station was located in an enclosed area. Although the Inch weather station sensor was located high above the walls of the enclosure, it is possible that the wall and/or nearby buildings were responsible for obstructing local winds, particularly those coming from the north. This may explain some of the discrepancies in wind speeds and directions at the two sites.

Overall, these data suggest that meteorological conditions at both sites are broadly similar, save for some discrepancies, which may stem from the specific locations of the weather stations in relation to the coast and built structures. As such, subsequent data analysis was performed on the Ventry data.

For each of the events identified from the WAM data (table 8.1), corresponding meteorological conditions were extracted in Excel. These included mean wind speed, maximum gust speed and prevailing wind direction, which are given in table 8.2.

Mean wind speeds for the events that occurred during the morphologic monitoring period ranged from 0.78 m/s to 10.68 m/s. The histogram shown in figure 8.11 shows how mean wind speeds for the events breaks down. Mean wind speeds for the majority of events were between 4 and 6 m/s. Few events had mean wind speeds of <2 m/s, and 2 had mean wind speeds of greater than 10 m/s. The events with the highest mean wind speeds were 2014-01-23 22:00:00 to 2014-01-28 11:00:00 (10.68 m/s) and 2014-02-07 04:00:00 to 2014-02-10 10:00:00 (10.3 m/s).

Maximum gust speeds for the events that occurred during the monitoring period ranged from 2.2 m/s to 28.35 m/s. The histogram shown in figure 8.12 shows how maximum gust speeds for the events break down. Gust speeds for the

majority of events were between 5 m/s to 10 m/s. Two events had maximum gust speeds of between 20 m/s and 25 m/s. These were 2013-12-13 06:00:00 to 2014-01-08 15:00:00 (the long duration 633-hour event) (20.38 m/s) and 2014-01-31 04:00:00 to 2014-02-04 17:00:00 (23.46 m/s). The maximum gust speed recorded during the monitoring period occurred during the event 2014-02-10 16:00:00 to 2014-02-16 07:00:00 (28.35 m/s).

Figure 8.13 shows the frequency of events with prevailing wind directions from the north, northeast, east, southeast, south, southwest, west, and northwest. The prevailing wind direction associated with the majority of events was westerly, the second most frequent being from the southwest and the third from the south.

8.2 Linear and multiple regression analyses

Linear regression analyses have been used to investigate morphological relationships between beach and dune morphological change and various parameters (*e.g.* Nolan *et al.*, 1999; Saye *et al.*, 2005). For this study, this approach was chosen to investigate the relationships between observed morphological change at Inch and Rossbehy and various storm characteristics.

Information about event characteristics from WAM data and local weather station data was broken down by morphological monitoring period and graphed. Tide data associated with events were also included in the analyses. These data were obtained from the Irish National Tide Gauge Network Real Time Data for Castletownbere (the nearest gauge to Inch and Rossbehy) and corrected to reflect water levels at Inch/Rossbehy (high/low tides at Inch/Rossbehy are approximately 20 minutes ahead of Castletownbere).

Scatter plots were produced and correlation coefficients calculated to examine whether or not there existed any first order linear relationships between:

1. rates of mean volume change and event frequency for events identified as having occurred over the duration of each monitoring period;
2. rates of mean volume change and mean event duration for events identified as having occurred over the duration of each monitoring period;

3. rates of mean volume change and maximum event duration for events identified as having occurred over the duration of each monitoring period;
4. rates of mean volume change and maximum tidal level for events identified as having occurred over the duration of each monitoring period;
5. rates of mean volume change and mean lag time between events identified as having occurred over the duration of each monitoring period;
6. rates of mean volume change and mean significant wave height for events identified as having occurred over the duration of each monitoring period;
7. rates of mean volume change and maximum significant wave height for events identified as having occurred over the duration of each monitoring period;
8. rates of mean volume change and mean wave period for events identified as having occurred over the duration of each monitoring period;
9. rates of mean volume change and mean wind speed for events identified as having occurred over the duration of each monitoring period; and
10. rates of mean volume change and maximum wind speed for events identified as having occurred over the duration of each monitoring period.

Relationships between rates of volume change and mean wave direction were not examined, as the mean wave direction for all events that occurred during the monitoring period varied by only 5 degrees (storm waves for all events that occurred were from the west – between 257-262°). This was likely due to the constraining influence of headlands to north and south.

A semi-quantitative analysis of the relationship between rates of volume change and prevailing wind direction was performed. This involved plotting the frequency of events with prevailing wind directions in a given direction for each monitoring period and qualitatively comparing these with rates of volume change. Because morphological monitoring periods differed for each site, these analyses were completed separately for Inch and Rossbehy.

The relationships between volume change and the individual storm variables under investigation are illustrated for Rossbehy and Inch in figures 8.14 to 8.46. These illustrations include:

- graphs showing rates of volume change for each of the morphological monitoring periods
- graphs showing the magnitude of each storm variable under investigation (event frequency, mean and maximum event duration, etc.), also broken down by morphological monitoring period, and
- scatter plots showing rate of volume change versus each of the storm variables, with correlation coefficients and their significance.

Rates of volume change and event frequency at Rossbehy (figures 8.14-8.15)

Figures 8.14 and 8.15 illustrate the relationships between rates of volume change and event frequency at Rossbehy beach and foredune, respectively. These figures show rates of volume change for each of the morphological monitoring periods (figures 8.14a and 8.15a), event frequency (8.14b and 8.15b) and corresponding scatter plots (8.14c and 8.15c).

The highest frequency of events (15) occurred over the period 16 January to 4 May 2014. During this period, little change in beach and dune volumes occurred. During the period when the highest rate of volume change occurred on the foredune (between 11 December 2013 and 16 January 2014), relatively few events (4) were recorded, although this was the period when back-to-back storms resulted in a 633-hour-long duration event. When volume change is plotted against event frequency (figures 8.14c and 8.15c), there is considerable variation around the trend lines. There were weak positive correlations between rates of volume change and event frequency for both the beach ($r=0.09$, $p=0.86$) and the foredune ($r=0.12$, $p=0.76$). Neither of these was statistically significant. These results indicate that event frequency alone is not a predictor of volume change at Rossbehy.

Rates of volume change and event frequency at Inch (figure 8.16)

At Inch, there is a similarly tenuous relationship between rates of volume change and event frequency. Figure 8.16 shows the rate of volume change for each of

the monitoring periods there (8.16a), event frequency broken down by these monitoring periods (8.16b), and the corresponding scatter plot (8.16c). The period in which the most events occurred was 20 June 2013 to 12 March 2014, yet relatively little change occurred at the site. During the period where the rate of volume change was highest (6 August to 6 October 2012) relatively few events occurred. Like at Rossbehy, there was a weak positive relationship between rate of volume change and event frequency. This relationship was not statistically significant ($r=0.30$, $p=0.47$).

Rates of volume change and event duration at Rossbehy (figures 8.17-8.20)

Figures 8.17-8.20 illustrate the relationship between rates of volume change and event duration at Rossbehy beach and foredune. These figures show rates of volume change for each of the morphological monitoring periods (figures 8.17a-8.20a), mean duration of events for each of the morphological monitoring periods (8.17b and 8.18b), max duration of events for each of the morphological monitoring periods (8.19b and 8.20b), and corresponding scatter plots (8.17c-8.20c).

The period with the longest mean duration of events was 11 December 2013 to 16 January 2014. This was also the period with the longest duration event (633 hours). The highest rate of volume change for the foredune was recorded during this period. During the period with the second highest mean and maximum duration of events (15 Nov 2012 to 30 Jan 2013), the second highest rate of volume change for the foredune was recorded. While there were no statistically significant correlations observed between mean duration of events and rate of beach volume change ($r=-0.59$, $p=0.166$) or maximum duration of events and rate of beach volume change ($r=-0.39$, $p=0.393$), there were strong negative relationships between both mean duration of events and rate of foredune volume change ($r=-0.96$, $p<0.001$) and maximum duration of events and rate of foredune volume change ($r=-0.93$, $p<0.001$). These results suggest that event duration may be a strong predictor of foredune volume change, but not necessarily beach volume change at Rossbehy.

Rates of volume change and event duration at Inch (figures 8.21-8.22)

Figures 8.21 and 8.22 illustrate the relationship between rates of volume change and event duration at Inch. These figures show rates of volume change for each of the morphological monitoring periods (figures 8.21a and 8.22a), mean and max duration of events (figures 8.21b and 8.22b), and corresponding scatter plots (figures 8.21c and 8.22c).

While like at Rossbehy there were positive relationships between rates of volume change and event duration, these were weak. Also, neither of the correlations was statistically significant ($r=0.51$, $p=0.20$ for rate of volume change vs. mean duration of events; $r=0.37$, $p=0.37$ for rate of volume change vs. max duration of events).

Rates of volume change and maximum tidal levels associated with events at Rossbehy (figures 8.23 and 8.24)

Figures 8.23 and 8.24 illustrate the relationship between rates of volume change and the maximum tidal levels associated with events at Rossbehy beach and foredune, respectively. These show rates of volume change (figures 8.23a and 8.24a), max tidal level for events that occurred during the corresponding monitoring periods (figures 8.23b and 8.24b), and corresponding scatter plots (figures 8.23c and 8.24c).

Given the fact that water levels range from 0 m to +4.36 m ODM at the site (MSL = +2.3 m ODM), the maximum tidal levels associated with storm events during the monitoring period were not particularly high (up to 2.94 m ODM). For both the beach and the foredune, there was a moderate negative relationship between rate of beach volume change and maximum tidal levels associated with events ($r=-0.45$, $p=0.31$ for the beach; $r=-0.48$, $p=0.19$ for the foredune). Neither was statistically significant.

Rates of volume change and maximum tidal levels at Inch (figure 8.25)

Figure 8.25 illustrates the relationship between rates of volume change and the maximum tidal levels associated with events at Inch. Figure 8.25 shows the rate of volume change for each of the monitoring periods there (8.25a), maximum

tidal levels associated with events that occurred during the monitoring periods (8.25b), and the corresponding scatter plot (8.25c).

Unlike at Rossbehy, there was a weak positive relationship between rate of volume change and maximum tidal levels associated with events. This was not, however, statistically significant (r=0.33, p=0.42). The trend reversal is therefore not likely to be meaningful.

Rates of volume change and time between events at Rossbehy (figures 8.26 and 8.27)

Figures 8.26 and 8.27 illustrate the relationship between rates of volume change and time between events at Rossbehy beach and foredune, respectively. These show rates of volume change (figures 8.26a and 8.27a), mean time between events (figures 8.26b and 8.27b), and corresponding scatter plots (figures 8.26c and 8.27c).

For the beach, there was a weak negative relationship between rate of beach volume change and mean time between events (r=-0.32, p=0.48). For the foredune, this trend was reversed (r=0.56, p=0.11). Given the fact that neither of these correlations were statistically significant, this trend reversal is likely not meaningful.

Rates of volume change and time between events at Inch (figure 8.28)

Figure 8.28 illustrates the relationship between rates of volume change and time between events at Inch. Figure 8.28 shows the rate of volume change for each of the monitoring periods there (8.28a), mean time between events broken down by the monitoring periods (8.28b), and the corresponding scatter plot (8.28c).

Similar to Rossbehy, there was a moderate negative relationship between rate of volume change and mean time between events. This was not statistically significant (r=-0.44, p=0.27).

Rates of volume change significant wave heights at Rossbehy (figures 8.29-8.32)

Figures 8.29-8.32 illustrate the relationship between rates of volume change and significant wave heights at Rossbehy beach and foredune. These figures show rates of volume change for each of the morphological monitoring periods

(figures 8.29a-8.32a), mean H_s associated with events for each of the morphological monitoring periods (8.29b and 8.30b), max H_s for events for each of the morphological monitoring periods (8.31b and 8.32b), and corresponding scatter plots (8.29c-8.32c).

There was little variation in mean significant wave heights associated with events for each period. Maximum significant wave heights for events that occurred during these periods ranged from 1.6 m (occurred between 29 June and 5 August 2012) to 2.97 m (occurred between 11 December 2013 and 16 January 2014).

Moderate to strong negative correlations between rates of volume change and mean and max significant wave height for events were observed, although none were statistically significant (r=-0.67, p=0.10 for rate of beach volume change vs. mean significant wave height for events; r=-0.5, p=0.17 for rate of foredune volume change vs. mean significant wave height for events; r=-0.56, p=0.2 for rate of beach volume change vs. max significant wave height for events; r=-0.58, p=0.10 for rate of foredune volume change vs. max significant wave height for events).

Rates of volume change and significant wave heights at Inch (figures 8.33 and 8.34)

Figures 8.33 and 8.34 illustrate the relationships between rates of volume change and significant wave heights. These figures show rates of volume change for each of the morphological monitoring periods (figures 8.33a and 8.34a), mean H_s associated with events for each of the morphological monitoring periods (figure 8.33b), max H_s for events for each of the morphological monitoring periods (8.34b), and corresponding scatter plots (figures 8.33c and 8.34c).

Unlike at Rossbehy, there were moderate to strong positive relationships between rates of volume change and significant wave heights. At Rossbehy, such relationships were negative (but not statistically significant). While the correlation between rate of volume change and maximum H_s at Inch was not statistically significant (r=0.62, p=0.100), the strong positive correlation between rate of volume change and mean H_s was (r=0.74, p=0.036). This result indicates higher significant wave heights during storms are associated with higher rates of volume gain at the site.

Rates of volume change and peak period at Rossbehy (figures 8.35 and 8.36)

Figures 8.35 and 8.36 illustrate the relationship between rates of volume change and mean peak wave periods associated with events at Rossbehy beach and foredune, respectively. These show rates of volume change for the beach and foredune (figures 8.35a and 8.36a), mean peak period associated with events for the corresponding monitoring periods (figures 8.35b and 8.36b), and corresponding scatter plots (figures 8.35c and 8.36c).

There is little variation in mean peak wave periods associated with events for each period, ranging from 7-10 seconds. There was a weak negative relationship between beach volume change and mean peak period associated with events ($r=-0.34$, $p=0.46$) and a neutral relationship between foredune volume change and mean peak wave period ($r=0$, $p=0.998$). Neither were statistically significant.

Rates of volume change and peak period at Inch (figure 8.37)

Figure 8.37 shows the relationship between rates of volume change and mean peak wave periods associated with events at Inch. There was a moderate positive relationship between rate of volume change and mean peak period associated with events ($r=0.57$, $p=0.14$). At Rossbehy, this relationship was either negative or neutral. Given the fact that none of these correlations were statistically significant, this trend reversal is likely not meaningful.

Rates of volume change and event wind speeds at Rossbehy (figures 8.38-8.41)

Figures 8.38-8.41 illustrate the relationships between rates of volume change and wind speeds associated with events at Rossbehy. These figures show rates of volume change for each of the morphological monitoring periods (figures 8.38a-8.41a), mean wind speed associated with events for each of the morphological monitoring periods (8.38b and 8.39b), max gust speed associated with events for each of the morphological monitoring periods (8.40b and 8.41b), and corresponding scatter plots (8.38c-8.41c).

There is little variation in mean wind speed associated with events for each period. Maximum gust speeds for events that occurred during these periods ranged from 10.65 m/s (7 Oct to 15 Nov 2012) to 28.35 m/s (16 Jan to 4 May 2014). Relationships between rates of volume change and mean wind speeds

associated with events were positive (r=0.09, p=0.84 for rate of beach volume change vs. mean wind speed associated with events and r=-0.17, p=0.66 for rate of foredune volume change vs. mean wind speed associated with events). Conversely, relationships between rates of volume change and max gust speeds associated with events were negative (r=-0.39, p=0.39 for rate of beach volume change vs. max gust speed associated with events and r=-0.29, p=0.45 for rate of foredune volume change vs. max gust speed associated with events). None of the correlations were statistically significant.

Rates of volume change and event wind speeds at Inch (figures 8.42 and 8.43)

Figures 8.42 and 8.43 illustrate the relationship between rates of volume change and wind speeds associated with events at Inch. These figures show rates of volume change for each of the morphological monitoring periods (figures 8.42a and 8.43a), mean wind speed and max gust speed associated with events (figures 8.42b and 8.43b), and corresponding scatter plots (figures 8.42c and 8.43c).

There was a very weak negative relationship between rate of volume change and mean wind speeds associated with events (r=-0.04, p=0.92). Conversely, there was a moderate positive relationship between rate of volume change and maximum gust speeds associated with events (r=0.46, p=0.24). None of the correlations were statistically significant.

Rates of volume change and wind direction at Rossbehy and Inch (figures 8.44-8.46)

Figures 8.44-8.46 qualitatively illustrate the relationships between prevailing wind directions associated with events and rates of volume change at Rossbehy beach, Rossbehy foredune, and Inch, respectively. These figures show rates of volume change (figures 8.44a-8.46a) and event frequencies with events characterized by predominantly northerly, northeasterly, easterly, southeasterly, southerly, southwesterly, westerly, and northwesterly winds broken down by monitoring period. For most events during the monitoring period, winds were westerly (n=32) or southwesterly (n=21). Qualitatively, there is no one direction that appears to coincide with high rates of volume change for either Inch or Rossbehy.

Tables 8.3, 8.4, and 8.5 provide summaries of the data used in the above analysis for Rossbehy beach, Rossbehy scarp, and Inch, respectively, with p-values of statistically significant correlations highlighted[15]. The lack of statistically significant regression results suggests that either no relationships exist between morphological change and the variables under investigation, the sample population of data was too small to model such relationships, the relationships were not linear, or the relationships were dependent upon multiple variables together. If either of the latter two options is true, it is possible to conduct further analyses.

Given the high degree of variability of the data presented in the scatter plots, it would require considerable effort to determine the function that would provide the optimal fit for the data, if one even exists. As such, a non-linear regression analysis was not attempted. Less effort is required to perform a multiple regression analysis. To further investigate any additional potential relationships, this approach was pursued.

Relationships between the rates of volume change and the following variables (data presented in tables 8.3, 8.4, and 8.5) were included in the multiple regression analysis:

Rossbehy Beach:

- Event frequency
- Mean event duration
- Max tidal level
- Mean time between events
- Mean H_s
- Mean peak period
- Max gust speed

Rossbehy foredune:

- Event frequency

[15] Statistical significance is an indicator of the likelihood that observed correlations are not due to chance or sampling error.

- Mean event duration
- Max tidal level
- Mean time between events
- Max H_s
- Mean peak period
- Max gust speed

Inch

- Event frequency
- Mean event duration
- Max tidal level
- Mean time between events
- Mean H_s
- Mean peak period
- Max gust speed

Choice of mean versus max event duration, H_s, and wind speed was made based on which was more strongly correlated with rates of volume change (which had the higher correlation coefficient) in the simple linear regression analysis.

When performing a multiple regression analysis, there are some principal assumptions that must be satisfied in order to obtain a valid result. These are described and commented upon with respect to this analysis as follows:

1. **The dependent variable should be measured on a continuous scale**. The dependent variables, rates of volume change in this case, which are measured in (m^3 m^2 day), satisfy this criterion.
2. **There are two or more independent variables, which can be either continuous (i.e., an interval or ratio variable) or categorical (i.e., an ordinal or nominal variable).** The seven independent variables, introduced previously, meet this criterion.
3. **The values of the model residuals are independent (independence of observations)**. This can easily be checked by looking at the Durbin-Watson test statistic, which ranges from 0 to 4. A value of 2 means that

there is no autocorrelation in the sample. The values for the proposed analyses were as follows:

a. Rossbehy beach: 3.03
b. Rossbehy scarp: 2.305
c. Inch: 3.37

Because this assumption was violated for the Rossbehy beach and Inch data, these were not included in the multiple regression analysis.

4. **The relationship between variables is linear**. Scatter plots shown previously in this chapter indicate the relationships between variables are roughly linear.
5. **The variance of residuals is constant (homoscedasticity).** This can be checked by plotting standardized residuals against predicted values and checking that scores are roughly concentrated in the centre (about the 0,0 coordinate). The plot for the Rossbehy foredune data, shown in figure 8.47, confirms that this assumption is met.
6. **The values of residuals are normally distributed**. The distribution of the residuals, shown in fig. 8.48, generally follows a normal curve.

Since the Rossbehy beach and Inch data violated the independence of observations assumptions, the multiple regression analysis was only performed for the Rossbehy scarp data.

Analysis of variance (ANOVA) results showed that the seven variables together (event frequency, mean event duration, max tidal level, mean time between events, max H_s, mean peak period and max gust speed) were not statistically significant predictors of foredune volume change at Rossbehy, $F(7, 1)=17.814$, $p=0.181$, $R^2=0.992$. The variable with the highest standard error was the maximum tidal level for events (Std. error=0.266). Taking this into account, the analysis was repeated, this time without the maximum tide variable.

The new model indicated that the remaining six variables together (event frequency, mean event duration, mean time between events, max H_s, mean peak period and max gust speed) are statistically significant predictors of foredune volume change at Rossbehy, $F(6, 2)=40.513$, $p<0.05$, $R^2=0.992$. However, only

mean event duration was found to statistically significantly contribute to the prediction, $p<0.05$.

Further examination of the variables aside from event duration (event frequency, mean time between events, max H_s, mean peak period and max gust speed) was performed to see whether or not they could together be useful predictors of foredune erosion. However, the independence of observations assumption was violated (DW statistic = 0.646) and the results were not statistically significant ($p=0.079$).

A stepwise linear regression was performed to investigate the multiplicative effect of the storm parameters. This was performed using only the event frequency, mean event duration, mean time between events, max H_s, mean peak period and max gust speed variables. This time, two variables were found to significantly affect rates of foredune erosion, event duration, $F(7, 1)=87.448$, $p<0.001$, $R^2=0.926$, and maximum significant wave height, $F(6,2)=129.323$, $p<0.001$, $R^2=0.977$. The R^2 values are positive, but, in the case of multiple regression analysis, this does not mean an increase in event duration is correlated with a decrease in dune erosion. The R^2 values simply represent the amount of variability in dune erosion that is accounted for (explained) by the variables.

The results of the stepwise linear regression indicate that together, event duration and significant wave height may be predictors of dune erosion at Rossbehy. This result is fairly intuitive, but useful in that it helps to quantify the relative importance of these two variables against the others under investigation.

Summary

- Strong negative simple linear relationships exist between mean duration of events and rate of foredune volume change ($r=-0.96$, $p<0.001$) and maximum duration of events and rate of foredune volume change ($r=-0.93$, $p<0.001$) at Rossbehy, which suggests that event duration may be a strong predictor of foredune volume change at the site.
- There was a strong positive relationship between rate of volume change and mean H_s associated with events at Inch ($r=0.74$, $p<0.05$). This result indicates, counter intuitively, that higher significant wave heights during

storms are associated with higher rates of volume gain at the site.

- Multiple regression analysis of variance (ANOVA) results showed that together, event frequency, mean event duration, mean time between events, max H_s, mean peak period and max gust speed are statistically significant predictors of foredune volume change at Rossbehy, $F(6, 2)=40.513$, $p<0.05$, $R^2=0.992$. Event duration was found to be the most influential variable in the model.
- The results of a stepwise discriminant analysis suggest that of the event frequency, mean event duration, mean time between events, max H_s, mean peak period and max gust speed variables, event duration and maximum significant wave height are the most influential variables in predicting foredune erosion at Rossbehy ($F(7, 1)=87.448$, $p<0.001$, $R^2=0.926$ for event duration and $F(6,2)=129.323$, $p<0.001$, $R^2=0.977$ for maximum significant wave height).
- The results of the multiple regression and stepwise discriminant analyses highlight the importance of event duration in affecting rates of foredune volume change at Rossbehy.
- Storm waves for all events that occurred during the monitoring period were from a narrow westerly band (257-262°).
- Prevailing winds during storm events are primarily westerly or southwesterly.

9 Investigation of sediment transport pathways at Rossbehy using a sediment tracer method

Sediment tracing is a useful technique for investigating nearshore sediment transport processes and fluxes (Allen, 1988). It is the only technique for measuring sediment transport that can be applied at a broad range of spatial and temporal scales, and, as such, has been widely used by coastal researchers and engineers (Balouin *et al.*, 2005; Schwartz, 2006). Applications of tracers in nearshore coastal environments include quantification of sediment transport in the swash and inner surf zone (*e.g.* Masselink and Russell, 2006), measurement of longshore drift, measurements of which have assisted in development of empirical longshore transport formulae (*e.g.* Komar and Inman, 1970; Komar, 1977), assessment of the sediment transport pathways near complex tidal inlets (*e.g.* Vila-Concejo *et al.*, 2004), and, more recently, the validation of numerically modeled transport (*e.g.* Cronin *et al.*, 2011).

In order to better understand the transport processes in operation at Rossbehy in the context of the overall morphological dynamics of the system, sediment tracing was employed at this study site. Two sediment tracer experiments were performed – one under high-energy (storm) conditions and another under low-energy summer conditions. This chapter will provide some background on tracer applications and methodologies in the nearshore environment and present the methods and results of the experiments performed in this study. The results of these experiments will also serve to help verify that modelled transport pathways (chapter 10) are broadly in agreement with observations.

9.1 Background

Tracers are sediment particles with some unique characteristic (*e.g.* fluorescence) that make them easily identifiable within a large mass of grains. They can thus be used to make direct observations of sediment transport. There are two general types of tracers. Natural tracers consist of mineral assemblages inherited from the characteristics of the provenance basin, while artificial tracers are particles tagged with dye or radioactivity. This research makes use of an artificial tracer. The first group of researchers known to experiment with artificial tracers on a large scale were Zenkovitch (1960) and Zenkovitch and Boldyrev (1965), who

used fluorescent dyed sand to study nearshore transport in the former Soviet Union. They injected marked sand on a beach and collected samples at fixed distances from the injection point at regular time intervals (*e.g.* minutes to days). They also experimented with injecting differently coloured samples at different beach levels to assess differential transport. Contemporaneous with Zenkovitch (1960) and Zenkovitch and Boldyrev (1965) was Abecassis *et al.* (1962), who had little success with fluorescent dyed sand due to the fact that natural luminescence was present in the sands in which they were working. They, therefore, pioneered the use of artificially tagged radioactive sands as tracers, with radioactivity levels measured from these sands using a Geiger counter. While radioactive tracers were popular at first, they are no longer widely used due to associated environmental impacts (Ciavola, 2006).

In the 1970s, the use of artificial tracers to provide direct measurements of longshore drift assisted in the development of empirical longshore transport formulae, such as those of Komar and Inman (1970) and Komar (1977). Other studies around this time made use of artificial tracers to study large-scale sediment transport patterns (Chapman and Smith, 1977) and the effect of wave motion on sediment transport (Miller and Komar, 1979). In the 1990s and 2000s, European researchers made many improvements to old methodologies. For example, with regard to sample analysis, Vila-Concejo *et al.* (2003) developed an automated method of tracer quantification whereby fluorescent particles could be detected and counted by photographing the samples under a UV light with a high-resolution digital camera.

Tracer studies continue to be widely used in the study of sediment transport. In addition, they have emerged as being of great importance for the calibration and validation of numerical models (Cronin *et al.*, 2011). A simple search on the Internet database Google Scholar reveals that the number of studies containing the words “sediment tracer” and “numerical model” almost doubled between the periods 1990-1999 and 2000-2010.

With regard to the technique itself, there are three basic assumptions associated with the use of artificial tracers. These are:

1. The tracer should have a similar hydraulic behavior to the local sediment – *e.g.* tracer particles should mimic the density and grain size of the local beach material;
2. Advection should be prevalent over diffusion and dispersion – *e.g.* the downstream transport of the tracer associated with the flow should dominate over spreading of the tracer associated with other forces; and
3. The transport system is in equilibrium over the course of the experiment – *e.g.* transport processes do not change with time.

In some cases, it may be difficult to meet these criteria. For example, the complexities of rising and falling tides, changes in wave energy and current velocities and the on- and offshore movement of sediment can hinder the application of tracer theory (Cronin, 2010). Despite such issues, for lack of a better alternative, the technique is still widely used and, although simple, it has been an invaluable tool for estimating longshore drift (Ciavola *et al.*, 1997).

At present, the use of fluorescent tracers is popular in the literature. This is because marking can be done easily and rapidly; fluorescent tracers do not pose a threat to human health or the environment (unlike radioactive tracers); and the sensitivity of the technique is high (on the order of 1 ppm) (Ingle, 1966; Ciavola, 2006). Fluorescent tracers can either be prepared by the researcher (*e.g.* by dying local sediment using various types of resins or paints) or purchased from elsewhere. Some companies (*e.g.* Environmental Tracing Systems Ltd.) manufacture and sell tracer particles to mimic the size, density and surface charge of the local sediment. This can be specified by the researcher, who may use the results of a particle size analysis for this purpose.

There are three methods of injecting a tracer into a system. These are summarised after Madsen (1987) as follows and illustrated diagrammatically in figure 9.1.

1. Time Integrated Method (TIM) – A Eulerian method whereby a known quantity of tracer is released at low tide and variations in tracer concentration are subsequently monitored over time at one location down drift of the release point.

2. Continuous Injection Method (CIM) – This method is similar to TIM, but differs in that the tracer is injected continuously at the release point at a known rate.
3. Spatial Integration Method (SIM) – A Lagrangian method whereby a known quantity of tracer is released at low tide and tracer movements are monitored in both space (*e.g.* across a grid) and time. With this method, the velocity of transport can be computed by tracking the movement of the centroid (centre of mass) of the tracer. Tracer recovery can also be estimated by extrapolating point concentrations to representative control volumes.

Choice of injection method depends on the application, although the spatial integration method (SIM) appears to be the method of choice in an overwhelming majority of longshore transport studies. Ciavola *et al.* (1998) developed a detailed methodology for quantifying longshore transport based on the SIM approach. This method is commonly cited as a guide for sediment tracer studies. Some segments from that paper that are of relevance to this study are summarised as follows.

1. To measure the mass of recovered tracer sand, Ciavola *et al.* (1998) propose the following equation:

$$Mass\ of\ recovered\ tracer\ sand = AhD\rho_s(1-p) \quad (14)$$

Where:

- A=area of which the core is representative (*e.g.* area of sample grid cell)
- h=thickness of sub-sample within core sample (they divided the cores into 5 cm thicknesses)
- D=dilution of tracer (mass of counted grains per total mass of sample)
- ρ_s=density of sand (they used 2650 kg/m^3, the density of quartz sand)
- p=sand porosity

The mass of recovered tracer sand should be calculated for each sample for each depth interval. To obtain total mass recovered per total mass injected, the sum of all masses together should be calculated and divided by the total mass injected. In their study, an estimated 90% of the

injected mass was recovered. It is important to recover a high percentage of tracer because, otherwise, estimates of transport velocities may be biased.

2. Ciavola *et al.* (1998) also provide a method for calculating the position of the tracer cloud centroid within each layer of each core sample. Calculation of this parameter is necessary for calculating transport velocity. The longshore position of the tracer cloud centroid can be obtained from the following equation:

$$Y = \frac{\sum M_i d_i}{\sum M_i} \quad (15)$$

Where:

i=samples from 1 to i
M_i=mass of tracer for sample i
d_i=distance of sample i from injection point

M_i and d_i must be calculated for each sample across the grid. Centroid positions should be calculated for each layer.

3. To calculate the velocity of transport, Y can be divided by the time between two successive low tides. Velocities should be calculated for each layer.
4. Mixing depth is defined as the depth above which 80% of the tracer is recovered.

This methodology was employed during this study to calculate the mass of recovered tracer sand, mixing depths, and velocities of sediment transport at Rossbehy.

9.2 Method and Results

For this experiment, first a particle size analysis (PSA) was carried out to identify local sediment characteristics such that a suitable tracer could be prepared. On 19 April 2013, a sediment sample was obtained mid-way between the upper and lower beach at the study site at Rossbehy (on the seaward side of the barrier at its distal end adjacent to the breach). Sediment sieving was performed and analysis of smaller particles was undertaken using a Malvern Mastersizer 2000 Laser Granulometer. Grain-size parameters were calculated in Excel after Folk and Ward (1957). The sample was found to be a fine- to

medium- well sorted sand (d50=218.3 μm, σ=0.45). A UV light helped to confirm that the sample did not display any background fluorescence, thus a fluorescent tracer was deemed suitable for this location.

Ten kg of a blue-dyed fluorescent tracer was obtained from Environmental Tracing Systems Ltd. (dry tracer shown in figure 9.2). The particles were made from barium sulphate and designed to mimic the size, density, and surface charge of the local sediment. The size fraction of the tracer was 125-350 μm, with a d50 of 225 μm.

Prior to the experiments, half pipes were prepared to preserve the stratigraphy of samples. The pipes served as receptacles inside which samples could be collected, stored and transported for later analysis. Three 6 m length PVC drain-pipes were purchased and cut into roughly 30 cm length half pipes with a jig saw. One hundred and eight half pipes were prepared.

9.2.1 December 2013 Tracer Experiment

The first tracer experiment was planned to coincide with high-energy conditions in the winter of 2013. On 9 December, Met Eireann forecast strong winds up of up to 100 km/hr (28 m/s) in the coming days, so the experiment was planned to begin on 11 December. Two and a half kilograms of tracer was deployed at low tide at two locations at approximately 5:50 am. As per the instructions of the manufacturer, the tracer was mixed in a container with seawater and an equal amount of sand from the site of deployment, then raked into the surface sediment layer over an area of approximately 1 m by 1 m (figure 9.4). Two kg of tracer was deployed approximately 150 m northeast of the foredune terminus at an elevation of 2.91 m ODM and 0.5 kg of tracer was deployed approximately 90 m seaward of the dune toe near the barrier terminus at an elevation of 2.50 m ODM (red "x"s in figure 9.4). The two locations were strategically chosen to assess whether or not sediment was being transported alongshore and into the breach. Also, the elevations of injection sites guaranteed that they would be inundated during part of the tidal cycle, with approximate inundation periods of 4 hours (2.91 m ODM injection site) and 6 hours (2.50 m ODM injection site). The GPS coordinates of the injection sites were recorded with a Trimble ProXH differential GPS.

Sampling took place one tidal cycle (approximately 10-12 hours) later, from 4-6 pm. The planned sampling strategy was Lagranigan and modelled after strategies described by Cronin (2010). In her study on sediment transport in estuarine tidal flats, Cronin (2010) used four sampling strategies. These are summarised as follows:

- sample along intersecting lines in the main compass point directions (N, S, E, W) around the point of injection at regular (or semi-regular) intervals up to a maximum distance (up to 12 m)
- sample in an alongshore direction on either side of the point of injection
- sample in a regular or semi-regular grid around the point of injection (up to a maximum distance of 15 m)
- sample in concentric circles around the point of injection (up to a maximum distance of 40 m)

The different strategies were employed based on the observed movement of the tracer. For example, in one experiment in which sampling was performed over a semi-regular grid, the dominant direction of tracer movement was found to be to the west and the south. As a result, Cronin (2010) added a south westerly-component to the sampling grid.

For this study, the first samples were collected after the first tidal cycle from 16:00 to 18:00 on 11 December (approximately 10 hours after tracer injection). No tracer could be positively identified in the field around either point of injection, even with a UV light. Similar to the first strategy mentioned previously, samples were collected along intersecting lines around the site of the 2 kg deployment. The lines were orientated in an alongshore and cross-shore direction, rather than in the main compass point directions, in an effort to assess the alongshore and cross-shore transport components. For the 0.5 kg injection site, samples were collected in a shore-parallel line seaward of the 0.5 kg deployment (blue dots in figure 9.4).

Sampling consisted of plunging the half pipes into the sand as deep as they could go, and pulling them up with the aid of the trowel (figure 9.5). The excess sediment was then shaved off the top across the diameter of the pipe and the

samples were preserved in cling film. The GPS coordinates of each sample were recorded. Fifteen samples were obtained with depths of approximately 20 cm. During sampling, weather conditions were deteriorating and gusts were picking up. This, unfortunately, resulted in the accidental deployment of additional tracer when a container containing a backup 2 kg of tracer was blown over. In an effort to try and salvage the experiment, the tracer was quickly emptied onto this site and raked into the sediment. The GPS coordinates of the third injection site were recorded (elevation = 1.468 m ODM) and its location is shown in figure 9.4 (yellow "x"). No further samples were obtained after injection on that visit.

On the second day, after the third tidal cycle from the first injection, sampling took place at low tide. Twenty one samples were collected, and their locations are shown in figure 9.4 (pink dots). No tracer was positively identified in the field around any of the points of injection.

Six days after the first injection (twelve tidal cycles later), one final sampling campaign took place. Twenty samples were obtained, and their locations are shown in figure 9.4 (green dots). Again, no tracer was positively identified in the field. Figure 9.6 summarises the times of injection and sampling in relation to the tidal cycle over the duration of the experiment. Information about wind speed and direction during the experiment was also obtained (from the Ventry weather station data) and is shown in figure 9.7. Winds were predominantly offshore, with average speeds of 5.5 m/s. Offshore winds enhance the development of plunging breakers (Galloway, 1989), which are known to be associated with increased rates of longshore transport (Wang *et al*., 2002).

The samples were transported back to UCC and stored in a refrigerated unit until they were ready for analysis so that they would remain moist and intact. Samples were analysed in 1.5 cm layers, whereby each layer was carefully removed, broken, and sifted through (figure 9.8) with a blunt knife. With the aid of a UV light and a magnifying glass, the presence and number of individual tracer particles was noted for each 1.5 cm layer in each of the 56 samples. This method was both tedious and laborious, but given the extremely low concentration of tracer recovered in the samples, a photographic method of estimating tracer concentration was regarded as inappropriate. Figure 9.9 shows

the presence of two tracer particles under both ordinary and UV light in a core sample and illustrates the difficulty associated with identifying such particles in low quantities.

A total of 60 individual tracer particles could be positively identified from the first set of samples (those collected after the first tidal cycle on 11 December). Figure 9.10 shows the distribution of samples containing tracer and the number of particles per sample collected at that time. Most samples containing tracer were concentrated around the site of the 2 kg deployment, although no clear direction of transport could be inferred. Table 9.1 illustrates the number of particles found in each depth layer for each sample. Tracer was found to a depth of 21 cm, although more than 80% of the particles were found above 9 cm. The estimated percent recovery was calculated using the method of Ciavola *et al.* (1998). For each sample at each depth interval, the mass of recovered tracer sand was calculated using equation 14, with:

- A=area of which the core is representative (*e.g.* area of sample grid cell)=5 m^2
- h=thickness of sub-sample within core sample=0.015 m
- D=dilution of tracer (mass of counted grains per total mass of sample) – the mass of the counted grains was estimated by multiplying the number of particles by the average mass of a grain of fine to medium sand, taken to be 0.0045 g. This was then subtracted from the total sample weight and the ratio between the two masses was calculated.
- ρ_s=density of sand (taken as 2650 kg/m^3, the density of quartz sand)
- p=sand porosity (0.6 for wet sand)

To obtain total mass recovered per total mass injected, the sum of all masses together were calculated and divided by the total mass injected. For the 11 December samples, the estimated percent recovery was only 8.25%.

Only nine individual tracer particles were positively identified from the second set of samples (those collected after the third tidal cycle on 12 December). Figure 9.11 shows the distribution of samples containing tracer and the number of particles per sample collected at that time. Tracer was found only around the first and second injection points, but again, no clear direction of transport was obvious. Table 9.2 illustrates the number of particles found in each depth layer for each sample. Tracer was found to a depth of 15 cm, with more than 80% of

the particles found above 10.5 cm. The estimated percent recovery was only 0.69%.

No tracer was found to be present in any of the third set of samples, which were collected after the 12th tidal cycle from the first injection on 17 December.

Unfortunately the collective result of this experiment was that there was not enough information to make any meaningful observations about the rate or direction of transport at the site, except the qualitative observation that, perhaps, sediment may have been moving too quickly or have travelled much greater distances from the point of injection for transport to be observed in a single tidal cycle. As such, no further analysis was performed on these data and a second experiment was planned, but this time under less energetic conditions.

9.2.2 June 2014 Tracer Experiment

A second sediment tracer experiment took place in June 2014. This time, the entire remainder of the tracer, 5.5 kg, was deployed at a single location at low tide on 17 June 2014 (figure 9.12). The tracer was deployed approximately 75 m seaward of the dune toe at an elevation of 2.158 m ODM, ensuring inundation for approximately 7 hours. The first sampling campaign took place after the first tidal cycle. In an attempt, at a minimum, to identify maximum distance travelled, sampling took place at a range of locations within the breach. However, it eventually became clear that much of the tracer was spread at distances of up to 15-20 m around the point of injection. As such, to try and maximise tracer recovery, further sampling took place only where tracer could be positively identified in the field, either on the surface or within a core sample. A total of 17 samples were collected after the first tidal cycle (blue dots in figure 9.12); 21 samples were collected after the second tidal cycle (green dots in figure 9.12); and 19 samples were collected after the third tidal cycle (orange dots in figure 9.12). Some additional samples were collected in the field, but if, on inspection, no tracer particles were observed to be present across the diameter of the core, its GPS coordinate was recorded (as "no tracer") and the sample discarded. Figure 9.13 summarises the times of injection and sampling in relation to the tidal cycle over the duration of the experiment. Average wind speed during the experiment

(8.6 m/s) was higher than that during the December 2013 experiment and winds were predominantly southeasterly (figure 9.14).

Like with the December experiment, the samples were stored in a refrigerated store until they were ready for analysis. The same method of analysis was employed, except this time samples were analysed in 2 cm layers to speed up the process. Tracer was present in many of the cores in substantial quantities. Figure 9.15 shows the presence of many tracer particles relatively evenly distributed within one of the core samples. For layers where more than 100 particles were present, counting was impractical. The sample was therefore evenly divided into subsets, the number of particles in one of the subsets was counted, and this figure was multiplied by the total number of subsets. While this method assumes an even distribution of particles within the layer, qualitative observations of particle distribution within the cores confirmed this assumption.

Tracer distribution after the first tidal cycle

The estimated number of tracer particles positively identified from the first set of samples (those collected after the first tidal cycle at mid-day on 17 June) was 5927, and the estimated percent recovery was 74%. Figure 9.16 shows the distribution of samples containing tracer and the number of particles per sample collected. Most of the tracer was concentrated within 10 m of the injection point. A single particle was found in the top 4 cm of a core located approximately 240 m from the point of injection, although this is insufficient evidence to confidently state that the tracer travelled this distance from the point of injection.

Table 9.3 illustrates the number of particles found in each depth layer for each sample. Maps showing the locations of samples (labelled with corresponding sample IDs) are shown in figure 9.17. Tracer was found to a depth of 12 cm. Eighty percent of the particles were located at depths between 0-6 cm.

The cores were divided into three equally sized layers between 0-12 cm to further investigate tracer distribution within the mixing zone. Figures 9.18, 9.19 and 9.20 show tracer distribution for each of the layers 0-4 cm, 4-8 cm, and 8-12 cm, respectively. At all levels, the tracer centroid appears to be moving alongshore, but the absence of samples that would confirm that no tracer was

present in other directions means that this cannot be confirmed. The priority in the field at the time of sampling was to make the most of the time spent collecting samples where there actually *was* tracer present. It was not possible to anticipate where tracer would or would not be found if it was buried. Nonetheless, these results appear to suggest that the tracer centroid moved alongshore following injection and through the first tidal cycle.

The longshore position of the tracer cloud centroid and velocity of transport was calculated for each layer using the method of Ciavola *et al.* (1998) (see equation 15). This was computed as follows:

1. For each sample, the mass of recovered tracer particles in the 0-4 cm, 4-8 cm, and 8-12 cm depths (M_i in equation 15) was estimated by multiplying the total number of recovered particles by the mass of a grain of fine to medium sand (0.0045 g).
2. The distance between the injection point and the samples was measured in ArcGIS (d_i in equation 15) and recorded.
3. The estimated masses calculated in step 1 were multiplied by the distance from the injection point measured in step 2.
4. The tracer centroid (Y in equation 15) for each layer was calculated by taking the sum of the $M_i d_i$ term (calculated in step 3) and dividing it by the sum of the masses of recovered tracer particles estimated in step 1.

The velocity of transport for each layer was calculated by dividing Y by the time between two successive low tides.

Table 9.4 summarises the results of these calculations. The longshore position of the tracer cloud centroid decreased from 4.6 m (top 0-4 cm) to 4 m (4-12 m) with depth, and the velocity of transport decreased by an order of magnitude with depth.

Tracer distribution after the second tidal cycle

The estimated number of tracer particles positively identified from the second set of samples (those collected after the second tidal cycle at midnight on 17/18 June) was 7169, and the estimated percent recovery was 90%. Figure 9.21 shows the distribution of samples containing tracer and the number of particles

per sample collected. Most of the tracer was concentrated within around 15 m of the injection point.

Table 9.4 illustrates the number of particles found in each depth layer for each sample. Maps showing the locations of samples (labelled with corresponding sample IDs) are shown in figure 9.22. Tracer was found to a depth of 16 cm. Eighty percent of the particles were located at the surface, at depths between 0-2 cm.

Again, the cores were divided into three equally sized layers between 0-12 cm to further investigate tracer distribution within the mixing zone. Figures 9.23, 9.24 and 9.25 show tracer distribution for each of the layers 0-4 cm, 4-8 cm, and 8-12 cm, respectively. In the top 0-4 cm layer, the tracer centroid appeared to be moving in a westerly direction, oblique to the shoreline, but still generally alongshore. At 4-12 cm depths, the centroid appears to have moved more or less parallel with the shoreline. The absence of tracer in samples surrounding the tracer centroid in the 8-12 cm layer lends credence to the inference that transport is alongshore, but does not necessarily confirm this.

The position of the tracer centroid and velocity of tracer movement was calculated for the samples using the method described previously. The distance between the centroid after the second tidal cycle (t_2) and the centroid after the first tidal cycle (t_1) was calculated by subtracting the distance between the centroid and the injection point at t_2 from the distance between the centroid and the injection point at t_1. The velocity between t_2 and t_1 was calculated by dividing this distance by one tidal cycle.

Table 9.5 summarises the results of these calculations. The longshore position of the tracer cloud centroid decreased from 6.1 m (top 0-4 cm) to 5.4 m (4-12 m) with depth, thus the velocity of transport was again highest at the surface.

Tracer distribution after the third tidal cycle

The estimated number of tracer particles positively identified from the third set of samples (those collected after the third tidal cycle at midday on 18 June) was 941. The estimated percent recovery was only 12%. Figure 9.26 shows the distribution of samples containing tracer and the number of particles per sample

collected. This time, the tracer cloud was more widely distributed, with most of the tracer concentrated within around 50 m of the injection point.

Table 9.6 illustrates the number of particles found in each depth layer for each sample. Maps showing the locations of samples (labelled with corresponding sample IDs) are shown in figure 9.27. Tracer was found to a depth of 14 cm. Eighty percent of the particles were located between 0-4 cm.

Figures 9.28, 9.29 and 9.30 show tracer distribution for each of the layers 0-4 cm, 4-8 cm, and 8-12 cm, respectively. The pattern of tracer distribution within these samples provides the most compelling evidence for the dominance of alongshore transport. At all levels, the tracer centroid appeared to move alongshore, albeit at an oblique angle to the point of injection (in an onshore direction).

Again, the position of the tracer centroid and velocity of tracer movement was calculated for the samples using the method described previously. The distance between the centroid after the third tidal cycle (t_3) and the centroid after the second tidal cycle (t_2) was calculated by subtracting the distance between the centroid and the injection point at t_3 from the distance between the centroid and the injection point at t_2. The velocity between t_3 and t_2 was calculated by dividing this distance by one tidal cycle.

The longshore position of the tracer cloud centroid was found to be furthest from the injection point (36.8 m) at the lowest depths (8-12 cm) and decreased upwards (table 9.7). It should be noted, though, that the low percentage of tracer recovered after the third tidal cycle means these results should be treated with caution.

Summary

- The estimated percent recovery of tracer[16] after each tidal cycle is summarised as follows:
    - After 1st tidal cycle = 74%
    - After 2nd tidal cycle = 90%

[16] Percent recovery is extrapolated per m^2.

- After 3rd tidal cycle = 12%

- Mixing depths calculated based on samples collected after each tidal cycle are summarised as follows:
  - Samples collected after 1st tidal cycle = 0-6 cm
  - Samples collected after 2nd tidal cycle = 0-2 cm
  - Samples collected after 3rd tidal cycle = 0-4 cm
- Table 9.8 summarises the longshore position of tracer cloud centroids and velocities of transport for subsample layers 0-4 cm, 4-8 cm, and 8-12 cm from samples collected after each of the three tidal cycles.
- The observed direction of sediment transport is likely to be predominantly alongshore, but more evidence is required to confirm this.

10 Process-based modelling of the impacts of storms under SLR

A primary objective of this research was to evaluate the importance of storms as a driver of morphologic change at the Inch-Rossbehy barrier system under potential future sea-levels. To address this objective, numerical modelling was employed. Numerical models are the only tools that allow a quantitative projection into the future (Kraus, 1996). While their use as predictors of coastal behaviour has been subject to some degree of criticism (*e.g.* Thieler *et al.*, 2000; Pilkey and Pilkey-Jarvis, 2007), a thoughtful, well designed experiment can yield instructive insights into the dynamics of the coastal environment.

In this study, scenario testing was employed to gain insight into the relative impacts of storms under differing sea-levels. The Intergovernmental Panel on Climate Change have described scenarios (in numerical modelling terms) as follows:

> "Scenarios are images of the future, or alternative futures. They are neither predictions nor forecasts. Rather, each scenario is one alternative image of how the future might unfold…Scenarios help in the assessment of future developments in complex systems that are either inherently unpredictable, or that have high scientific uncertainties…Good scenarios are challenging and court controversy, since not everybody is comfortable with every scenario, but used intelligently they allow policies and strategies to be designed in a more robust way" (Nakicenovic and Swart, 2000).

Although written in relation to GHG emissions scenarios, this description is relevant to this research in that the Inch/Rossbehy barrier system is (morphodynamically) highly complex. Given the almost infinite number of possible future morphological realisations (in response to future storms, sea-level change, human activities, etc.), it can be argued that the future of this system is unpredictable at worst or highly uncertain at best. As such, an experiment was designed to isolate the effects of storms under varying sea-levels, whereby all model inputs were held constant except that of sea-level. This chapter first describes the model used to run the experiment, MIKE21, and the specific set-up for the Inch/Rossbehy barrier system, which was developed by members of

UCC's (former) Hydraulics and Maritime Research Centre (now part of MaREI). It then goes on to describe the model scenarios and inputs used in the experiment. While the scenarios were run using the HMRC/MaREI setup, the experimental simulations were designed by myself, the inputs for the experimental simulations were prepared by myself, and all analysis of the modelled output were performed by myself explicitly for this PhD thesis.

Model results for all simulations are also presented in this chapter. These include include:

- Bed level changes over the course of typical and extreme storm event simulations under three different sea-level scenarios
- Changes in transport magnitude and direction during typical and extreme storm event simulations under different sea-level scenarios
- Comparison of the above results with those of a control simulation ("fair-weather")
- Comparison of modelled morphologic change with observations from the TLS monitoring campaign

Analysis of model results is limited to the nearshore area around Rossbehy.

10.1 Model Description

For this research, numerical modelling was undertaken using the MIKE21 suite of software, developed by DHI. MIKE21 is a 2D numerical model that simulates flows, waves, sediments and ecology in rivers, lakes, estuaries, bays, coastal areas and seas. It is one of the most commonly used coastal modelling software suites. Because it is 2D, is should be applied only where stratification of the water column can be neglected.

Three of the MIKE21 modules were used in this research:

- MIKE21 Hydrodynamics (HD) - simulates water level variations and flows (*e.g.* tidal currents) in response to a variety of forcing functions.
- MIKE21 Spectral Waves (SW) – simulates the growth, decay and transformation of wind-generated waves and swell.

- MIKE21 Non-Cohesive Sediment Transport Module (ST) – simulates non-cohesive sediment transport due to currents or combined waves and currents.

Various physical processes are included in these modules, including:

- Wave growth by action of wind
- Non-linear wave-wave interaction
- Dissipation due to white-capping
- Dissipation due to bottom friction
- Dissipation due to depth-induced wave breaking
- Refraction and shoaling due to depth variations
- Wave-current interaction
- Effect of time-varying water depth and flooding and drying

The MIKE21 HD and ST modules are based on a flexible mesh – a triangular grid whose resolution can vary across the model domain. This provides some degree of flexibility in the representation of complex geometries and enables smooth representations of boundaries. Equations are solved across the mesh using a cell-centred finite volume solution technique, whereby the spatial domain is discretised by subdivision of the continuum into non-overlapping elements or cells (see DHI Software, 2013).

Basic required model inputs include:

- Model domain (extent of model area) and grid
- Digitised bathymetry
- Time step
- Duration of simulation
- Boundary conditions (*e.g.* wave parameters, surface elevation or flux at all open boundaries, sediment properties, etc.)

For modelling storms, winds and surge heights can also be incorporated. For more detailed model descriptions, see the MIKE21 Flow Model Hydrodynamic Module User Guide (DHI Software, 2007a), the MIKE21 SW Spectral Waves

FM Module User Guide (DHI Software, 2007c) and the MIKE21 ST Non-Cohesive Sediment Transport Module User Guide (DHI Software, 2007b).

As with all numerical models, the quality of the model results provided by MIKE21 is limited by the quality of the model inputs, especially the model bathymetry, as well as other user specified information, such as flow boundary conditions, eddy viscosity, and bottom roughness values. An inherent limitation of numerical models is that they are required to make some form of simplifying approximations in order to solve the governing equations (Toombes and Chanson, 2011). Some of the most important simplifying approximations always or often employed by MIKE21 include:

- Solution of the flow field in 2 dimensions,
- Assumption of depth-average properties,
- Omission of fluid properties that are assumed to have a negligible influence on flows (*e.g.* constant density and temperature; no inclusion of viscosity or surface tension), and
- The use of empirical formulae (*e.g.* Manning's equation) to approximate flow characteristics.

10.2 Model set-up

The Dingle Bay model was set up in MIKE21 by members of UCC's Hydraulics and Maritime Research Centre (HRMC), including Michael O'Shea and Jimmy Murphy, as part of a PhD thesis on predicting the medium-term (to 2030) evolution of inner Dingle Bay (O'Shea, 2015). The same model set-up was used to simulate the nearshore wave data used to identify storm events in chapter 7. An overview of the model domain and flexible mesh are shown in figure 8.1 (chapter 8) and the nearshore mesh is shown in more detail in figure 10.1. Bathymetry data for the model domain were sourced from the following:

- The INtegrated Mapping FOr the Sustainable Development of Ireland's MArine Resource (INFOMAR) seabed survey data for Dingle Bay (10 m resolution) – covers the majority of the bay (figure 10.2)

- Aerial LiDAR data – obtained in April 2011 by Kerry County Council (2 m resolution) – covers the nearshore area around Inch, Rossbehy, and Cromane to a depth of approximately 10 m ODM (figure 10.3)
- Admiralty charts – various sources
- Nearshore bathymetry surveys carried out by Michael O'Shea in March and September 2013 – cover only a small area seaward of the breach at Rossbehy

The INFOMAR, aerial LiDAR data, and admiralty chart data were sourced by Sala (2010). Bathymetric data was interpolated to a mesh created in the MIKE21 Mesh generator (figure 10.1). Relevant land boundaries were also imported. The initial bathymetry used by O'Shea (2015), which reflects a breached barrier configuration, was used in this research. Meshing and interpolation was initially carried out by Sala (2010) and later updated by O'Shea (2015) to include higher resolution nearshore bathymetry data.

Model validation was performed by O'Shea and Murphy (2013) and O'Shea (2015). While past studies indicate MIKE21 is generally capable of modeling flows and transports at tidal inlets reasonably well (*e.g.* Siegle *et al.*, 2004 and Sennes *et al*., 2007), the sites under investigation in these previous studies do not share similar characteristics to the Inch-Rossbehy system. These systems are often dominated by longshore transport, whereas at Inch and Rossbehy, longshore sediment transport to the barriers is limited by their unusual configuration within the long, narrow bay. O'Shea and Murphy (2013) and O'Shea (2015), however, argue that the MIKE21 model set-up is capable of modelling alongshore and cross-shore transports at Rossbehy using the cross shore transport formula of van Rijn (1998) and the alongshore transport formula of van Rijn (2009). They compared dune recession modelled using these formulae (and others) with measured dune recession from a seasonal monitoring campaign carried out between July 2009 and October 2011. Their results show some agreement between modelled and observed dune recession (tables 10.1 and 10.2). In the swash-aligned zone, the difference between the calculated range of recession using the cross-shore formula and the measured range was <1 m. In the drift-aligned zone, the alongshore formula was found to be more appropriate,

although it tended to over predict recession, especially on the newly formed island at the distal end of the barrier (O'Shea and Murphy, 2013). Further comparison of modelled dune recession and observations from the TLS monitoring campaign performed in this PhD research is described later in this chapter.

Nine simulations were planned for this experiment. Observations of recent morphodynamic behaviour at Rossbehy from the TLS monitoring campaign suggested non-extreme storms could be just as destructive as extreme events. As such, two scenarios were developed – one which would represent a "typical" storm and one which would represent an "extreme" storm. Given that the purpose of the experiment was to investigate the impacts of storms under varying sea-levels, three sea-levels were chosen under which each of the scenarios would be run – 0 m (the baseline), 10 cm (a bottom-of-the-range 21st century GMSLR forecast), and 50 cm (a middle-of-the-range 21st century GMSLR forecast). The choice of these specific figures was guided by the UKCP09 projections (see chapter 4), which suggest that under the IPCC AR4 medium emissions scenario, the southwest coast of Ireland can expect a relative SLR of roughly 45-50 cm by the year 2095. A control scenario, whereby non-storm conditions would occur, would also be run for comparison. All scenarios were run using the same initial bathymetry and for the same period of time so that any modelled changes in the resulting morphologies could more likely be attributed to SLR and not to differences in initial bathymetry or changes in event duration.

Sediment transport by wind was not modeled. Given the short periods over which the scenarios were run, it was assumed aeolian transport would be negligible. O'Shea (2015) reported that rates of accretion in a sand trap placed near the dune terminus at Rossbehy were 0.13 m/month (based on one year's worth of data). This is well below the ability of the model to resolve over the timescale used in this study (53 hours).

For each scenario, a single simulation was performed. Given the probabilistic nature of the model, it should be noted that multiple simulations using the same inputs may yield different results. Further research would be required to examine internal variation in sediment mass changes.

The following section describes how the model inputs for each of the storm and non-storm scenarios were derived.

10.2.1 Model inputs and scenarios

In an effort to simulate realistic conditions, the non-storm and storm scenario inputs were derived from hindcast nearshore wave data and real local weather station data. Storm events and their characteristics over the period 2011-2014 had been previously identified from simulated nearshore wave data (see chapter 8). From these data, characteristics associated with all events characterized by modal wave conditions over this period were identified. These included mean event duration, mean wave direction, and peak period and are given in table 10.3. An event with characteristics similar to these was identified from the pool of events. Its characteristics are shown in table 10.4. Additional model inputs were derived for this event from the Ventry weather station data. Half-hourly time series of wind speeds, directions, and sea-level pressures were extracted from the data for the event, which ran from 2012-01-24 18:00:00 to 2012-01-26 23:00:00. Sea-level pressure was used as a proxy for the storm surge, with the surge height inferred to increase by 1 cm for every mb below 1015 mb. For simplicity, the influence of funneling of the surge into Dingle Bay and geostrophic events were not considered. As a result, surge levels may be underestimated in the typical and extreme events scenarios.

Figures 10.4, 10.5, and 10.6 show the wind speeds, wind directions, and surge heights, respectively, used to drive the "typical event" scenario. These data were prepared by myself and provided to the HMRC in Excel spreadsheets.

Identifying a representative "extreme event" from the pool of data was problematic because sometimes storms would merge into one another and a series of events was therefore regarded as a single storm, as discussed in chapter 6. For example, during the winter 2013/2014, which is regarded as one of the stormiest winters in Ireland on record (Matthews *et al.*, 2014; Met Eireann, 2014; National Directorate for Fire and Emergency Management, 2014), a series of back-to-back events in December/January were recorded as a single 633-hour event. This "event" appeared to be a good candidate for representing the extreme scenario, as it epitomized the impacts of back-to-back events and the

characteristics of some of the (sub) events that occurred during this period were comparable to the most extreme events ever recorded in Ireland (examples to follow later in this section). The characteristics of this long duration event, as extracted from the WAM data, are summarized in table 10.5. Required model inputs (wind speeds, wind directions, and surge heights) for this event were derived from the Ventry weather station data and provided to the HMRC for the "extreme event" simulation. However, several attempts to run this scenario failed, most likely because the lengthy duration of the event caused the simulation to be too computationally expensive. It was therefore later decided that a subset of the 633-hour event should be selected, over which the most extreme conditions occurred. The chosen subset spanned a 53-hour duration, which made it more reasonable to compare the results with those of the typical event (which was also run for a 53-hour duration). To decide on which subset to extract, wind speeds and surge heights were further analysed for the 633-hour duration event. Maximum wind speeds during the event occurred on 23/24 December (up to 14.4 m/s) and on 26 December (up to 20.3 m/s). Maximum surges occurred on 24 December (49.4 cm) and 27 December (49.8 cm). Two 53-hour periods centred around these times were selected as candidates for the extreme event scenario. To see how the two events compared with each other and other events, they were plotted against data compiled by Orford *et al.* (1999) on the most extreme events recorded in Ireland. Figure 10.7 shows severe storm power in terms of minimum pressure and wind speed for extreme events that have affected Ireland in the 19th and 20th centuries (blue) and for the 23/24 December 2013 and 26/27 December 2013 events (red). While both events fall at the lower end of the extreme spectrum, the 26/27 December event was more extreme in terms of wind speeds. This event, which ran from 2013-12-26 11:30:00 to 2013-12-28 16:30:00, was therefore chosen for the extreme storm simulation. Figures 10.8, 10.9, and 10.10 show the wind speeds, wind directions, and surge heights, respectively, used to drive the "extreme event" scenario. These data were provided to the HMRC in Excel spreadsheets.

For the control scenario ("fair-weather conditions"), typical wind speeds and directions for the times outside of which events occurred (relatively calm periods) were identified from the Ventry data such that a suitably representative

"non-event" period could be chosen. This would therefore reflect the natural variability in meteorological conditions. The mean wind speed for periods outside of which events occurred was 4.31 m/s with a standard deviation of 2.85 m/s. Southerly winds were found to be predominant. A 53-hour period was chosen from the non-event data which was representative of these conditions. This period lasted from 2012-02-01 00:00:00 to 2012-02-03 05:00:00, had a mean wind speed of 4.58 m/s with a standard deviation of 2.64, and was characterized by predominantly southerly winds. Figures 10.11 and 10.12 show the wind speeds and directions used to drive the control scenario. These data were provided to the HMRC in Excel spreadsheets. No surge was superimposed on this scenario.

To summarise, based on the described derived model inputs, three "event" scenarios were defined. These are summarised in table 10.6. All events were run for a duration of 53 hours – the mean duration of events extracted from the simulated nearshore wave data – with varying wind speeds, directions, and surges based on real measurements. Each of the three event scenarios was run under sea-levels of 0 cm, 10 cm, and 50 cm.

The model was run for each of the nine scenarios and the output provided in the form of .dfsu files, which could be read by MIKE Zero (the MIKE21 post-processing module). Sediment and wave output were recorded at 2 hour intervals, while the flow output was recorded at 1 hour intervals.

10.3 Analysis of model outputs

Modelled data were exported from MIKE21 such that they could be imported to and analysed with GIS and in Excel. Exported data included:

- Initial bathymetry (the same for all 9 scenarios)
- Final bathymetry for all 9 scenarios
- Time series of sediment transport magnitude and direction at a point near the breach in the intertidal zone (specifically, the same location as the June 2014 sediment tracer experiment injection point)
- Time series of water levels at that coordinate

- Time series of water levels at a coordinate in the nearshore (sub-tidal) zone seaward of the breach

Bed level data (vertical datum = LAT) were imported to ArcGIS (v. 10.2) in the form of triangular irregular networks (TINs) and converted into raster DEMs representing the intial bathymetry and the final bathymetries. DEMs of difference (DODs) were generated by subtracting the final raster DEM (t_2) from the initial raster DEM (t_1) for each scenario. These DODs provided a good indication of where and to what extent erosion and deposition took place.

Volume change maps were also generated using the cut and fill tool in ArcGIS. To quantitatively assess volume change in the nearshore zone, vector polygons were generated from the initial raster DEM, within which volume changes were assessed. These polygons represented the upper (>0 m depth contour) and lower (-5 to 0 m depth contour) shore. The polygon representing the area between the -5 and 0 m contour was arbitrarily bounded approximately 3.5 km west of the barrier and 2 km east of the barrier. These polygons were used as a way of quantitatively comparing volume change on the lower and upper shore across each of the scenarios over the same area and were also helpful in the qualitative assessment of bed level change on the elevation change maps.

Changes in transport magnitude and direction near the breach were assessed in the intertidal area near the breach at the same coordinate as the injection point of the June 2014 sediment tracer experiment (UTM coordinate 433532.807, 5770466.742 – shown in figure 10.13). Time series of transport magnitude (bed load) and direction, and water levels were extracted at this point for each scenario. Because that coordinate was often supratidal, a time series of water levels at a coordinate further offshore (UTM coordinate 431732.69, 5770864.57 – shown in figure 10.14) was also extracted to give a full picture of the tidal state throughout each of the scenarios. An aerial view of the area covered by the maps presented in the following section relative to the 0 m and -5 m depth contours and the site of the sediment tracer experiment is shown in figure 10.15.

10.3.1 Results: Bed level changes in the nearshore zone

Fair-weather conditions – 0 m SLR

Figure 10.16 shows bed level changes near Rossbehy for the fair-weather conditions - 0 m SLR scenario. Overall magnitudes of erosion and deposition were lower than for the typical event scenario (described in the ensuing section). For example, bed level change ranged from -0.72 m to 0.44 m for the typical event – 0 m SLR scenario as opposed to -0.14 m to 0.19 m for the fair-weather conditions – 0 m SLR scenario. High levels of erosion were concentrated along the length of the barrier on its seaward side (above and straddling the 0 m contour) from its proximal end to the southern margin of the breach. Erosion here was not as extensive as it was for the typical event scenario and did not extend as far into the dunes as for the typical event scenario. The breach was characterised more by deposition than erosion. Similar to the typical event scenario, high levels of erosion could be observed in pockets approx. 750 m due east of the breach, on the seaward margin of the ebb shoal located seaward of the barrier, and on the steep margins of the main inlet channel.

Figure 10.17 shows volume gains and losses for this scenario, which generally reflect the elevation change map. Net volume change above the 0 m contour was positive (+4,397 m^3). This was the only simulation of all nine in which net volume change above the 0 m contour was positive. Net volume change between the -5 m and 0 m contours was also positive (+498 m^3). As such, the overall net volume change above the -5 m contour was positive (+4,895 m^3), indicating an overall onshore movement of sediment. This was one of only two simulations out of all nine in which net volume change above the -5 m contour was not negative (the other was the fair-weather conditions, 0.5 m SLR simulation).

Fair-weather – 0.1 m SLR

Figures 10.18 and 10.19, respectively, show bed level changes and volume losses and gains, respectively, near Rossbehy for the fair-weather – 0.1 m SLR scenario. Patterns of erosion and deposition are similar to those under the 0 m SLR scenario. Differences between the two scenarios are more apparent upon examination of net volume changes. Net volume change above the 0 m contour

was slightly negative (-724 m^3, compared to +4,397 m^3 for the 0 m SLR scenario) and net volume change between the 0 to -5 m contour was slightly positive (+463 m^3, compared to +498 m^3 for the 0 m SLR scenario). The overall net volume change above the -5 m contour was -261 m^3 (compared with +4,895 m^3 for the fair-weather – 0 m SLR scenario).

Fair-weather – 0.5 m SLR

Figures 10.20 and 10.21 show bed level changes and volume losses and gains, respectively, near Rossbehy for the fair-weather – 0.5 m SLR scenario. Again, patterns of erosion and deposition are similar to those under the 0 m and 0.1 m SLR scenarios. It is therefore more instructive to assess volume changes, compared to the 0 m and 0.1 m SLR scenarios. Figure 10.22 summarises volume change both above the 0 m bathymetric contour (top) and between the -5 to 0 m bathymetric contours (bottom) for each of the 3 fair-weather SLR scenarios. Net volume change above the 0 m contour for the 0.5 m SLR scenario was negative (-1,015 m^3) and net volume change between the -5 m and 0 m contours was positive (+1,127 m^3). Overall net volume change above the -5 m contour was therefore slightly positive (112 m^3). These results indicate that for the fair-weather scenarios, under higher sea-levels less sediment is deposited in the nearshore zone (above the -5 m bathymetric contour).

Typical Event – 0 m SLR

Figure 10.23 shows bed level changes near Rossbehy for the typical event - 0 m SLR scenario. High levels of erosion were concentrated in the following areas:

- along the length of the barrier on its seaward side (above the 0 m contour) from its proximal end to the breach (adjacent to, and possibly affecting, the foredunes),
- in small pockets approx. 750 m due east of the barrier neck/breach area,
- on the seaward margin of the ebb shoal located seaward of the barrier, and
- on the steep margins of the main inlet channel.

High levels of deposition were generally concentrated near areas where high levels of erosion occurred, specifically:

- seaward of and adjacent to the 0 m contour,
- between pockets of erosion approx. 750 m due east of the breach,
- on the upper part of the ebb shoal seaward of the barrier, and
- interspersed between areas of erosion on the steep margins of the main inlet channel.

During the TLS monitoring campaign, a similar pattern emerged near the breach, whereby following erosive storms, heightened beach levels accompanied foredune erosion.

Figure 10.24 shows volume gains and losses for this scenario, which broadly reflect the elevation change map. Net volume change above the 0 m contour was negative (-12,435 m^3). Between the -5 to 0 m contour, net volume change was slightly positive (+3,328 m^3), but not sufficient to offset the losses suffered above the 0 m contour. Net volume change above the -5 m contour was therefore negative (-9,107 m^3), indicating an overall offshore movement of sediment.

Modelled dune volume change in the area where the TLS monitoring campaign was conducted (adjacent to the breach) was consistent with observations. During the period 19 April to 5 June 2013, two events occurred with similar characteristics to the "typical" event. Dune volume change observed during this period at the TLS monitoring site was -0.07 m^3/m^2. The modelled dune volume change across the same area (for the typical event – 0 m SLR scenario) was -0.05 m^3/m^2. While this does not necessarily validate the model (the TLS monitoring campaign took place over a very small area compared to the model domain), it suggests the model may be able to reasonably predict dune erosion volumes in the area adjacent to the breach during such events.

Typical Event – 0.1 m SLR

Figures 10.25 and 10.26 show bed level changes and volume losses and gains, respectively, near Rossbehy for the typical event – 0.1 m SLR scenario. Patterns

of erosion and deposition are similar to those under the 0 m SLR scenario. Differences between the two scenarios are more apparent upon examination of net volume changes. Like with the 0 m SLR scenario, net volume change above the 0 m contour was negative (-13,615 m^3, compared to -12,435 m^3 for the 0 m SLR scenario) and net volume change between the 0 to -5 m contour was positive (4,010 m^3, compared to 3,328 m^3 for the 0 m SLR scenario). Again, overall losses did not offset gains, resulting in a net volume loss above the -5 m contour of -9,605 m^3 (compared to -9,107 m^3 for the typical event - 0 m SLR scenario).

Typical Event – 0.5 m SLR

Figures 10.27 and 10.28 show bed level changes and volume losses and gains, respectively, near Rossbehy for the typical event – 0.5 m SLR scenario. Again, patterns of erosion and deposition are similar to those under the 0 m and 0.1 m SLR scenarios, although in many places, the extent of severe erosion has increased. This is particularly apparent in the breach, where more severe erosion extends further north, and along the length of the dunes, where the linear pattern of erosion becomes less discontinuous than is seen in the other scenarios.

Figure 10.29 summarises volume change both above the 0 m bathymetric contour (top) and between the -5 to 0 m bathymetric contours (bottom) for each of the 3 typical event SLR scenarios. Like the 0 m and 0.1 m SLR scenarios, net volume change above the 0 m contour for the 0.5 m SLR scenario was negative (-18,309 m^3, compared with -12,435 m^3 for the 0.1 m SLR scenario and -13,615 m^3 for the 0.5 m SLR scenario). Similarly, net volume change between the -5 to 0 m contour was positive (5,407 m^3, compared with 3,328 m^3 for the 0 m SLR scenario and 4,010 m^3 for the 0.1 m SLR scenario). Net volume change above the -5 m depth contour was negative (-12,903 m^3, a 29% increase in volume losses over the 0 m SLR scenario and a 26% increase in volume losses over the 0.1 m SLR scenario). Overall, these results indicate that under higher sea-levels, the "typical event" results in more material being eroded from the upper beach (>0 m contour) and moved offshore (below the -5 m contour).

Extreme Event – 0 m SLR

Figure 10.30 shows bed level changes near Rossbehy for the extreme event - 0 m SLR scenario. Bed level change ranged from -0.69 m to 0.89 m, as opposed to -0.72 m to 0.44 m for the typical event – 0 m SLR scenario. Like with the typical event scenario, high levels of erosion were concentrated along the length of the barrier on its seaward side, this time extending further seaward of the 0 m contour and thus affecting the mid- to lower- shore. High levels of erosion also occurred on the ebb shoal west of the breach, and the area over which they occurred was more extensive than that of the typical event scenario. Finally, there were high levels of deposition on the southern wall of the main inlet channel, straddling the -5 m depth contour along the length of the barrier on its seaward side, on the northern margin of the ebb shoal, and across an extensive area to the southwest of the shoal.

Figure 10.31 shows volume gains and losses for this scenario, which generally reflect the elevation change map. Net volume change above the 0 m contour was negative (-32,398 m^3). Between the -5 m and 0 m depth contours, net volume change was slightly positive (+842 m^3), resulting in an overall net volume change above the -5 m depth contour of -31,555 m^3 (as opposed to -9,109 m^3 for the typical event – 0 m SLR scenario).

Modelled dune volume change in the area where the TLS monitoring campaign was conducted (adjacent to the breach) was not entirely consistent with observations. The extreme event used to drive the model occurred during the monitoring period 2013-12-11 to 2014-01-16. Observed dune volume change during this period was -37.76 m^3/m^2. Modelled dune volume change for the extreme event – 0 m SLR scenario was only -0.10 m^3/m^2. It should be noted that a number of back to back events occurred during this monitoring period, and as such, the extreme change in volume observed may have been as a result of multiple events and not just the one used to drive the model simulation. Nonetheless, the modelled dune volume change appears to be a conservative estimate, given the gross underestimation, and thus calls into question the ability of the model to simulate dune volume change in the breach area as a result of the simulated extreme event. Perhaps one of the transport formulae used is

inappropriate under extreme conditions. Further research is required to examine this possibility.

Extreme Event – 0.1 m SLR

Figures 10.32 and 10.33 show bed level changes and volume losses and gains, respectively, near Rossbehy for the extreme event – 0.1 m SLR scenario. Patterns of erosion and deposition are similar to those under the 0 m SLR scenario. Differences between the two scenarios are more apparent upon examination of net volume changes. Net volume change above the 0 m contour was negative (-34,882 m^3, compared to +32,398 m^3 for the 0 m SLR scenario) and net volume change between the 0 to -5 m contour was slightly positive (+1,897 m^3, compared to +842 m^3 for the 0 m SLR scenario). The overall net volume change above the -5 m contour was -32,985 m^3 (compared with -31,555 m^3 for the extreme event – 0 m SLR scenario).

Extreme Event – 0.5m SLR

Figures 10.34 and 10.35 show bed level changes and volume losses and gains, respectively, near Rossbehy for the extreme event – 0.5 m SLR scenario. Again, patterns of erosion and deposition are similar to those under the 0 m and 0.1 m SLR scenarios, although in some places, the extent of severe erosion has increased. This is particularly apparent in the breach and along the length of the foredunes.

Figure 10.36 summarises volume change both above the 0 m bathymetric contour (top) and between the -5 to 0 m bathymetric contours (bottom) for each of the 3 extreme event SLR scenarios. Like the 0 m and 0.1 m SLR scenarios, net volume change above the 0 m contour for the 0.5 m SLR scenario was negative (-41,260 m^3, compared with -32,398 m^3 for the 0.1 m SLR scenario and -34,882 m^3 for the 0.5 m SLR scenario). Net volume change between the -5 to 0 m contour was slightly positive (+276 m^3, compared to +842 m^3 for the 0 cm SLR scenario and +1,897 m^3 for the 10 cm SLR scenario). Overall net volume change above the -5 m depth contour was negative (-40,984 m^3, compared to -31,555 m^3 for the 0 m SLR scenario and 32,985 m^3 for the 10 cm SLR scenario).

The results for the "extreme event" simulations indicate that under higher sea-levels, 29% (0 m SLR scenario) to 31% (0.5 m SLR scenario) more material was removed from the nearshore zone (above the -5 m contour) and moved offshore (below the -5 m contour) than for the typical event. Figure 10.37 graphically summarises net volume changes above the -5 m depth contour for each model scenario, which were each presented in this section.

In general, budget volume changes were greater for the area above the 0 m contour than for the nearshore zone between -5 m and 0 m depth. This was likely due to the vulnerability of the foredunes, which are large stores of sediment, to storm waves. The near lack of budget volume changes for the -5 to 0 m zone (including in the backbarrier salt marsh) suggests the material eroded from the foredunes and upper beach was transported further offshore, although some of this material appears to have moved seaward of the 0 m depth contour in the form of a longshore bar.

In the breach, erosion was concentrated on the seaward side of the breach in the storm scenarios. This area of erosion was directly adjacent to an area of accretion on the landward side. As sea-levels increased (for the typical and extreme event scenarios), both the area characterised by erosion and the area characterised by accretion increased. This suggests that as SL rises, more material from the shoreface could be dumped into the backbarrier during storms. From the backbarrier, this material will likely eventually make its way into the main inlet and either onto the ebb delta or further seaward.

10.3.2 Results: Changes in transport magnitude and direction

Figure 10.38 shows nearshore water levels at UTM coordinate 431732.69, 5770864.57 (shown in fig. 10.14) during the three fair-weather simulations (0 m SLR, 0.1 m SLR, and 0.5 m SLR). While the simulation began on 1 Feb 2012 at 00:00, a 24 hour spin-up[17] meant water levels did not reach statistical equilibrium until 2 Feb 2012 at 01:00, therefore water levels are shown only from this point. At no point during any of the three "fair-weather" simulations (0 m SLR, 0.1 m SLR, 0.5 m SLR) did water levels reach the height of the sediment tracer

[17] Spin-up is the time taken for a model to reach a state of statistical equilibrium under the applied forcing (The NOM Group, 2003).

injection point coordinate (UTM coordinate 433532.807, 5770466.742 – shown in fig. 10.13), therefore transport magnitude (bed load) was zero for the duration of all three (fair-weather) simulations.

Figure 10.39 shows water levels for the nearshore coordinate (UTM coordinate 431732.69, 5770864.57 – shown in fig. 10.14) and for the sediment tracer injection point coordinate (UTM coordinate 433532.807, 5770466.742 – shown in fig. 10.13) during the three typical event simulations (0 m SLR, 0.1 m SLR, and 0.5 m SLR). Nearshore water levels are shown to put into context water levels at the sediment tracer injection point, which was only inundated for part of the tidal cycle during all three typical event scenarios. Maximum water levels at the sediment tracer injection point coordinate for the 0 m SLR scenario, 0.1 m SLR scenario, and 0.5 m SLR scenario were 1.87 m, 1.97 m, and 2.36 m (above MSL), respectively[18]. The maximum durations of inundation for the 0 m SLR scenario, 0.1 m SLR scenario, and 0.5 m SLR scenario were 7 hours, 7 hours, and 9 hours, respectively.

Figure 10.40 shows a time series of bed load transport magnitude for the sediment tracer injection point coordinate under the typical event scenario for all three SLR scenarios. Because water levels only reached statistical equilibrium on 25 January 2012 at 18:50, only data from after this time is presented. The mean bed load transports recorded for the 0 m SLR scenario, the 0.1 m SLR scenario, and the 0.5 m SLR scenario were 0.0004 $m^3/s/m$, 0.0004 $m^3/m/s$, and 0.0006 $m^3/s/m$, respectively (summarised in figure 10.41).

The directions of transport for the typical event scenario 0 m, 0.1 m, and 0.5 m SLR scenarios were plotted on a compass rose (figure 10.42). For all three scenarios, the overall majority of transport was onshore (from the west). In contrast, observations from the TLS tracer experiment suggest a dominant alongshore transport, although the June experiment was not conducted under

[18] It is pertinent to note that water levels for the extreme event scenario were lower than those of typical event scenario. This is due to the fact that the extreme event did not coincide with a higher tidal range. Given the fact that extreme events are less likely to occur, and therefore less likely to coincide with a higher tidal range, than typical events, it may be more realistic to run the extreme event over the course of such a tidal cycle. For a graphic summary of maximum water levels for all nine scenarios, see fig. 10.37.

storm conditions and, due to the nature of the experiment, it was unclear how directions of transport varied while the beach was submerged.

Figure 10.43 shows water levels for the nearshore coordinate (UTM coordinate 431732.69, 5770864.57 – shown in fig. 10.14) and for the sediment tracer injection point coordinate (UTM coordinate 433532.807, 5770466.742 – shown in fig. 10.13) during the three extreme event simulations (0 m SLR, 0.1 m SLR, and 0.5 m SLR). Again, nearshore water levels are shown to put into context water levels at the sediment tracer injection point, which was only inundated for part of the tidal cycle during all three typical event scenarios. Maximum water levels reached at the sediment tracer injection point coordinate for the 0 m SLR scenario were 0.1 m SLR scenario, and 0.5 m SLR scenario were 1.58 m, 1.67 m, and 2.05 m (above MSL). Figure 10.44 illustrates how this compares with the fair-weather and typical event scenarios. The maximum durations of inundation for the 0 m SLR scenario, 0.1 m SLR scenario, and 0.5 m SLR scenario were 5 hours, 5 hours, and 10 hours, respectively. Figure 10.45 illustrates how this compares with the fair-weather and typical event scenarios.

Figure 10.46 shows a time series of bed load transport magnitude for the sediment tracer injection point coordinate under the extreme event scenario for all three SLR scenarios. Because water levels only reached statistical equilibrium on 27 December 2013 at 12:30, only data from after this time is presented. The mean bed load transports recorded for the 0 m SLR scenario, the 0.1 m SLR scenario, and the 0.5 m SLR scenario were 0.0003 m^3/m/s, 0.0003 m^3/m/s, and 0.0004 m^3/s/m, respectively (figure 10.47). These figures were slightly lower than those of the typical event (which were 0.0004 m^3/s/m for the 0 m SLR scenario, 0.0004 m^3/m/s for the 0.1 m SLR scenario, and 0.0006 m^3/s/m for the 0.5 m SLR scenario).

While for the typical event the dominant direction of transport was mostly from the west (onshore), for the extreme event simulation there were two dominant directions of transport for the three scenarios - from the west (onshore) and from the northeast (alongshore; figure 10.48). Contrary to expectation, no alongshore transport occurred during the simulation.

10.3.3 Summary of results

- **As sea-levels increased, there was an increase in the volume of sediment removed from the nearshore zone (above the -5 m depth contour) during typical and extreme events.** As sea-level increased under the each of the typical and extreme event simulations, so did the volumes of sediment lost on the upper beach (the zone above the 0 m contour) and in the nearshore zone as a whole (above the -5 m contour). Figures 10.22, 10.29, and 10.36 graphically summarise these volume changes.
- **As sea-levels increased under the fair-weather scenario, net deposition decreased dramatically.** Net volume change decreased from +4,895 m^3 under the 0 m SLR scenario to -261 m^3 under the 0.1 m SLR scenario and +112 m^3 under the 0.5 m SLR scenario).
- **As sea-levels increased under the typical event and extreme event scenarios, so did the duration of inundation in the intertidal area near the breach.** At the sediment tracer injection point (UTM coordinate 433532.807, 5770466.742 – shown in fig. 6.13), the durations of inundation for the typical event 0 m SLR scenario, 0.1 m SLR scenario, and 0.5 m SLR scenario were 7 hours, 7 hours, and 9 hours, respectively. Durations of inundation for the extreme event scenario were 5 hours, 6 hours, and 10 hours, respectively.
- **Sea-level does not appear to have a major impact on bed load transport magnitude or direction**. Average and maximum values of bed load transport were not affected by sea-level for any of the simulations. Neither were the dominant directions of transport.
- **Comparison with observations from the TLS monitoring campaign and sediment tracer experiment suggest that the model may be capable of reasonably simulating sediment transport patterns, and even, in the case of the typical event scenario, predicting dune volume changes near the breach.** Patterns of erosion and deposition in the area near the breach were consistent with observations from the TLS monitoring campaign for the typical and extreme event scenarios. Modelled dune volume change under the typical event – 0 m SLR scenario was consistent with observations in the area where the TLS monitoring campaign was conducted (modelled dune volume

change = -0.05 m^3/m^2; observed dune volume change = -0.07 m^3/m^2). However, under the extreme event scenario, the model under predicted dune volume change by a factor of more than 300 (modelled dune volume change = -0.1 m^3/m^2; observed dune volume change = -37.76 m^3/m^2).

Further discussion on the limitations of the model and the experimental design is presented in chapter 11.

11 Discussion

This chapter focuses on the implications of the findings of the previous three chapters and how they relate to one another in the context of the wider literature. It then goes on to argue that existing conceptual models of barrier breaching and subsequent evolution do not fully address the influence of storms on barrier breach evolution. Finally, a conceptual model of storm impacts on breached barrier dunes under the influence of SLR is presented based on the findings of this research.

11.1 Summary and discussion of findings

The aim of this study was to evaluate the importance of storms as a driver of morphological change on a breached barrier system under present and potential future sea-levels. A series of objectives related to this aim were outlined in chapter one. These were collectively addressed in this thesis using a variety of methods, including topographic monitoring and modelling, GIS analysis, sediment tracing, and numerical modelling. While the findings were presented in chapters seven through ten, they are discussed in depth in the following sections.

11.1.1 The viability of TLS as a monitoring technique in vegetated coastal dune environments

Terrestrial laser scanning is quickly becoming a favoured tool of geomorphic inquiry, yet there is a lack of standardised approaches to its employment in the published literature. This study, therefore, had to address numerous methodological issues, specifically those related to its application in coastal dune environments, and it built on the work of Feagin *et al.* (2014), Montreuil *et al.* (2013), and others. Methods for collecting and processing TLS data were developed for Inch/Rossbehy and presented in chapter 7, including methods for registering multi-temporal scans, filtering vegetation from TLS point clouds, generating digital elevation models (DEMs) and DEMs of difference (DODs), and calculating volumetric change.

One way in which this study was unique was in that it employed a relatively sophisticated technique for vegetation filtering. Others working with laser scanned data in vegetated coastal environments (*e.g.* Montreuil *et al.*, 2013; Feagin *et al.*, 2014) commonly use simple filtering techniques like lowest points

analysis (LPA), which has been shown to perform poorly in such environments. For example, in tests conducted at a vegetated saltmarsh on the west coast of Ireland, LPA was found to introduce errors of up to 16 times larger than those deriving from any other single potential source of error, including scan registration, GPS error, georeferencing, and target position definition (Coveney and Fotheringham, 2011). In this study, qualitative preliminary tests showed that LPA performed poorly on the Inch and Rossbehy test datasets when compared with the more sophisticated multi-scale dimensionality criterion classification technique, developed by Brodu and Lague (2012). Statistical assessments of error associated with vegetation filtering using this technique confirmed that residual errors associated with filtered clouds were significantly lower than those associated with unfiltered clouds for both Inch and Rossbehy test datasets (see chapter 7). Given the satisfactory performance of the technique at the two sites, it's fair to suggest that this technique could also be applied elsewhere in similar environments. Also, it would be interesting to see if clouds filtered using LPA would yield similar results to that of the multi-scale dimensionality criterion technique, and whether or not errors associated with LPA are significantly greater than those associated with that of the multi-scale dimensionality criterion technique. As this was beyond the scope of this study, such an assessment was not carried out here.

Despite attempts to reduce the introduction of error to the TLS datasets where possible, in multiple cases (15 out of 26), the volumetric error margin exceeded the actual magnitude of volume change, thus rendering rates of volume change inconclusive for many of the monitoring periods. The best results were achieved in the case of the Rossbehy dune data, where rates of volume change were relatively high (average rate of volume change = -0.165 m^3/m^2/day). There, in seven out of ten cases, the volumetric error margin did not exceed the magnitude of volume change. For the beach, however, the volumetric error margin exceeded the magnitude of volume change in five out of eight cases. When compared to dune volume change, the average rate of beach volume change was relatively low (0.003 m^3/m^2/day). How these errors compare with those reported in other similar type studies is difficult to comment upon, as many studies fail to report on volume error and/or only report on error associated with the generation

of individual DEMs, not for DODs (*e.g.* no assessment of error propagation is performed). For example, in their study on seasonal variations in embryo dune morphology, Montreuil *et al.* (2013) only reported their elevation error, even though volume change was calculated. Elevation error associated with DEM generation from 1 cm resolution scans obtained in that study was ±0.068 m, which is comparable to the elevation change error margin for Inch (±0.05 m), but not for Rossbehy (±0.41 to ±0.44 m). Feagin *et al.* (2014) also only reported on elevation error, again, despite having presented volume calculations. For 1 cm resolution scans in a vegetated dune environment, elevation errors associated with scan registration ranged from ±0.064 to ±0.229 m. In a study by Pietro *et al.* (2008) that involved the use of TLS to monitor beach nourishment performance, volumetric error margins associated with DEM generation were reported, but an analysis of error propagation was not performed. In that study, for 0.2 m resolution scans, volumetric errors ranged from ±973 m^3 (±1% of total volume) to ±1248 m^3 (±1.4% of total volume). The range of volumetric error margins in the present study was substantially wider, ranging from ±179 m^3 (1.2% of total computed volume change) to ±2983 m^3 (almost 20x the magnitude of the computed volume change). This is likely due to the introduction of error from vegetation filtering, which was not an issue in the Pietro *et al.* (2008) study.

The use of TLS in this research illustrates the importance of error propagation in chronotopographic and volume change studies, specifically in vegetated coastal environments. The careful identification and quantification of potential sources of error are important in such studies, as they limit the level of change detection that is possible using TLS (or even other point cloud) data. While TLS may be a useful monitoring tool, this is only true provided error margins do not exceed actual measurements of change. In the published literature, there is a particular emphasis on the quantification of registration errors. However, errors associated with vegetation filtering, DEM and DOD generation, and volume change should also be quantified in chronotopographic studies, as they can contribute to a substantial widening of error margins, which was demonstrated in this research.

While it's true that TLS can play an important role in coastal process studies (through the delivery of precise terrain information), it's reasonable to question whether such high degrees of precision are truly necessary to advance our

understanding of morphodynamics, especially given the issues described previously. In addition, as data acquisition becomes ever more efficient, there is a danger of *data quantity* overriding *data quality*. In this study, the collection of highly precise high-resolution data was essential to ensure that enough data would be obtained to reasonably capture the spatial heterogeneity of the ground surface, especially as many of the ground returns from the laser-scanned data were obscured by vegetation. In a broader context, though, the question of "why do we need *so much* data" demands attention. LiDAR technology – both airborne and ground-based – allows for the efficient capture of immense datasets. Once these datasets are collected, they can be stored almost indefinitely on relatively small devices. Such datasets may be of major importance in as yet unthought-of ways in future coastal process studies. Oftentimes old datasets are revisited for, for example, topographic change detection studies. Given the efficiency of data capture – it takes about as much time to collect *tens to hundreds of millions* of measurements in the field using a laser scanner as it does to collect, perhaps, *tens to hundreds* of measurements using an EDM or GPS – there's little reason not to capture such datasets. In terms of data quality, LiDAR data is unparalleled. More precise datasets mean smaller scale changes can be detected and more detailed studies can be conducted at the micro-scale. There are, therefore, compelling reasons to argue that TLS can meaningfully contribute to coastal process studies. One example is this. Traditional ground surveying methods are laborious and inefficient, a problem which has, for decades, prompted sampling in the form of the 2D beach profile. But studies based on cross-shore profiles completely ignore the mechanisms by which coasts function in the third dimension (perpendicular to and, especially, oblique to the cross-shore profile). Many studies supplement beach profile data with information derived from aerial photographs or, more recently, aerial LiDAR data to account for this, but these types of data are most often only available at infrequent and/or irregular intervals. TLS, however, bridges this gap by allowing for the provision of 3D coastal elevation data on-demand. Not all studies require highly precise 3D coastal elevation data. In the context of this research, it can be argued that the precision and level of detail may have been greater than necessary to investigate the key issues of process functioning, but the technique and data produced were of distinct value in that a lot of information could be gathered

regularly and at any given time (*e.g.* in the immediate aftermath of a storm). No other technique can deliver such datasets.

TLS technology has advanced significantly since the introduction of the Leica ScanStation (the instrument used in this study) and an update on this merits attention. While the instruments (new and old) are all more or less capable of capturing the same information, newer scanners have become more portable and more efficient, increasing their practical value. For example, the Leica ScanStation could collect 4,000 measurements per second (Leica Geosystems, 2006). Leica's newest model, the P40, can collect up to 1,000,000 measurements per second (Leica Geosystems, 2016), meaning significantly less time is required to obtain the same amount of information in the field. The ScanStation required a hefty kit to operate, including a laptop and large external batteries or a generator. The P40 is powered by internal batteries and can be operated from a full colour touchscreen built into the instrument itself. The Faro Focus, which can also capture up to 1,000,000 measurements per second, is even smaller and more portable than the P40. Faro scanners, though, are not built for rugged outdoor work, which Leica prides itself on (Paul Burrows, *pers. comm.*). The Faro Focus is also phase based, while the Leica ScanStation and P40 are pulse based which is (slightly) less accurate. Over the last decade, there has been a shift in the industry from pulse-based scanners to phase-based scanners and back to pulse-based scanners (John Meneely, *pers. comm.*). Pulse-based scanners are capable of capturing data over much wider distances than phase based scanners, but data acquisition is slower. Choice of scanner depends entirely on project objectives. The newest laser scanning innovations are in mobile scanning technology – from instruments mounted on motorised vehicles to UAVs. These require either inbuilt or external intermal measurement units (IMUs) to correct for motion. Mobile scanning technology can be useful in meso-scale coastal process studies where the coverage of large areas (on the order of thousands of square metres – say an entire coastal barrier system) is not possible using traditional survey methods.

11.1.2 The influence of storms on Inch/Rossbehy

Previous research by Sala (2010) and O'Shea *et al.* (2011) suggests that storms played an auxiliary role in barrier breaching at Rossbehy. While there is some evidence to suggest that the period between 2004-2009 had a higher than average concentration of winter storms, Sala (2010) argued that breaching was more likely due to a decline in beach volume in the early 2000s rather than the impact of one or more events. But given the fact that tidal exchange is still occurring through the breach at the site eight years on, the role of storms in shaping the system with this new morphological configuration now requires attention.

Attempting to answer the question of 'are storms the primary driver to morphodynamic change at Inch/Rossbehy' is difficult. This is because firstly, observations from this research suggest the two barriers respond differently to storms of the same magnitude. This is evidenced by the existence of statistically significant relationships between storm duration and dune retreat at Rossbehy and the non-existence of such a relationship at Inch. This is not to say storms are or are not a dominant control on either of the two barrier systems. Orford *et al.* (1999) showed that at the meso-scale (30-50 year timescales), storms are, in fact, an important control on the functioning of Inch, at least since the 19th century and probably earlier (*e.g.* Devoy *et al.*, 1995; Delaney *et al.*, 20120; Devoy, 2015). While the results of this study indicate storms may not be a dominant control on Inch over short (2-year) time scales, it does not rule out their dominance over longer time scales. In fact, this study may lend support to the premise that storms are a dominant control on Inch at the meso-scale in that little change was actually observed over the short-duration monitoring period. Such an observation might be expected if it is only events separated by relatively long (30-50 year) time scales that are likely to induce significant change – *e.g.* such events are not likely to occur in the space of a short-duration (2 year) monitoring period. At Rossbehy, on the other hand, storms may be important at this scale. Strong (statistically significant) negative correlations between event duration and rates of dune volume change ($r=-0.96$, $p<0.001$) suggest that event duration may be a strong predictor of dune volume change. This indicates that storms may play an important role in the morphodynamic evolution of Rossbehy. The difference in morphologic behaviour between Inch and Rossbehy is likely due to

the fact that Rossbehy is still in a post-breaching phase, and it is still unclear whether or not the system will self-organise and tend toward a new equilibrium or self-regulate and return to its pre-breach morphology.

Statistical analyses (*e.g.* linear regression and multiple regression analysis of variance) showed that the strongest predictor of dune volume change associated with storms that occurred over the 2-year monitoring period at Rossbehy was event duration. The other variables investigated - event frequency, event lag time, significant wave height and wind speed associated with events – did not meaningfully contribute to a predictive model of dune volume change. It is, however, possible that one or more of these storm variables could be important over time scales longer than that of the monitoring period. Event duration is likely of importance because the longer storm waves persist (*e.g.* $H_{sig}>H_{crit}$), the more likely it is that the storm surge will coincide with high tide, elevating water levels and bringing storm waves into contact with foredunes. This would be further exacerbated by an increase in sea-level, which sensitivity analysis results suggest could set in motion a cycle of foredune erosion that could potentially lead to the disintegration of the barrier.

Results from the sensitivity analysis experiment performed in MIKE21 (described in chapter 10) suggest that increases in sea-level could lead to an increase in the net volume of sediment removed from the upper beach and nearshore zone (the area above the -5 m depth contour) during storms and a decrease in deposition in this zone under non-storm conditions. In other words, under the SLR scenarios, more material would be stored further offshore than under a no SLR scenario, resulting in a sediment deficit. For example, under the typical event – 50 cm SLR scenario, 29% more volume was lost in the nearshore zone than under the typical event – no SLR scenario. Such a situation would mean the beach could become sediment starved, leaving the foredunes vulnerable to further erosion under future storm events and, possibly, leading to the eventual disintegration of the barrier (depending on how quickly sea-level rises). Other experts in the field (*e.g.*FitzGerald *et al.*, 2008; Devoy, 2015) have already posited that SLR may lead to the deterioration of barrier islands in general (as a result of an increase in long-term beach erosion). According to FitzGerald *et al.* (2008), the loss of nearshore sediment "results from complex, feedback-

dependent processes that operate within the littoral zone" (FitzGerald *et al.*, 2008, p. 604). The argument presented here, as evidenced by numerical model simulation results, is that it may be possible that such a feedback could be responsible for the eventual disintegration of Rossbehy. However, other larger-scale (non-storm related) process controls also play a role in the evolution of the system, namely sediment supply. This is, arguably, one of the most important controls on barrier morphology (Orford *et al*., 1996), and it's thought that a reduction in sediment supply in the early 2000s is what led to breaching in the first place at Rossbehy (Sala, 2010). In addition, there is evidence from Rossbehy and many Atlantic coasts showing that long-term (10^3 years) sediment supply from offshore sources is diminishing (Cooper *et al.*, 1995; Delaney *et al*., 2012). The removal of sediment from the nearshore zone during typical and extreme storms under higher sea-levels may further add to this reduction in supply. If this is the case, storms could be considered to play a secondary role in decade- to century- scale evolution of the barrier.

The idea that storms under higher sea-levels could lead to a net removal of sediment in the nearshore zone is supported by arguments put forth by FitzGerald *et al*. (2007). FitzGerald *et al*. (2007) maintain that beach erosion and the eventual destruction of the foredune ridge along barrier coasts can generally be expected as a result of the long-term effects of storms, negative or reduced sediment supply, and SLR. This is because, firstly, back-barrier saltmarshes are particularly vulnerable to SLR because vertical accretion is limited (slow). If rates of SLR are high enough, marshes will drown and begin to convert to open water. The change in the basin geometry would result in increased tidal exchange through the inlet (*e.g.* an increase in the tidal prism). An increased tidal prism facilitates erosion of sand from onshore, contributing to the more frequent segmentation (breaching) of coastal barriers. FitzGerald *et al.* (2007) suggest that such segmentation could even transform barrier island chains into island-only systems.

Despite the alignment with previously published literature, it should be stressed that the model results presented in chapter ten were based on hypothetical situations simulated by a model that requires further validation to increase confidence in model outputs. There is a need to validate the Dingle Bay model

using data collected during storm conditions. While an attempt was made to collect such data (see chapter 9), strongly energetic conditions meant that the experiment failed to yield useful information about rates and direction of sediment transport during storms. This is one drawback of modelling storms – collecting data for model validation can be (in practice) difficult under such conditions.

Another potential issue with the MIKE21 model (and others like it) is that it is 2D, and therefore there is an assumption that there is no vertical stratification. It is often assumed in the literature that there is no stratified flow in coastal areas because of low water depths. This is not always the case. For example, constant coastal winds can induce steady stratified circulation in the nearshore zone (Kong *et al.*, 1995). This could potentially be an issue in the Dingle Bay model, given the strong wind climate on the southwest coast of Ireland, especially during storms. In light of this, it's interesting to note that while the model was able to reasonably simulate observed dune volume change under a typical storm event, it under predicted dune volume change by a factor of more than 300 during the extreme event simulation. Further investigation into why it performed so poorly in this case would make for an interesting modelling study.

On the basis of the evidence presented in this thesis, storms are thought to play a secondary role in the evolution of the barrier. Previous research conducted at Rossbehy and elsewhere (Orford *et al*. 1996; Sala, 2010) suggests that sediment supply is likely the most important process control on barrier evolution. Given the fact that this PhD research suggests that storms can affect nearshore sediment budgets at the site, it is argued that they may play an auxiliary role in the future evolution of the system. However, limited information about the role of other process controls, such as vegetation, anthropogenic influences, etc., exists. As such, further research is required to better understand the role of storms relative to these controls.

There is no evidence from this research to suggest that storms play a key role in the evolution of Inch, where rates of volume change were relatively low throughout the duration of the study and bore little relationship with storm characteristics (although there was a statistically significant positive correlation

($r=0.74$, $p<0.05$) between significant wave height for events and rate of volume change, which may form the basis for further research). Storm duration was found to be an important predictor of dune volume change at Rossbehy and provisional numerical model results suggest that under rising sea-levels, storms may contribute to further dune erosion, possibly through a feedback mechanism that may eventually result in barrier disintegration.

The difference in rates of volume change between Inch and Rossbehy may be explained in part by the orientation of the main channel in the back barrier basin in relation to the Inch and Rossbehy field sites. The sinuous shape of the channel controls the initial direction of ebb outflow into the main inlet throat, and, to a lesser extent, into the new inlet. The Rossbehy site is located on the outer bend in the channel. In meandering channels, the outer bend is typically characterised by erosion due to an increase in the velocity of currents. This may help to create a tidal imbalance resulting in larger ebb flows at the Rossbehy site. As a result, the Rossbehy site is more vulnerable due to its orientation with respect to the main inlet channel.

11.2 Storms and future SLR as drivers morphological change at Rossbehy

Based on the findings of this research, a critical re-examination of conceptual models related to barrier breaching is proposed. In the published literature, there are a number of conceptual models describing the evolution of breached barriers (Balouin and Howa, 2001; Kraus *et al.*, 2002; Vila-Concejo *et al.*, 2003; Hartley and Pontee, 2008; Giese *et al.*, 2009). O'Shea and Murphy (2013) have pointed out, though, that only a small proportion of these apply specifically to conditions similar to Rossbehy in that these conceptual models were developed for barriers where breaching occurred:

(1) through storm response – this was not the case for Rossbehy, where steady shoreline recession was occurring for a decade prior to breaching;
(2) as part of inlet migration – also not the case for Rossbehy; or
(3) through a combination of both.

As such, O'Shea (2015) put forward a five stage conceptual model of the evolution of Rossbehy, which was described and illustrated in chapter 3 and is briefly revisited as follows:

- Firstly, the removal of the swash platform between 2004-2008 (just prior to breaching) left the drift aligned zone of Rossbehy vulnerable to wave attack (stage one) which eventually led to breaching.
- In stage two, the growth of the ebb tidal bar was said to facilitate dune erosion in the drift-aligned zone, as when waves pass over the bar, they change direction and approach the barrier perpendicular to the foredunes. The establishment of a channel between the ebb bar and the barrier facilitates the removal of sediment from the system on the ebb flow, leading to further erosion of the drift-aligned zone. As of 2015, Rossbehy was said to be in this stage of development.
- O'Shea (2015) used MIKE21 to simulate the long-term (30-year) morphodynamic evolution of Rossbehy. The output of that simulation forms the basis of stages 3-5 of the five-stage conceptual model. During stage three, which was forecast to begin in 2015 and continue until 2025, O'Shea (2015) argued the breach will continue to widen and the ebb bar will begin to migrate toward the drift aligned zone. Eventually (by 2025), the bar is expected to weld onto the drift shore (through channel infilling), dune retreat is expected to slow, and embryo dunes are expected to develop in the breach. The island is expected to disappear by 2030 (end of stage 4).
- In stage 5 (2030-2035), dune repair will be facilitated by the growth of the swash platform in the drift-aligned zone.

There are two major issues with this model. Firstly, it does not take into account the impact that storms might have on barrier recovery and secondly it does not account for SLR. If storms are, indeed, a dominant control on the evolution of Rossbehy over the short term, as evidence from this study strongly suggests, it would be prudent to consider what role they might play in all stages of this model. For example, some questions that might be addressed in this context include the following:

- Could storms slow or even prevent the shoreward migration of the ebb tidal delta during stages three and four?
- How might storms affect dune regeneration?
- How might storms affect the re-emergence of a swash platform in the drift-aligned shore in stage 5?

The influence of storms on the progressive breakdown of a gravel-dominated coastal barrier at Dunwich–Walberswick, Suffolk, U.K. has been documented by Pye and Blott (2009). During one significant, but non-extreme, storm in 2006, the combined effect of sustained high water levels and large onshore waves flattened almost 2 km of the barrier. Because a substantial volume of sediment was lost offshore during the storm, proposed artificial maintenance works had to be abandoned, as it was not possible to fully rebuild the barrier. Yet, many existing conceptual models of barrier evolution fail to take into account how this offshore movement of sediment after such events could impact barrier evolution. For example, Hartley and Pontee (2008) developed and trialled a method for assessing breaching risk in coastal barriers, which they purported could predict the permanence of a breach. This was essentially based on the ratio of the tidal prism and annual drift rates. However, this (commonly used) "averaging" approach trivialises the potential of storms, particularly extreme storms, to redistribute sediment. A similar conceptual model that employs this "averaging" effect includes that of Giese *et al.* (2009), on which the model of O'Shea (2015) was based. This model, based on the evolution of Nauset Beach, Cape Cod, MA (USA), describes the system's evolution in two phases. The inlet development phase describes breaching, which is said to be followed by a period of instability characterised by the development of multiple inlets. In the inlet migration phase, the tidal current becomes dominant over alongshore sediment transport, and the inlet migrates downdrift until, eventually, the (nearshore) remnants of the barrier weld back onto the barrier. The influence of storms between breach inception and closure, however, is not addressed. Others *have* integrated such events into conceptual models of breach evolution. For example, Balouin and Howa (2001) developed a conceptual model of short-term (months to years) inlet morphodynamics based on observations from the Barra Nova inlet, south Portugal. They found that major erosion of the barriers downdrift coast during

short storm events was an important contributor to barrier evolution. As such, they argued that inlet migration was dependent on both onshore drift rates *and* storm frequency. Similarly, Vila-Concejo *et al.* (2003) have argued that inlet migration is not a progressive evolution – it is characterised by the periodic interruption of storms, which can occur at any given time.

In addition to neglecting to address the potential impacts of storms on the sediment budget of Rossbehy, O'Shea's and other conceptual models of barrier breaching do not consider the potential impacts of future sea-level change (not to mention potential changes in storminess as a result of climate change…). The experiments conducted in this study suggest that under higher sea-levels, storms may result in the removal of more material in the nearshore zone, which could be detrimental to predicted slowing of Rossbehy's dynamic evolution. While O'Shea (2015) has argued the presence of recurves suggests that Rossbehy can recover and rebuild, this may not be the case in the context of an accelerated SLR. As Cooper *et al.* (1995) and Delaney *et* al. (2012) have pointed out, rates of SLR appear to have been important determinants of morphosedimentary behavior at Rossbehy in the past, albeit over much longer time-scales than were considered by O'Shea (2015). However, the potential for accelerated SLR due to potentially catastrophic Antarctic ice sheet mass loss means large increases in sea-level may occur more quickly than previously thought (on the order of decades) (Hansen *et al.*, 2015). As such, the short-term impacts of SLR on barrier coasts sedimentary budgets require attention.

The most widely cited model of coastal response to SLR is the simple 2D equilibrium profile model of Bruun (1954), which includes a number of variants (*e.g.* Davidson-Arnott, 2005). Despite numerous criticisms (*e.g.* Cooper and Pilkey, 2004), this, and models based on it, remain to a large extent highly popular. These models, however, are unsuitable for describing the evolution of breached barriers in that they cannot account for longshore transport, which is an essential component in the functioning of inlets. 2DH numerical models, on the other hand, can simulate the longshore component, and are therefore more useful for understanding the potential impacts of SLR on such systems. To the author's knowledge, this is the first such study to investigate the short-term impacts of storm events under different SLR scenarios using a 2DH numerical model.

11.3 Conceptual model of the short-term evolution of Rossbehy

As expressed in the previous sections, there is a need for a conceptual evolutionary model of the short-term evolution of Rossbehy under the influence of storms and SLR. Such a model, based on that of O'Shea (2015), is proposed. The model is herein referred to as the S-SLR (storms-sea-level rise) model. This model may be applied to barriers where breaching persists (*e.g.* a naturally occurring breach either fails to heal or widens over an extended period) during the course of its tendency towards a new equilibrium under a rising sea-level.

The proposed conceptual model is presented graphically in figure 7.1. Nested within each stage of the O'Shea (2015) model, a post-storm sub stage is introduced. Like the PS stage in the model developed by Vila-Concejo *et al.* (2003), this is a discrete phenomenon caused by high energy events which can be entered during any or all stages of barrier evolution. The influence of the sub-stage on each of the 5 original stages is outlined briefly as follows:

- **Stage 1 PS (2001-2007)** – Drift aligned dune erosion is facilitated by storms. Event frequency and lag time may be important during this sub-stage.
- **Stage 2 PS (2008-2015)** – The expansion of the drift-aligned zone adjacent to the breach is facilitated by storms. The duration of this stage may be governed by the frequency of long-duration events that occur during this period.
- **Stage 3 PS (2015-?)** – This stage is critically dependent on whether or not the migration of the ebb bar toward the barrier can keep up with further dune recession/breach widening, which is dependent on storms (and rates of post-storm dune recovery). If the migration of the ebb bar cannot keep up with the widening of the marginal flood channel from the landward side, stages 4 and 5 of the original model are not possible and the barrier may drown or segment.
- **Stage 4 PS** – A possible sediment deficit due to the migration of sediment further offshore affects embryo dune establishment. Their permanence is also threatened by storms.

- **Stage 5 PS** – A possible sediment deficit (as above) inhibits dune regeneration.

Evidence presented in this thesis and from the wider published literature supporting the development of these post-storm sub stages is presented as follows.

11.3.1 Storm influence on swash platform removal and growth of drift-aligned zone

Between 2004 and 2008 the removal of the swash platform left the drift aligned zone of Rossbehy vulnerable to wave attack during storms. Based an examination of events that occurred during this period, Sala (2010) speculated that an increase in the frequency of storms and a decrease in the summer recovery period might have amplified the swash platform erosion process. Erosion of swash bars due to (not particularly severe) winter storms has also been documented elsewhere by Balouin and Howa (2001). As such, the duration of stage one may be a function of the frequency and/or characteristics of these events. The addition of this sub-stage does not affect the original model except that it governs the potential duration of this stage when applied to other situations.

During stage two (post-breaching), the growth of the drift-aligned zone is facilitated by storms. Evidence from this study shows there is a statistically significant correlation between event duration and dune recession. The duration of this stage may therefore be governed by the frequency of long-duration storms, although this is tentative, as correlations do not necessarily imply causation.

11.3.2 Ebb tidal bar migration, channel infilling, and keeping up with breach widening

During stages 3 and 4, eventual breach healing is critically dependent upon whether migration of the ebb bar toward the barrier can keep up with further dune recession/breach widening. O'Shea (2015) established a limit of the width of the breach/dune edge erosion, based on the position of historical re-curves, dune vegetation line surveys, and model generated bathymetry, and used this to extrapolate the initiation of recovery (due to take place in 2033). It is argued

here that the potential redistribution of sediment during storms may affect the predicted timing of recovery or even whether or not recovery occurs at all. Model results from this study suggest that even under a small SLR (0.1 m), there may be an increase in material removed from the nearshore zone during storms and stored further offshore. This would mean that there is less sediment available for channel infilling, which was not accounted for in the original conceptual model. If the landward migration of the ebb tidal bar cannot keep up with dune recession/breach widening, it is possible that the barrier may drown.

11.3.3 Influence of sediment deficit on dune regeneration

If sediment budget is significantly affected by storms under a rising sea-level, this could have implications for dune regeneration. The effects of sediment deficits on dune development are well documented (Sherman and Bauer, 1993). A sediment-starved beach means less material is available for dune building. As such, this could affect the onset/duration of stages four and five. O'Shea (2015) alluded to this in that it was argued there is a high potential for dune regeneration at Rossbehy, but only provided storms don't destroy the embryo dunes.

11.3.4 Evolutionary timeline

The integration of these substages into the original model inhibits the establishment of a precise timeline of the evolutionary cycle of Rossbehy, as it is impossible to predict the instances and characteristics (magnitude, duration, coincidence with high tide, etc.) of storms over several decades and an "averaging" of these characteristics could drown out the importance of extreme events. As such, this revised conceptual model is more conservative in its predictive capability than the original model.

The influence of the post-storm substages at all stages is that they can delay or prevent the onset of the subsequent stage. Further research into the influence of storms under SLR on Rossbehy or similar type barriers at all stages would be required to establish more precisely thresholds that may limit barrier evolution.

11.4 Implications of this research

A classical problem in process geomorphology is posed by the fact that all landscapes are inherently unique, and unique drivers are acting on these unique

systems to produce unique landforms. As such, the question of "how representative can one system be of another" is relevant to this study, this discipline, and the environmental sciences in general. The issue of generalization in geomorphology has been extensively debated from a philosophical perspective. In an essay on the scientific nature of geomorphology, Richards (1996, p. 184) has pointed out that "process interpretations based on research in one location may be appropriate for that site, but not for a different one." As such, he argues, "it is essential to identify the boundary conditions provided by the field location, in order that generalization can proceed of the mechanisms inferred from observation" (Richards 1996, p. 171). In other words, it is easier to make generalizations about systems with more similar boundary conditions (*e.g.* coastal orientation with respect to wave approach, meteorological characteristics, antecedent morphology, etc.) than it is to make generalizations about systems with less similar boundary conditions. This issue relates to this study in that it begs the question, "are the findings presented, including the proposed conceptual model, applicable to more general situations?" It's difficult to test the proposed conceptual model given the somewhat uniqueness of the Inch-Rossbehy barrier system. Other breached barrier systems either heal quickly or are often closed artificially. For example, Bartra Island, in Killala Bay, Co. Mayo, is a barrier dune system with a history of breaching in Ireland. It is similar in geographical extent to Rossbehy. Breaching has broken the dune system into three distinctive sections but not to the extent that new tidal inlets have formed between the intact dune sections (Cooper and Jackson, 2011). If coastal barriers begin to disintegrate under SLR, it's reasonable to expect that some may behave in a similar manner to the Inch-Rossbehy system, in which case this study can serve as a baseline from which to compare other process studies of such systems. It would be expected that these systems would have similar boundary conditions (*e.g.* embayed, storm-dominated, etc.) to Inch-Rossbehy, such that generalizations about breach evolution could reasonably be made.

Even in the event that the Inch-Rossbehy system is unique, the findings of this study still have value in that they can help to inform future management practices in the local area. Given the uncertainties associated with SLR, the future of the

system is in question. This has implications for future planning and current and future land use practices for the entire Castlemaine Harbour area. These findings should, therefore, be of interest to local authorities and/or An Bord Pleanála, who are tasked with making informed decisions with regard to land use planning.

In addition to making a contribution to our understanding of the impacts of storms on Inch and Rossbehy (and arguably on breached barrier systems in general), this study has also made important methodological contributions, including the following:

- The development of a GIS technique for assessing dune scarp volume change using TLS data;
- The successful application of the multi-scale dimensionality criterion classification technique to vegetated coastal dune environments; and
- The development of an experimental design for testing the influence of storms under SLR on coastal barriers (using numerical modelling techniques).

Finally, this research has contributed to a wider body of literature on the impacts of SLR on coastal barriers. The findings are in line with other studies (*e.g.* Morton, 2008; FitzGerald *et al.*, 2007) that suggest that SLR will result in barrier retreat and the possible disintegration of coastal barriers. Future models of short- to medium-term barrier coast evolution must include the influence of storms and SLR – we can no longer ignore these components. From this work, it is evident that storms are capable of significantly affecting local sediment budgets. This has major implications for the worldwide management of coastal barriers and the areas that they protect. Hundreds of millions of people are at risk of losing their properties and/or livelihoods as a result of coastal barrier retreat and/or associated flooding. Results from coastal process studies, such as this one, are essential to inform effective management practices and mitigate or prevent potentially catastrophic damages to coastal communities.

11.5 Research limitations

This study evaluated the influence of storms on a breached barrier system on the west coast of Ireland and proposed a conceptual model of barrier breach

evolution. One limitation of this research is that due to the relative uniqueness of the system under investigation, it's difficult to test the applicability of the proposed model to other systems. As a result, caution must be exercised in making generalisations with respect to other barriers until an opportunity arises to test this model.

It should also be noted that observations about the storm driver are only valid for the timeframe over which the study was conducted. The influence of extreme but infrequent events, for example, may also be an important control on the system, but only at longer time-scales (as at Inch). Long-term monitoring would be required to test this.

In addition, while statistically significant correlations were observed between event duration and dune volume change at Rossbehy, it's important to remember that statistical correlations do not necessarily imply causation. While it seems logical that longer duration events are more likely to result in dune erosion, there may be other factors at play. In this study, attempts were made to identify these factors – *eg.* through multiple regression analysis – but only with respect to the storm driver.

With regard to the use of laser-scanned data from vegetated dune environments, propagated error was found to limit the level of change detection possible. During this research, efforts to limit individual error sources were made at all levels. It is recommended that similar studies should strive to do the same.

This study did not consider vegetation as an agent of erosion/deposition of the dunes at Inch and Rossbehy. Vegetation plays a key role in post-storm dune recovery and protects dunes from wind erosion. It can even reduce the impact of large storm surges (Feagin *et al*., 2010). The aim of this study, however, was not to examine the role of vegetation as a control on morphological change, but to examine the role the storm driver. It is true that some information about vegetation was available from the TLS data. However, to make sense of this would have been complicated. Feagin *et al.* (2012) attempted to quantify changes in vegetation volume from TLS data by interpolating the upper and lower surfaces of the vegetation and subtracting them. This approach doesn't take into account vegetation density or distribution. This is a major problem and

can be illustrated by the following example. After a major storm, the upper canopy of vegetation might be lower than it would have been before the storm (because it would have been depressed by storm winds and rain). Assume the volume of vegetation present is the same before and after the storm. Applying Feagin *et al*. (2012)'s technique, the two situations would yield different volumes. Given the experimental difficulties associated with using the TLS data to evaluate the role of vegetation, it was not considered in this research.

With respect to the numerical modelling experiments, there are two categories of limitations – those related to the experimental design used in this study and those related to the use of numerical modelling for coastal process studies in general. The experiments described in this study were based on local wave and weather station data and only events deemed "representative" were simulated. There is a continuum of event severity, and the analysis presented in this thesis is only based on these rather broadly defined "fair-weather", "typical event" and "extreme event" scenarios. As such, the results are only valid for these situations.

In the case of the Dingle Bay model, there is a need to validate this model under storm conditions. While such an attempt was made in the form of the December sediment tracer experiment, insignificant amounts of tracer were recovered, so transports could only be compared to the June experiment, which was performed under fair weather conditions. This is a limitation of modelling the impact of storms on coastal morphology in general in that collecting data for model calibration and validation can be difficult under high-energy conditions.

Finally, while it is maintained here numerical model outputs can yield considerable practical insights into the possible impacts of storms under different SLR scenarios, numerical modelling is not, nor should be, intended as a substitute for past experiences on Rossbehy or similar beaches, but rather as a useful ancillary tool for assessing the likely future behavior of barrier coasts under SLR. Additional similar experiments to those performed here, perhaps using different models, might (or might not) provide further evidence supporting the findings of this study.

12 Conclusion

The primary aim and objectives of this research set out in section 1.1 have been achieved using a combination of topographic monitoring and modelling, GIS analysis, sediment tracing and numerical modelling. Key findings of this study include the following:

- The Rossbehy barrier dunes in the drift-aligned zone of the barrier are particularly vulnerable to both extreme and non-extreme storms. Storm duration was found to be a key determinant of dune recession near the breach in the short term. This was not the case with Inch, which remained relatively stable over the duration of the 2-year study.
- Under future SLR, storms will likely contribute to a net offshore movement of sediment in the near shore zone of breached coastal barriers. If this material is transported to depths at which it cannot be returned to the shoreface, this will inevitably lead to shoreline retreat and the possible drowning of these protective landforms.
- The viability of TLS as a monitoring technique in coastal dune environments is limited by error propagation from multiple sources, although methodological advances, such as advances in the classification of complex scenes, are helping to minimise error from individual sources. The practicability of the technique in terms of precise data collection on-demand is unparalleled.

Key areas of innovation include:

- The development of the S-SLR conceptual model of the evolution of the Inch-Rossbehy barrier system.
- The development of a method for assessing scarp volume change using GIS.
- The development of a methodology for assessing the impacts of storms under future SLR using numerical modelling and TLS data.
- The successful application of the multi-scale dimensionality criterion vegetation classification technique to the Inch-Rossbehy datasets.

This chapter highlights the findings of this research, provides recommendations for future research in light of these findings and calls attention to the practical implications of this work.

12.1 TLS as a monitoring technique in vegetated coastal dune environments

A primary objective of this research was to assess the viability of TLS as a monitoring technique in vegetated coastal dune environments. It was found that, in practice, laser scanning could be a powerful tool for regular coastal monitoring, especially at timescales over which ALS is not feasible. However, issues related to scan registration, vegetation filtering, DEM generation and error propagation were found to be of particular importance, as they can limit the level of change detection that is possible in coastal dune environments. In the context of this study, many volume change measurements were inconclusive due to large error margins, particularly where little volume change occurred. As such, future studies might consider in the early planning stages what constitutes an acceptable margin of error and how might it be possible to work within that error margin.

An innovative aspect of this study relating to the analysis of the laser-scanned data was the development of a method for assessing scarp volume change using GIS software. Rapid retreat of the scarp at Rossbehy meant that many surveys did not overlap in plan view. To address this issue, the coordinate systems of the scarp clouds were translated prior to the generation of DEMs so that chronotopographic and volumetric change analysis could be performed in the horizontal in ArcGIS. To the author's knowledge, no other studies on cliff face change have been performed in this way. Other studies that involved using laser scanning to assess cliff face change either required point cloud overlap in plan view, involved the use of specialist software packages, or required programming knowledge (*e.g.* Lim *et al.*, 2005; Rosser *et al.*, 2005; Wawrzyniec *et al.*, 2007; Olsen *et al.*, 2012). This method of assessing rapid shoreline change may appeal to coastal researchers and/or managers with a background in GIS.

A second innovative aspect of this study in relation to processing the laser-scanned data was the application of the multi-scale dimensionality criterion classification technique of Brodu and Lague (2012) for filtering vegetation from

TLS datasets. Using ground truthing data collected at both field sites, statistical analyses helped to validate this method. Residual errors of filtered clouds were significantly lower than for unfiltered clouds, with mean residual errors for clouds at Inch reduced from 0.32 m (SD=0.64) to 0.06 m (SD=0.32), $t(111.876)=3.192$, $p<0.001$, and mean residual errors for clouds at Rossbehy reduced from 0.43 m (SD=0.18) to 0.33 m (SD=0.25), $t(139.915)=3.005$, $p<0.005$. Similar ground-truthing experiments could help others to quantify errors associated with vegetation filtration in vegetated coastal environments.

12.2 The influence of storms as a driver of morphological change

Over the duration of the morphological monitoring period (May 2012 to July 2014), at least 72 storms occurred. During this period, net volume change at the Inch field site (-231 m^3) was much lower than net volume change at the Rossbehy field site (-23,684 m^3)[19]. Maximum volume losses occurred at Rossbehy (-15,337±179 m^3) during the 2013-12-11 to 2014-01-16 monitoring period, and maximum volume losses occurred at Inch (-719±124 m^3) during the 2012-08-06 to 2012-10-06 monitoring period. The observed differences may be due to the different locations of the field sites with respect to the main inlet channel, with the Rossbehy site being more vulnerable due to its location on the outer bend of the channel.

Statistical analyses (simple linear regression and multiple regression analysis of variance) based on the TLS monitoring campaign data and WAM data obtained during this study revealed strong correlations between event duration and rates of scarp volume change at Rossbehy. This may be an indicator of the ability of storms, particularly longer duration storms, to drive morphological change at Rossbehy. As such, this information was integrated into a conceptual model developed to emphasize the influence of storms at various stages of Rossbehy's post-breaching evolution.

It has also been argued in this thesis that the lack of such a relationship at Inch indicates that storms may not be such an important control on the evolution of the Inch barrier system, at least in the short term.

[19] Net volume change at Rossbehy includes both the beach and scarp together.

The potential influence of storms under a rising sea-level was investigated using numerical modelling techniques described in chapter 10. The results of the experiment conducted in this study suggest that as sea-level increases, the volumes of sediment lost in the nearshore zone (quantitatively defined here as the zone above the -5 m depth contour) during storms would increase. Results also indicated that under fair weather conditions, net deposition in the nearshore zone under higher sea-levels would decrease. This would result in a net sediment deficit in the nearshore zone, which would have implications for the future evolution of the barrier. A recently developed conceptual model of the evolution of Rossbehy (O'Shea, 2015) suggests that infilling of the marginal flood channel (fronting the drift aligned zone of the breach) would eventually result in the welding of the ebb bar onto the barrier, facilitating breach repair. However, it is argued in this thesis that a sediment deficit in this zone caused by storms under a rising sea-level could have a negative impact on channel infilling and potential dune regeneration. As such, the original model was revised to reflect these new findings. The new S-SLR model is explicitly different from that of the original in that it integrates the influence of storms under a rising sea-level, includes additional scenarios, such as possible barrier drowning, and does not make precise predictions about the onset and timing of each stage. The model includes a post-storm sub stage that takes into account the influence of storms under a rising sea-level at each stage of the original O'Shea (2015) model. These sub-stages include the following:

- **Stage 1 PS** – Drift aligned dune erosion is facilitated by storms. Event frequency and lag time may be important during this sub-stage.
- **Stage 2 PS** – The expansion of the drift-aligned zone adjacent to the breach is facilitated by storms. The duration of this stage may be governed by the frequency of long-duration events that occur during this period.
- **Stage 3 PS** – This stage is critically dependent on whether or not the migration of the ebb bar toward the barrier can keep up with further dune recession/breach widening, which is dependent on storms (and rates of post-storm dune recovery). If the migration of the ebb bar cannot keep up with the widening of the marginal flood channel from the landward

side, stages 4 and 5 of the original model are not possible and the barrier may drown.

- **Stage 4 PS** – A possible sediment deficit due to the migration of sediment further offshore affects embryo dune establishment. Their permanence is also threatened by storms.
- **Stage 5 PS** – A possible sediment deficit (as above) inhibits dune regeneration.

12.3 Further research

Over the course of this research, various opportunities for further research were recognized. The following would be worthy of consideration:

- Comparable sensitivity analyses to those set out in this study, perhaps at other similar sites, to further examine the role of storms under a rising sea-level – the identification of thresholds under which barrier stability is threatened would be useful information for coastal managers and planners.

- A quantitative examination of the relative effectiveness of vegetation filtering techniques, including lowest points analysis and the multi-scale dimensionality criterion technique, in vegetated dune environments using field observations, including ground truthing data;

- Further examination of role of vegetation as agents of erosion/deposition during and between storms;

- Research on the reduction of propagated error in the GIS analysis of point cloud data – especially with regard to measuring volume change;

- Validation of the Dingle Bay numerical model using data collected under storm conditions (for modelling impacts of storms on Inch/Rossbehy) – an assessment of sediment transport patterns from a larger scale sediment tracer experiment or analysis of grain size trends may help to address this;

- Investigation into the under prediction of dune volume change for the Dingle Bay numerical model during extreme storm events - this could be done through a sensitivity analysis, whereby repeated simulations are

performed under alternative assumptions to try to identify which variable or variables are responsible for this under-prediction. Once identified, a correction might be developed and applied during future simulations;

- Further examination of the role of sequential as opposed to single events, perhaps using single-event outputs as inputs for additional event simulations;

- An investigation into the observed statistically significant positive correlation ($r=0.74$, $p<0.05$) between significant wave heights during storm events and rates of volume change at Inch;

- Additional similar numerical modelling experiments to those performed in this research, perhaps using different models such as Delft 3D - While the Delft 3D model can simulate flows and transports in 3-dimensions, it can also be run in 2DH mode, like MIKE21. If the Delft 3D model can reasonably reproduce the outputs of the MIKE21 model experiments, this would lend credence to the findings of this study.

- Examination of internal model variation in sediment mass changes to help verify model results – This can be performed by running multiple simulations using the same model inputs.

- Further modelling of long-term impacts of storms on Rossbehy under SLR, perhaps using a model similar to that described by Dai (2011) – This could help to forecast potential barrier responses to SLR, such as those presented in the S-SLR conceptual model, by providing quantitative information about, for example, the likelihood of barrier disintegration.

- Further research on the influence of storms on the nearshore sediment budget at Rossbehy – Additional field monitoring – contributing to the documentation of a long-term record of storm impacts – could help to quantify the influence of storms on the near shore sediment budget.

12.4 Conclusion

This thesis has made a contribution to our understanding of the impacts of storms under SLR on a breached beach-dune barrier system using a combination of

techniques, including topographic monitoring and modelling, GIS analysis, sediment tracing, and numerical modelling. It has demonstrated the importance of storms on short-term barrier evolution and the influence of storms under a rising sea-level on the nearshore sediment budget. Of particular interest, mean and maximum storm duration were shown to be significantly correlated with dune erosion at Rossbehy ($r=-0.96$, $p<0.001$ for mean duration of events and $r=-0.93$, $p<0.001$ for max duration of events). While a causal relationship between storm duration and dune erosion cannot be drawn, these findings lend support to the premise that storm duration is an important contributor to dune erosion at Rossbehy. More in-depth statistical exploration, coupled with direct observation and monitoring of post-storm beach response, are required to validate this relationship. Also of interest were the model experiment results, which suggested that under SLR, storms could contribute to a net nearshore sediment deficit. These findings are in agreement with other studies on the impacts of SLR on barrier islands (FitzGerald *et al.*, 2007; Morton, 2008). It is important to note here, though, that numerical model results are tentative and not intended as a substitute for past experiences on Rossbehy or similar type beaches, but rather as a useful supplementary tool for assessing the likely future behavior of the barrier under SLR. Additional similar experiments to those performed here, perhaps using different models, might provide further evidence supporting the findings of this study.

A conceptual model based on these findings has been proposed. The proposed S-SLR model can be integrated into future management policies and could contribute to more effective and sustainable management practices at Rossbehy and elsewhere. Presently, an overwhelming majority of coastal management policies do not take into account SLR and/or the possibility of an increase in extreme storm events as a result of climate change. The S-SLR model can help to inform future policies with regard to management practices by raising awareness of possible scenarios that may limit barrier evolution during different phases of barrier breaching. As an island nation, Ireland is inherently vulnerable to future SLR and must act accordingly. In addition, the proposed conceptual model can serve as a baseline for future coastal process studies on barrier breach evolution.

There is a need to better understand barrier-breaching processes as key to barrier development from a geomorphological perspective (as opposed to from a purely engineering perspective). Rather than studying barrier breaching as a "problem" that must be "solved," geomorphologists understand that it's part of the bigger picture of natural barrier coast evolution. Geomorphology, however, is still a field that is in its infancy in comparison to coastal engineering. As a result, many coastal processes are still not yet well understood. However, there is growing interest in (and a growing need to understand) the response of barrier coasts to SLR as a result of climate change. Geomorphologists and engineers are now working together to study these processes, and it is collaborative efforts like these that will ultimately help us all to better understand barrier breaching processes and coastal science in general.

Finally, an understanding of the impacts of SLR on coastal barriers is critical to the effective management of barrier coasts worldwide. As GMSLR accelerates, research such as that presented in this thesis is more relevant now than ever. Barrier coasts are extraordinarily important to society and life on the planet in general. They protect hundreds of millions of people from storm-induced flooding and erosion, provide important and irreplaceable habitats, prevent saltwater intrusion of freshwater tables, and their aesthetic qualities are arguably unrivalled. For these and many other reasons it would be foolish on our part as scientists and as a society not to aim to better understand these invaluable and dynamic systems.

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