Mesoscale temporal changes to foredunes at Inch Spit, south-west Ireland

by

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with 13 figures

Summary. The history of foredune development and loss is examined on a decadal-scale basis at Inch Spit at the head of Dingle Bay. Inch currently has a foreland area (2 km²) with four low foredune ridges fronting the southern section (2.5 km) of the barrier. Analysis of maps (1842 & 1898) ariophotographs (1949, 1967 & 1973) and field surveys (1993/94) of the shoreline has supported an interpretation of two to three possible sequences of sudden erosion of the foredunes and subsequent redeposition and reworking of foredunes over periods of 30 to 50 years. The erosion and deposition of the foredunes have been analysed in terms of variation in the following; human activity, sea-level, storm activity and longterm (annual to decadal) atmospheric forcing changes, but none of these factors individually account for foredune changes. It is argued on circumstantial evidence that the role of extreme storms and associated surge is the most probable key to understanding sudden foredune disappearance. Wave refraction modelling of extreme events has helped to specify the principal control on foredune loss. The result of specific microscale triggers (days) leading to mesooscale changes (decadal) is examined.


Introduction

The Atlantic west coast of Ireland is storm-wave dominated with a meso to macrotidal range. It has extensive unconsolidated glaciogenic deposits both along the contemporary coastline as well as offshore across the inner shelf. The post-glacial sea-level rise that decelerated in the early Holocene has extensively reworked shelf material to provide a range of sediment sizes to the coast line (Carter & Wilson 1993). The strongly discordant surface geology of resistant metamorphic and igneous lithologies of the Irish basement has provided a substantial range of headlands and bays in which an extensive range of littoral morpho-sedimentary environments has developed through the Holocene and recent times. A principal element of this Atlantic coastal assemblage is formed by dune systems (Orford & Carter 1988, Carter 1990, Wilson 1990, Carter & Wilson, 1990) of varying size (Curtis 1991) and varying depositional history (Carter & Wilson 1993). The initiation of primary dune emplacement has been suggested (Orford & Carter 1988, Carter & Wilson 1993), using the limited $^{14}$C dating control available, to have occurred during the main deceleration phase of relative Atlantic sea level rise set between 6–5k yr. B.P., depending on latitudinal position along the western Irish seaboard (Carter et al. 1989a, Shaw & Carter 1994). Most of these dune systems were emplaced in the many available re-entrants defined by the discordant coastline geology, emphasising the individuality and potential idiosyncratic nature of most wave-sediment cells in which dune deposition has taken place. The potential of reworking shoreface sand volumes onto the beach has been viewed by Carter & Wilson (1993) as decreasing in step with sea-level rise rate reductions in the late Holocene, such that they believed that much of the secondary changes in dune systems’ structure and morphology related to budgetary controls on diminishing sediment supply in association with climatic shifts, seen noticeably in storminess variation over time.

One possible indicator of such secondary dune mobility is shown in the development of foredunes or low wind-blown sand ridges parallel to the shoreline, landward of an upper beach position. Carter (1987) remarked on the intermittent temporal nature of these features along some Irish dune systems, a comment echoed by Pye & Tsoar (1990) in a regional context. Such comments underline the pertinent questions as to why foredunes appear and disappear, and what relation these low dunes have to the general development of the main coastal dune field. Although our under-
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standing of what controls foredune development directly in terms of beach morphodynamics, aeolian dynamics, sediment supply and vegetational control (e.g. Short & Hesp 1982, Hesp 1988, Sherman et al. 1998) is not in question here, the wider context of coastal changes which control long-term foredune development has only been partially addressed. Principal among these is the work of Psuty (1992) who recognised rhythmic spatial variation in foredune development along the barrier islands of Long Island, New York State, that had a temporal dimension measured on an interannual basis. Such foredune variation may well be related to rhythmic longshore wave control of beach state which in turn controls upper beach sediment supply. This is not viewed as a probable answer to Irish foredunes which may persist spatially for several decades and then totally disappear, only to reappear after a few years. Pye (1983) identifies the major interest in causes of long-term switches or even cycles of dune erosion and deposition particularly when the period of interest is mesoscale. Mesoscale in the coastal context is taken to cover time changes driven over interannual to decadal to century time periods and follows Terwindt & Kroon's (1993) definition. Sherman & Bauer (1993) following Psuty (1989) commented on how the contrast in beach and dune sediment budget status identifies different phases in both foredune and back-barrier dune development. Pye & Tsoar (1990) draw attention to the likelihood of anthropogenic forcing of coastal change by which foredunes, as a sensitive back-beach assemblage, are likely to be early victims of fluctuating longshore sediment supply problems. In this regard there is a requirement to establish likely causes of mesoscale foredune disappearance-reappearance oscillations along Irish coastal dunes from a set of potential forcing agents: atmospheric and anthropogenic.

This paper examines for the first time such a history of mesoscale foredune changes on a major Irish dune system (Inch Spit, Dingle Bay, Co. Kerry: fig. 1) and

![Map of Ireland showing location of Inch Spit in Dingle Bay, Co. Kerry, south-west Ireland. Note the ebb-orientated delta linking Inch with Rossbeigh Spit.](image)
debates a cause for this type of occurrence. This debate has to attempt to establish two sets of contrasting processes: the activity and rate associated with both dune formation and (δ) dune removal. Both sets of processes may be operating at different time scales. Interest in the rate of process centres on the possibility of identifying sudden triggers or microscale thresholds to mesoscale changes.

Methodology

There is a caveat to this paper in that there is a scarcity of documented evidence of process and response for the Irish coast, therefore much of the following information is inevitably circumstantial, drawn as secondary data from existing historical sources. Principal data come from the detailed map evidence supplied by the Ist Ordnance Survey of Ireland (1:10560) with the Inch sheet surveyed in 1842. At this scale the detail on mean High Water position is sufficient to allow this as a baseline for later maps and surveys (a linear accuracy of 1 in 500: SEYMOUR 1980). The subsequent revision (1897) plus vertical air photographs (1949 (fig. 2), 1967 and 1973 (fig. 3)), limited ground photography in 1984 and an EDM topographic survey of 1993–4 allow us to assess a history of foredune presence. The photos and ground survey allow us to define a beach-vegetation boundary which though not HW per se, spatially mirrors the HW position sufficiently along the open Atlantic stretch of Inch Spit, if offset by

Fig. 2. Vertical airphotograph mosaic of the distal element of Inch Spit in 1949. Note the foredune area at the south west side of the spit, which shows extensive shore-transverse reworking through blowout development. (Courtesy of the Ordnance Survey of Ireland and the Geological Survey of Ireland.)
Fig. 3. Vertical airphotograph mosaic of the distal element of Inch Spit in 1973. Note the complete loss of the foredune area from the south-west side of the spit, that was shown on figure 2. (Courtesy of the Ordnance Survey of Ireland and the Geological Survey of Ireland.)

c 30–50 m. All surveys were digitised with respect to a series of known control points (up to nine obtained) and transferred to a common spatial basis (fig. 4) using a GIS procedure (IDRISI). A number of profile positions were selected orthogonal to the coastline to allow assessment of shoreline change. Control points on an uninhabited spit were difficult to establish from map to photo so the net of precision improves towards the proximal end of the spit. Such distortion will be experienced in the long-shore direction rather than the cross-shore direction.

The air photographs provide for a visual assessment of foredune number, position and state in terms of blowout transformation. Sediment samples were collected from mid-beach, high water, foredunes and eroded face of the main dune at regular intervals along the spit (400 m). Sediment samples (< 1200 microns) were analysed for size structure using a Galai CS1 laser registering on 300 channels with each channel equal to a 4 micron size interval. Sediment size data were aggregated on a 0.25 o basis and sample median size, sorting and skewness calculated. Sample carbonate percentage was also obtained.

The methodological framework for the study is to examine the range of potential forcing factors to establish whether mapping of forcing and known dune response occurs and at what time scale. Such mapping is a balance between differentially time-resolved forcing data and response data. As most of the forcing data is based on hourly determinations (wind and water levels), forcing data has to be transformed into a longer time-based characterisation for comparison with morphological response.
Wind data were recorded at the Irish Meteorological Service station at Velantia (30 km to the south-west of Inch). Digital data available and analysed were for 1940–93. Data were initially sorted by annual wind events, where an event was defined as at least one hour of average wind velocity exceeding i) 10.2 ms$^{-1}$ (≈20 knots) indicating potential aeolian activity and ii) 17 ms$^{-1}$ (≈34 knots) indicating storm wave potential. Frequency of events and mean direction defined in terms of the number ($n = 1$ to $n$) of hours of persistent wind in excess of these thresholds were obtained on an annual basis. Both data sets were analysed with principal components analysis to establish an index of wind power. The first component for the 10.2 ms$^{-1}$ data set accounted for 39.4% of the variation and was classified as a measure of wind magnitude or power. The second component (21.2%) related to event number and duration. The annual scores for both components were standardised so that a value of zero equates to a year of average conditions for the period recorded. The first component
for the >17 ms\(^{-1}\) data set accounted for 73% of the data variation and was classified as a measure of storm power. Negative scores indicated years with no defined storms.

The nearest long-term record of annual secular sea-level changes to Kerry is that from the Newlyn tide gauge, Cornwall, south-west Britain. This annual observed record was filtered by use of a 19-year (term) unweighted moving-average to remove the effects of the nodal tide constituent. A record of higher-order sea-level changes associated with storm activity (positive surges) has been established using the method advocated by ORFORD et al. (1992). This characterises the distribution of hourly recorded positive surge heights on an annual basis, by using the coefficients of the best-fit linear regression between the logarithm of surge number (y) and surge elevation (x). The regression coefficients act as surrogates for annual characterisation of surge frequency (B0) and surge magnitude (B1). The negative value associate with B1 means that the lower the absolute value of B1 the higher the annual surge potential, which indirectly is a measure of the state of annual storminess, in terms of wave activity.

Interest in microscale extreme wave conditions which might be triggers for mesoscale changes was pursued through modelling of extreme wave conditions (hindcast 90th percentile values of H:6.6 m and T:13.6 s) relative to modal wave conditions (H:0.4 m, T:7 s) on an initial wave orthogonal vector of 215\(^\circ\)N adjusted to a bathymetric datum of low water position. Wave refraction in Dingle Bay was undertaken using HISWA (HOLTHUYSSEN et al. 1989) however adjustment for surge presence was not available. This analysis enables wave energy direction and wave-induced bottom stress magnitude and direction to be assessed. On this basis an assessment of wave-sediment impact at the shoreline can be inferred.

**Inch Spit and its aeolian components**

Dingle Bay (fig. 1) in south-west County Kerry is a macrotidal (c. 4.0 m HW Springs), narrow, tapering ria, aligned west-south-west and opening directly into the Atlantic. The estuary is broadly rectilinear, c. 16.5 km wide at the mouth (36 m bathymetric contour) and c. 9 km wide at the site of two spits (Inch and Rossbeigh) c. 27 km up estuary from the mouth. These two spits are approximately north-south orientated, sand-dominated coastal barriers separated by an ebb-tide delta, facing direct Atlantic south-west to westerly wave and wind activity. The spring tidal range increases up the estuary from 3.5 m at the mouth to 4.2 m behind Inch barrier (in Castlemaine Harbour) indicating a potential for amplification by the estuary structure.

Inch Spit (the northerly barrier) is identified as containing the largest dune area (1250 hectares) in Ireland (CURTIS 1991). It is a dune system which has been massively transformed by transverse parabolic dune activity, orthogonal to the contemporary shoreline and parallel to the prevailing wind (Photo 1, GUILCHER & KING 1961). The dunes are both elongated hairpin and bowl shaped (PYE 1982) with winter lakes. The tendency to elongation indicates a high velocity wind regime (GAYLORD & DAWSON 1987). Cross-spit aeolian sediment activity through the parabolic alleyways is still on-going, albeit intermittently (CARTER et al. 1989b). The development of the spit was probably controlled by the development of the large ebb-tide orientated delta which links Inch to the northerly extending Rossbeigh Spit that is
attached to the south side of Dingle Bay and lying seaward of Inch (fig. 1). Much of the sediment blown into the rear of Inch (Castlemain Harbour) is probably recycled into the ebb-tide delta which local fisherman anecdotally believe to change cyclically particularly along the north and south flanking side bars to the main channel. Although Rosbeigh exhibits a sand dune system partially superimposed on a gravel ridge basement, there is no indication that Inch exists on any continuous gravel-ridge basement.

Inch Spit, c. 6 km long and varying in width from 0.5–1.5 km, is dominated by three aeolian dunes components; i) low (<10 m OD) undulating vegetated sand hills on the northern proximal area (c. 30% of the spit’s area); ii) both active and inactive high (>20 m OD) parabolic dunes and alleyways aligned south-west to north-east in the southern distal spit area; and iii) an area (2 km²) comprised of a series of low linear foredune ridges (<4 m high) fronting the southern section (2.5 km) of the barrier on the open coast side. The present wide dissipative Atlantic beach (slope of 1 in 75) meets a strongly eroding dune-face scarp (<6 m in height) for c.2 km along Inch Spit from its northern attachment point to the southerly foredune area. This eroded dune-line scarp continues south where it is up to 20 m high and indicates a backstop position to the foredunes as well as defining the blowout entrances to the cross-spit parabolic dune alleyways. Infra-red stimulated luminescence age determinations from dune sediments in the scarped dune line identify the main dune line as 200–300 years old (WINTLE et al. 1998) indicating the principal dune erosion and therefore the baseline of foredune deposition as being of recent origin. Through the survey period of 1993–95 there were no indications of foredunes developing along the northerly section of the spit.

There is no difference in the median sediment size between ridges in the contemporary foredunes (mean: c. +2.35 Ø). All ridges show a general trend in decreasing median size towards the northern end, from +2.2 Ø in the south to +2.5 Ø in the north indicating a possible sediment pathway from the south to the north. There is no similar trend in the contemporary high water samples (mean: +2.25 Ø), or in samples from the scarped dune line (mean: +2.3 Ø at the dune base and +2.4 Ø at the dune top).

Timing of mesoscale foredune change

Figure 4 shows historical shoreline variation which specifies two longshore sectors separated by a longshore position of minimal change at 2.5 km from the spit neck. Variation in shoreline position is maximised in the distal section. At all sections of the seaward margin of the spit, the shoreline was close to its most landward position in 1842. Accretion of 150–200 m had occurred along the entire length by 1894. By 1949, the proximal section of the spit had receded by about 50 m while in the distal section accretion by a similar amount had occurred (fig. 5). Air-photographs for 1969 and 1973 record a period during which the spit’s distal shoreline had receded to a position similar to that of 1842. In 1993 the shoreline position was advanced to a position similar to the most advanced position of 1894 in the south and a position somewhat landward of the 1894 position at the proximal end. Changes in shoreline position at the distal section of the spit are shown by the air-photography and field observations to coincide with the presence during accretional phases of a series of vegetated, shore-
parallel dune ridges. It is proposed that the variation recorded on the maps of 1842 and 1898 reflect a similar change of foredune development. There is some potential for possible understanding of changes in the recent past given the period of maximum change between 1949 and 1967 and the period of near maximum foredune accretion during 1973 and 1994.

**Foredune structure**

To date there are two recorded phases of foredunes occurrence. The 1949 airphotograph identifies a phase of foredune development running parallel to the coast which had disappeared by 1967. This phase of foredunes is more accurately described as hummocky dunes (Pye & Ts'oar 1990). The 1994 ground survey identifies a development of four foredune ridges (fig. 6) in the same longshore position as the earlier set (fig. 2). On both occasions the foredunes shadow the form of a depositional bulge in the upper beach resting against the eroded dune line. In the earlier phase (fig. 2), the hummocky dunes are defined principally by a substantial cross-beach transverse dissection related to blowout activity that has yet to appear in the latest phase. Trends of blowouts are remarkably parallel suggesting dominance of wind activity from the south-west. Such transverse reworking of the dune is imposed by blowout extension between adjacent hummocks of marram (Ammophila arenaria) which grows rapidly under the moist Atlantic climate and steady aeolian sediment supply. The loss of foredunes should not be viewed as a function of loss of vegetation, it is more likely that it is a reduction or
cessation of sediment supply that is responsible for dune loss. The high order irregularity by vegetative clumping on the ground did not appear to disturb the low order linearity of ridges in the longshore parallel direction of the last phase of foredunes, in contrast to the condition in the earlier phase of foredunes (fig. 2).

The cuspathe bulge of the foredune dominated zone (fig. 6) suggests that the dune cores were not stochastically placed in terms of shadow dunes in the lee of the higher eroded dune line. The longshore continuity of ridges suggests that the dunes had a wavelain core which formed part of a prograding beach foreland controlled spatially by wave refraction. As upper beach progradation continued, the last high water beach ridge became the core of a new shadow dune. Hence ridge spacing is probably a function of beach ridge spacing per se, rather than aeolian activity. Foredunes as parallel ridges indicate a medium range of supply, often continuous in time but where vegetation controls the ultimate fixing and development of the dune. In the 1994 phase, foredunes are low (< 4 m) and broad (c. < 80 m wide). A ridge's low elevation is a reflection of fast progradation rate (20 ma⁻¹) and the loss of beach supply by the spatial intervention of a newly emerging beach ridge at high water. CARTER (1990) noted that prograding foredunes in Ireland are rarely over 4 m in elevation due to progradation rate. The survival of these foredunes over periods of 30–40 years of transgressive reworking by wind is a reflection of the lack of variation in wind vectors as well as possible rapid rebuilding of dunes post their erosional phase. This may also relate to a sheltering effect of the low dunes under the cliffline of the eroding dune face to the rear. The lessening in transverse reworking of the last foredunes phase may simply reflect a 20 year existence compared to the possible 40+ years of the earlier phase (see below).
Periods of foredune development should not be viewed as independent of beach state. Although SHERMAN & BAUER (1993) have identified the common occurrence of back-beach dune sediment surplus with beach face deficit, it does not appear that this is likely at Inch given the position of the foredune bulge relative to the northern edge of the ebb-delta. It is more likely that foredune building reflects similar positioned changes in beach status (SHORT & HESP 1982), although there is no independent evidence of this element from the map and photo data. The foredune's sediment source may be considered through the carbonate content of the sediment. Figure 7 indicates the percentage of carbonate content in the foredunes relative to the present beach (HW position). The relative absence of carbonates in the contemporary high water sediments along the north of the spit and the peak of carbonates opposite the faunal haven of the ebb-tide delta suggest that it is the latter area from where the high water sediments in front of the foredunes, are derived. The foredune sediment is in turn derived from the high water sediment. The northern directed foredune sediment size pathway is possibly a reflection of a high water pathway. The flood-directed sand waves of the exposed intertidal northern delta rim (airphotograph evidence) suggest that there is a tangential shorewards directed sediment supply along the distal end of the spit. This sediment is reworked by onshore wave action through the tidal cycle to form the dune ridge basement and at low water is exposed to aeolian activity. This interpretation suggests that cycles of foredune development are a consequence of sediment unloading from the flanks of the delta. The latest foredune phase required a mean annual sediment budget of c. $18 \times 10^3$ m$^3$ a$^{-1}$ to be brought on to the upper beach, of which probably no more than 20% was by aeolian activity.

Potential controls on mesoscale foredune changes

There is a number of possible mechanisms by which mesoscale oscillatory foredune changes could be induced; sea-level variability, wind variability for aeolian activity,

![Figure 7. Longshore variation in calcium carbonate content of sediment from the high water position and the crest of the 2nd and 3rd foredune line.](image)
variability in storminess and extreme events linked to periods of foreshore erosion/deposition. Consideration of these processes is based on the identification of meso-scale periods of process variability which might be related to foredune variability at Inch. Indirect sediment supply changes related to human activity in the coastal system also need to be considered.

**Human activity**

Abundant shell middens on Inch testify to a long (millenium-scale) human occupancy of the feature. Documentary sources (Smith 1756) identify Inch as a source of sand for agricultural soil improvements in west Kerry. Evidence of marram grass cutting for thatching leading to localised dune destabilisation is also cited by Smith. While there is no evidence of temporal variation in human use of Inch, the effect of the Great Irish Famine of the 1840s when many tenants were displaced may have led to an over use of Inch as a principal common land area. However no sensible estimate of human activity for this source can be made. It is tempting to link the shoreline recession of 1842 with the effects of destabilisation, but the same mechanism could not be used to explain the possible episodic erosion in the early 1900s or the observed erosion of the 1960s, nor even the increased supply of sediment that would be required after erosional phases, for dune rebuilding.

Land reclamation has been widespread in Irish estuaries (Carter et al. 1984, Orford 1988) producing marked changes in the tidal prism and estuary mouth deposition, particularly side bar development. It is noted that changes to both Inch and Rossbeigh Spit termini (progradation between 1842–1894) might reflect hydraulic adjustment to tidal prism changes stemming from 160 hectares of reclamation of marsh in Castlemain harbour. There are indications of agricultural improvements of the inter-tidal zone on the north shore of Castlemain relating to pre-famine subsistence agriculture, but the area is marginal and a very small fraction of the inter-tidal zone as a whole (<3%) unlikely to cause the major disturbance to the tidal prism required to force estuary mouth side deposition.

**Sea-level variability**

The Newlyn tidal record (1917–1992: Orford et al. 1997) was analysed to examine any potential link between dune development and sea-level fluctuation at Inch, which has been observed elsewhere (Christiansen & Bowman 1986). Although the two areas are geographically separated, both lie to the south of marked influence of post-glacial isostatic readjustment, and show a similar Holocene trend in slow deceleration of sea level. They might therefore be assumed to have responded similarly in the past century. Figure 8 shows the annual mean sea-level position for Newlyn. The regression line has been fitted to the nodally detrended data to specify the mean sea-level rise rate of 1.6 mm a⁻¹. It is clear that there was a peak in elevation of MSL by c. 5 cm in 1960 compared with the late 1950s and middle 1960s, that may correspond in time with the last phase of dune erosion. However, the peak of MSL in the late 1930s does not appear to be associated with any known erosion at Inch – suggesting that MSL per se is unlikely to be the cause of major foredune erosion.
Higher-order sea-level variation can be observed in terms of storm-surge generation. Figure 9 shows the 5 year smoothed annual surge generation potential for the Newlyn tide gauge data (Orford et al. 1997). This analysis indicates a smoothed trend of low surge potential in the 1920s–early 1930s, rapidly increasing in the early-1940s thereafter holding a relatively consistent level except for two dips in the mid-1950s and late-1980s. While the 1980s surge reduction is not altogether incompatible with the observed foredune development at Inch, general surge generation as a function of increased storminess in the early 1960s at the time of foredune disappearance does not appear to be noticeably different from surge levels in the early-1940s and early-1980s – times of foredune persistence. Foredune disappearance does not appear to be related to variation in general storminess as witnessed by annual surge generation potential.

Wind variability

The annual scores on the first (wind power) and second component (event number and duration) of the Valentia wind event time series (> 20 knots: 1940–1993) are shown in fig. 10A and 10B. These data point to a period of reduced wind events (reduction in aeolian activity) in the mid-1950s with potential annual peaks of aeolian activity in 1950, 1974, 1985 and 1989. Though these changes in aeolian potential may be helpful in considering the observed accretion of the last phase of foredune
Fig. 9. Annual potential surge magnitude (B1) recorded at Newlyn tide gauge, south-west Britain, 1917–1992. B1 read as an absolute value shows greater annual surge generation as B1 decreases.

building up to the present time, they indicate little about the cause of the erosion of the dunes in the 1960s, other than a reduction in aeolian potential in what might be the antecedent period before the last erosion phase. Component 2 shows a major reduction in event number and duration in the 1950s with a peak of activity in the late 1960s which does not coincide with any peak in event power (Component 1). Post 1960s, there are peaks of event occurrence in 1978, 1985 and 1989. Only in 1989 does event duration coincide with a peak in event power.

Figure 10C shows the trend in storm power (>34 knots). Due to some years in which the storm threshold has not been exceeded (negative scores not shown) the data were not amenable to smoothing. Given the probable erosion of the foredunes in the early 1960s, there appears to be little difference in the power of stormy periods peaking in 1951, 1959, 1968, 1978 and 1990 to account only for this one period of foredune erosion. However, the intervening periods of low storminess may account for ridge building in the last progradation phase. The analysis of wind data whether for aeolian transport activity or storm activity shows a lack of correspondence between periods of atmospheric forcing and foredune activity at Inch.

**Extreme events**

The variation in foredune presence at Inch could be considered in terms of explaining the brief and dramatic loss of dunes in an otherwise continuing phase of dune deposition and maintenance. Such a loss could be self-inflicted through a threshold inherent in the sediment exchange between nearshore and beach face, which is reached at a point where the offshore bathymetry no longer protects the upper beach and foredunes from effects of severe magnitude storms. As there is a need to account for the relatively sudden and dramatic short-term loss of dunes compared to subsequent long-term net growth of foredunes, it is incumbent to consider a single extreme event process as the erosional cause, rather than seasonal periods of above-average storminess.

The historical record contains two periods of maximum shoreline retreat, one that occurred some time before 1842 associated with a period of active parabolic dune activity and a second that occurred sometime between 1949 and 1967. The latter date
Fig. 10. Annual characteristics of wind and storm events recorded at Valletta (1940–93) using principal components analysis. A: Component 1 Wind event power > 20 knots. B: Component 2 Wind event number and duration > 20 knots. Both A and B time series have been smoothed using an unweighted 5-year moving average. C: Component 1 Storm power > 34 knots.
can be constrained as Guilcher & King (1961) observed foredunes (<6.5 m high) with vigorous growing marram, along the distal spit in August 1958. Consequently erosion is timed between 1958 and 1967. A review of historical weather data revealed that a severe storm (originally Hurricane Debbie) struck the west coast of Ireland on September 16th 1961. This produced maximum wind speeds which are the highest recorded at the two nearest weather stations to Inch at Valletia (45 ms\(^{-1}\)) and Shannon (48 ms\(^{-1}\)), since records began in 1916 (Rohan 1975). Prior to this event the legendary ‘Night of the Big Wind’ of January 6–7th 1839 had been recorded as having “caused more widespread damage in Ireland than any storm in recent centuries” (Rohan 1975). Reconstruction of the meteorological conditions at that time (Shields & Fitzgerald 1989) suggested that maximum mean wind velocity of 45–70 knots (23–35 ms\(^{-1}\)) occurred. Gusts in excess of 100 knots (51 ms\(^{-1}\)) were suggested by the extent of the damage sustained to some buildings (Rohan 1975). Between these two events one additional storm on the night of 26–27 February 1903 was regarded as being of almost similar magnitude and the most severe to have affected Ireland since the Night of the Big Wind (Rohan 1975), producing wind speeds of 31 ms\(^{-1}\).

The storms of 1839 and 1961 may have caused maximum shoreline recession at Inch. Both storms impacted at high tide, a few days after spring tides. While no record of the direct effects of the storm at Inch are available, Carr (1993) cited newspaper records of widespread storm impacts on the west Irish coast during the 1839 event. These accounts refer to an unusually high tide at Limerick (100 km to the north) and the stranding of fishes on beaches and dunes at Galway (180 km to the north). Evidence for erosion during the storm of 1903 is lacking but there is equally no evidence which indicates that erosion and subsequent accretion did not occur between 1894 and 1949.

In the light of these observations, it appears likely that only two, and possibly three very high-magnitude, low-frequency storm events may be responsible for triggering the rapid erosion of foredunes observed on Inch Peninsula in the historical period. The intervening periods were characterised by progressive foredune accretion. The latest phase of post-storm (1960s to 1990s) accretion indicates that this can occur in approximately 30 years, so that ample time existed for anticipation of a similar period of foredune building post-1903 and prior to 1949, i.e. the available deposition window.

Discussion

Observations have been made of the impacts of extreme storms on coastal dune and beach morphology from other areas, in particular the eastern coast of USA, where hurricane activity exerts a marked influence on coastal behaviour. Dolan & Lins (1986) indicated that of the 100+ hurricanes which struck North Carolina between 1900 and 1985, over 50% had wind speed greater than 44 ms\(^{-1}\) and the most severe (H. Camille) was 66 ms\(^{-1}\). These indicate that the Irish winds of 1839 and 1961 are within the upper half of the North American hurricane range. Hayes (1978) referred to events with wind speeds >55 ms\(^{-1}\) as large hurricanes and noted that storm-surge ebb-directed currents produced during such storms, scour channels and inlets and deposited sediment on the barrier shoreface and inner shelf.
VELLINGA (1982, 1983) in an analysis and development of a model of coastal erosion during storm surges indicate that a relationship existed on the eastern North sea coast between storm surge level above MSL and the quantity of material eroded from the beach and dune. The measured retreat at Inch (assessed by comparison of maximum accreted position and maximum eroded position) ranges from 250 m at a point 4 km along the spit to 100 m at a point 1 km along the spit. These coincide with areas of high dunes (20 m) and low frontal dunes where the associated volumetric sediment loss under such landward retreat equates to 200 m$^3$ m$^{-1}$ and 1250 m$^3$ m$^{-1}$. These values are an order of magnitude higher than the maximum recorded in the North Sea (VELLINGA 1982). However the suggested exponential relationship between storm surge level and volumetric erosion rates implied that the surge level required by such a relationship at Inch would be between 6 and 7 m. VELLINGA (1982) also showed that as the initial beach profile steepened, storm surge erosion increased and that increased storm duration increased the volume of material eroded. It was also concluded (VELLINGA 1982) that the bulk of the eroded sediment was deposited within a short distance (150–200 m) of the shoreline. ANDREWS (1970) noted that storm surge maxima are up to 3 m, although in funnel-shaped estuaries water levels of 7 m have been recorded at the heads of bays. HAYES (1978) lists storm surges up to 7 m on the Texas shoreline generated by H. CARLA in 1961. Although no record exists of storm surge levels in Dingle Bay, its distinct funnel-shaped morphology and the known amplification of tidal range under non-surge conditions provide strong indications of a potential for surge amplification towards the head of the bay.

The importance of a surge element is inferred from the results of wave refraction simulation of extreme wave statistics as shown in fig. 11. This simulation is of

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**Fig. 11.** Wave refraction analysis to model: A. wave energy direction and B. wave induced bed stress. Conditions relate to the hindcasted 90th percentile values of wave height ($H$) and period ($T$) directed onshore at Inch. Vectors of wave induced bed stress indicate both magnitude and direction of resultant stress.
steep waves at low tide and shows the baffling effect of the ebb-delta on the incoming waves. Although surface wave energy is directed onshore along the whole spit, the shadow effect of the delta means that whereas onshore bottom stress is very high on the delta edge and high on the proximal beach face of the spit (40 N/m²), it is severely reduced in the area of the foredunes to the south. This scenario is scarcely likely to allow major erosion of the foredunes. To achieve massive foredune erosion, the still water level of wave action would have to be raised by c. 3 meters on top of high water to allow the freeboard for oscillatory and reformed waves to move across the otherwise reflective face of the delta. This means that although severe storms are required to achieve the surge element, there needs to be a very wide spectrum of wave periods and heights generated within the storm such that although the longer period elements are likely to be dissipated by the delta front, there is still sufficient wave energy in higher frequency periods to pass over the delta on the surge and break against the dunes. The importance then is not in storms per se, rather in the storms which have a major surge potential. Figure 11 may help to confirm why the foredunes do not appear to build beyond the distal end of the spit, in that during storms without significant surge increment there is sufficient wave energy to limit any northerly extension of the foredunes.

It would follow from the presence of an extreme surge against the beach face, that there would be seaward-directed return flows which would act as an agent of offshore-directed sediment transport. Given the effective wave field on the foreshore, such flows might well be directly initially, by effective head pressure, alongshore between the spit and the delta and emerge into the back spit area. Any ponding-up of surge water in this way would eventually break out into the bay via the main ebb-directed channel and lead to channel side-bars and delta-top deposition particularly to the northern side of the delta given the wave energy concentrations to the west and south side of the delta as still water level drop on a falling tide.

This wave simulation suggests a mechanism by which major beach and dune erosion can take place with dispersal of the eroded sediment firstly alongshore and then offshore during the storm return flow. Such an interpretation is consistent with observations made elsewhere (Vellinga 1982, Hayes 1978, Davis & Andronaco 1987) which point to storage of storm-eroded sediment in the beach, from whence it may be returned landward under lower wave energy.

Following the substantial erosion of the foredune area, excess sediment in the nearshore would feed a sediment return to the beach face under fairweather wave conditions. Under conditions of high sediment supply, accretionary ridges are likely to form at the distal section of Inch as a mixture of initial wavelain high water ridges and then enhancement by aeolian deflation of intertidal areas to produce foredunes. Figure 12 identifies the wave energy refraction orthogonalns which tie these wavelain peaks to the south end of the spit where sediment availability on the ebb-delta is maximised. Figure 12 also shows a path of consistent onshore wave induced bottom stress by which sediment is directed along the north flank of the delta, onto the south end of the spit. Although there are larger onshore wave induced stresses to the north of the spit, there is no consistent path by which sediment can be brought to this area as they are also bounded by an offshore directed stress.

The foredune's sediment size fining towards the north identifies the reworking of sediment from the south end of the spit which is moving shorewards from off the
Fig. 12. Wave refraction under hindcasted modal (50th percentile) wave conditions at Inch. A and B as in fig. 8.

ebb-delta front. Visual observations show that waves break progressively along the northern rim of the delta moving sediment to the north-east especially at low tide (fig. 12 identifies the wave energy directions to support this perspective). The flood orientated sand waves in the delta marginal channels at the delta/beachface margin indicate a returning sediment south-east along the beach face in front of the foredunes during the flood tide. Sand waves are reworked by incident waves which move sediment onshore into the prograded wavelain ridge series. The near point-source of the sediment from the delta indicates why the foredune area is effectively tied to the south end of Inch.

Whether these foredune changes can be accurately called mesoscale is a debatable point. There appears to be an extended period of some decades during which the foredunes are rebuilt and reworked by blowouts to generate the transverse grain to the area, however, the speed of foredune building is unknown though ground photograph evidence of foredune growth between 1984 and 1994 suggest that it is incremental and persistent on a decadal basis until some limit (e.g. accommodation space) is achieved. Although the spacing of the erosion events is mesoscale, they are caused by sudden triggers which might occur at any time, i.e. they do not necessarily have a mesoscale distribution. The extent to which antecedent conditions at Inch have to create a context of conditional probabilities by which the extreme triggers generate massive erosion is unknown. Such feedback may be of mesoscale proportions.

There is a further mesoscale action afforded by the foredunes at Inch due to their position as a buffer between beach and back beach. At times of foredune development the sediment resupply activity in the parabolic dunes of the back-barrier is reduced as the supply is reduced by foredune sealing. At times when the foredunes are absent, aeolian activity in the parabolic alleyways is operating. In this manner the
foredunes are a crude control on contemporary mesoscale changes to the main dunes on Inch Spit.

The initiation, development and loss of foredunes has been schematically set out in fig. 13. It is suggested that there is evidence of at least three main periods of foredune development/presence, each of which may have persisted for several decades, occupying essentially the same position on the spit. These periods appear to have been punctuated by major erosion incidents which have taken the shoreline well back into the back-barrier dunes. From luminescence determinations taken in the parabolic dunes of the main spit, much of this activity started about the mid-nineteenth century at a time when the shoreline was maximised in its landward position (WINTLE et al. 1998). Much of this parabolic dune activity may be episodically turned on/ off by the periods of foredune building shutting down the coastal entrances to the parabolic alleyways. Following each erosive phase, beach storage was presumed to grow by the exchange of sediment back from the ebb delta to the spit terminus. By inference, it would appear that the consequence of the erosive phase is a loss of beach sediment to the delta and then to the shoreface (or vice versa) by return storm flows or tidal current exchange. In pace with increased beach storage and upper beach progradation via wave-lain beach ridges, there is an aeolian exchange of sediment into foredunes which over time show an increasing transverse reworking by blowout development. The luminescence dating of the main dune line to the parabolic dune systems suggests that this type of oscillation is unlikely to have occurred pre-1800, however this should not deny that an initial period of foredune development might have occurred in the early nineteenth century to be terminated by the erosive phase responsible for the main erosional dune scarp first documented in 1842.

The schema identified in fig. 13 indicates a first approximate summary of foredune sequencing at Inch. Although we suggest that extreme storms of some severity are the initiating triggers to periods of total foredune loss, we are as yet uncertain as to the likelihood of periods of lesser foredune erosion as a consequence of 'seasonal' cut and fill where the periodicity of 'seasonality' may be of an interannual basis. Like-

![Diagram](image-url)

Fig. 13. Schematic view of inferred morphological changes of beach and dune over the last two centuries at Inch Spit.
wise we are uncertain as to the time basis by which the accretion volume in beach and dune storage exceeds a threshold volume which in its removal and depletion of the nearshore store, allows severe storm wave activity superimposed upon a major surge increment to directly erode such a substantial dune body. Clearly this volume is in its initial phases acting as a negative feedback control ameliorating the influences of storms. At some point the threshold is reached when the beach storage feedback status becomes positive and a storm of sufficient severity induces massive erosion.

Conclusions

The mesoscale periodicity (c 30–50 years) to the disappearance of foredunes at Inch should be more formally regarded as a microscale trigger operating at a mesoscale periodicity, where the trigger is an extreme storm which is capable of massive foredune erosion. Such an extreme storm is likely to be associated with a major surge element that is probably amplified by the funnelling nature of Dingle estuary. The relative effect of the extreme event may have been increased by the volume exchange of sediment from the protecting ebb-delta to the beach face (foredune extension in the intervening mesoscale period) thus oversteepening the beach face and contributing to larger waves getting closer inshore. Changes of storminess and sea-level identified in recent decades although not directly linked to erosion are likely to be of importance in the subsequent phases of foredune building. The human factor seems unlikely to be of any major importance in mesoscale foredune changes at Inch.

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