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Ch.22 Extreme events: impact and recovery

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Abstract:

Sandy coast changes on timescales from days to years and sometimes decades, primarily result from the erosion-recovery (im)balance that is controlled by the respective contributions of storms and recovery conditions. Over the last decade, our understanding and predictive ability of storm-driven erosion and subsequent multi-day to multi-annual recovery has greatly improved, notably thanks to long-term and rapid-response coastal monitoring programs. This chapter gives a broad overview of the definitions and processes that control storm-driven beach and dune erosion and subsequent (partial, complete or excess) recovery. Key conceptual concepts are illustrated using two well-documented case studies: response and (partial) multi-annual recovery from (1) over a severe winter period along the Atlantic coast of Europe characterised by unusually strong storm clustering episode; and (2) from a single severe storm with an anomalous wave direction along the southeast coastline of Australia. Finally, future perspectives and knowledge gaps in relation to impacts and recovery from extreme events are discussed.

Keywords: Coastal storms; beaches; coastal dunes; erosion; recovery; shoreline; subaerial beach

2322.1 Introduction

24 Coastal change on sandy coasts is driven by a wealth of processes interacting with each other across a wide
25 range of temporal and spatial scales (Stive et al., 2004; Cooper et al., 2004). The primary cause of major
26 episodic morphological variability at the coast are coastal storms, during which beach and dune erosion
27 (Castelle et al., 2015), cliff failure (Earlie et al., 2015) and transport up to even boulder deposits (Cox et al.,
28 2018) can be observed as a direct result of storm-driven wave action. Except along coastal embayments,
29 where longshore transport can dominate a shoreline change signal as a result of beach rotation (Ruiz de
30 Alegria-Arzaburu and Masselink, 2010), on most open coasts, storm-driven beach and dune erosion largely
31 involves cross-shore sediment transport (Yates et al., 2009; Splinter et al., 2014a). During storms, sediment
32 from the beach and nearby dune is transported rapidly seaward by bed return flow (undertow) (e.g., Hoefel
33 and Elgar, 2003). During post-storm conditions, wave nonlinearities slowly move sediment from the upper
34 shoreface back to the subaerial beach (Hoefel and Elgar, 2003; Ruessink et al., 2007; Dubarbier et al., 2015),
35 which can subsequently feed the eroded coastal dune through aeolian transport (Bauer et al., 2009). Longer-
36 term (i.e., decades and centuries) coastal change is largely affected by longshore processes (e.g., Hansen and
37 Barnard, 2010), sea level rise (Le Cozannet et al., 2019) and large-scale sediment budget (Cooper et al., 2001).
38 On shorter timescales, i.e. from weeks to years and sometimes decades, coastal change primarily results
39 from the erosion-recovery (im)balance that is controlled by the respective contributions of storms and
40 recovery conditions. Quantifying beach recovery is therefore as important as beach erosion, and is required
41 to increase our ability to understand and further predict coastal evolution on a wide range of timescales.

42 While a single, isolated, severe coastal storm can have a dramatic and long-lasting impact on the coast (Thom
43 and Hall, 1991; Harley et al., 2017), the combined impacts of a series of less severe storms can also lead to
44 severe coastal erosion (Ferreira, 2006). Indeed, coastal storm events sometimes occur in rapid succession
45 separated by a short time interval (~2-3 days), which is commonly referred to as a storm cluster, storm group
46 or sequence of storms. Storm clustering, for instance, is a frequent synoptic feature in the Euro-Atlantic area
47 with serial storm clustering occurring in both the flanks and downstream regions of the North Atlantic storm
48 track (Pinto et al., 2013). In addition, average storm event based on the probability distribution of wave
49 height can have a significant impact on the coast if coinciding with other hazards, such as spring high tide
50 and/or storm surge. It is therefore critical to address the *synchronicity* of environmental parameters to better
51 assess extreme events (Cooper et al., 2004; Guisado-Pintado and Jackson, 2018). As such, in the presence of
52 storm clusters, a positive storm surge may last for several days and the probability that it coincides with a
53 high tide or even brackets several tidal cycles, including a spring high tide, is greatly enhanced.

54 There has been considerable research interest in the impact of storms and storm groups on shoreline and
55 beach dynamics (e.g., Birkemeier, 1979; Zhang et al., 2002; Ferreira, 2005, 2006; Loureiro et al., 2012; Splinter
56 et al., 2014b; Coco et al., 2014; Karunarathna et al., 2014; Masselink et al., 2016a). Beach and dune recovery
57 after these events however, have received less attention by the coastal research community. This is despite

the fact that the timing and magnitude of recovery can provide useful proxy measures of coastal resilience, which is critical in the context of a rapidly changing climate. However, recently there has been increasing interest in the topic, with research providing new insights into patterns and timescales of recovery and modelling predictability (Houser et al., 2015; Scott et al., 2016; Castelle et al., 2017a; Philipps et al., 2017; Burvingt et al., 2018; Dodet et al., 2019; Phillips et al., 2019).

In this chapter, the definitions of extreme events, storm impact and beach/dune recovery are initially reviewed, highlighting a variety of concepts and thresholds and calling for clearer and more consistent definitions and communication to support understanding and management of extreme events. Secondly, we examine the means by which we can characterise storm impacts and recovery (considering both alongshore uniform and alongshore-variable responses). The third section describes recent examples of storm impacts and recovery on sandy coasts involving: response and (partial) recovery from: (1) a severe winter along the Atlantic coast of Europe characterised by exceptional serials of storm clusters and widespread coastal erosion; and (2) a single severe storm with an anomalous wave direction along the southeast coastline of Australia. In the last section, a summary and discussion is provided on future perspectives and knowledge gaps in our understanding of actual impacts and recovery from extreme storm wave events.

22.2 Definitions

22.2.1 Extreme events

In coastal sciences, like in other fields, there are many ways of deciding what defines an event ‘extreme’. In coastal research and engineering, there is no semantic difference between *storm* and *extreme storm*. Storm events along the coast can be characterised by variables such as storm magnitude (in terms of significant wave height and wind speed), storm direction, storm duration, tidal stage and water level surge relative to the storm event (Guisado-Pintado and Jackson, 2018). Other definitions can also include impacts (e.g., coastal dune erosion, storm demand, marine flooding), which blurs the distinction between extreme ‘event’ and extreme ‘impact’ (McPhilipps et al., 2018). Hereafter, the definitions of extreme events are essentially based on meteorological and oceanographic forcing variables.

On wave-dominated coasts, a storm is usually defined as an event in which the wave height exceeds a certain threshold H_T , known as the peak-over-threshold (POT) approach (Fig. 22.1a; see the review of Harley, 2017). Such a threshold has often been proposed subjectively (e.g. 2.5 m offshore of Sydney in Short and Trenaman, 1992; 6 m along the Portuguese coast in Ferreira (2005)) as a limit above which significant erosion is typically observed along the coast of interest. A more objective and transferable peak-over-threshold approach is the use of the probability distribution of the wave height, for example the 0.5% exceedance level (Luceno et al., 2006) or the 5% exceedance level (Castelle et al., 2015). Storm duration can therefore be defined as the duration over which wave height exceeds this threshold, but initiation and end of the event can also be defined as the time when wave height exceeds another quantile (e.g. the 25% exceedance level in Masselink

et al., 2014). The meteorological independence criterion (Harley, 2017) restricts the period of time between individual storm events, with anything shorter considered to be part of the same storm (Fig. 22.1a). This arbitrary criterion can vary from 30 hours (Almeida et al., 2012) to two weeks (Corbella and Stretch, 2012). Also arbitrarily defined is the maximum time for which two storms are considered part of the same cluster, for example this is defined as 9 days in Karunaranthna et al. (2014), 14 and 21 days in Ferreira (2005) and up to 39 days by Lee et al. (1998). The wide range of definitions and thresholds can produce dramatic differences in the number and duration of defined storm events (Fig. 22.1b).

Figure 22.1. (a) The Peaks-Over-Threshold (POT) method for defining storm events from a significant wave height time-series after Harley (2017) and (b) application to a real time series of significant wave height measured during the winter of 2013/2014 in about 50-m depth offshore of Truc Vert beach, southwest France after Castelle et al. (2015), using different arbitrary values of wave height threshold and independence criterion (I).

Storm wave conditions (defined by the above thresholds) coinciding with spring high tides and (positive) storm surge, or increases in incident wave exposure due to differing offshore wave directions, can also increase the coastal erosion hazard. The compound effects of waves and water levels can be assessed using complex process-based models, or by summing the tide level, the empirical vertical wave runup and storm surge estimation (Ruggiero et al., 2001; Young et al., 2016).

22.2.2 Beach and dune erosion

In basic terms, *beach erosion* is defined as a net loss of beach sediments over a particular vertical (2D) section of the beach profile and time scale of interest. This net sediment loss is manifested in a number of morphological signatures on the coast, including: a reduction in subaerial beach area; a landward migration of the shoreline; an overall lowering of the subaerial beach profile; presence of erosion scarps; and ultimately an undermining of dunes, cliffs and any back-beach infrastructure that may be present. Beach erosion is one of the most common impacts of extreme storm events, as elevated wave energy, in combination with winds, currents and elevated water levels drive sediment offshore from shallow to deeper waters. For particularly extreme events, this net loss of sediment might extend across the entire active beach profile i.e., from the

119 dune or cliff maxima down to a depth of negligible beach profile change known as the *depth of closure*. Such
 120 an immense loss of sediment may take decades for the beach system to recover from, if at all. In most cases
 121 however, storm events cause a simple redistribution of beach sediment, such that the subaerial beach
 122 experiences erosion while the subaqueous surf zone gains sediment lost from the subaerial beach. An
 123 example of a typical beach profile response from an extreme storm is presented in Fig. 22.2. In this example,
 124 sediment is redistributed by the storm waves from the subaerial beach to a single storm bar a few hundred
 125 metres offshore.

126
 127 *Figure 22.2. Example of beach and dune erosion resulting from an extreme storm: (a) pre and post-storm*
 128 *surveys of the entire beach profile, indicating the redistribution of sediment from the subaerial beach to an*
 129 *offshore storm bar; (b) close up of the subaerial beach response, highlighting the overall subaerial beach*
 130 *volume loss due to the storm ($\Delta V = 106 \text{ m}^3/\text{m}$ in this example) as well as the dune volume loss ($\Delta V_d = 12 \text{ m}^3/\text{m}$).*
 131 *Here the dune toe is defined by the 3 m elevation contour.*

132 Beach erosion due to storms is usually defined by the loss in beach sediment volume specifically above mean
 133 sea level (given in units m^3/m , refer Fig. 22.2b). For exposed, open-coast locations with large sandy dune
 134 systems, subaerial beach volume losses of up to 350 m^3 per alongshore metre of beach/dune system have
 135 been observed locally in the alignment of megacusp embayments (Castelle et al., 2017a). This subaerial
 136 volume change is often a good proxy for shoreline change (Robinet et al., 2016), since the majority of
 137 sediment lost from the subaerial beach occurs in the vicinity of the shoreline itself (Farris and List, 2007;
 138 Harley et al., 2011). The magnitude of subaerial beach volume change caused by a storm or storm cluster is
 139 an important variable in coastal engineering as it helps guide the determination of appropriate setback lines
 140 due to coastal storms (Callaghan et al., 2009). In planning coastal setback lines, it is typically assumed that a
 141 storm event or storm cluster with a given return period (e.g. a 1-in-100 year event) will result in an equivalent
 142 loss in subaerial beach volume known as the *storm demand*. Once this value is known, an appropriate buffer
 143 distance separating valuable infrastructure and residential properties from the shoreline can then be
 144 estimated (Kinsela et al., 2017).

145 Similar to subaerial beach erosion, dune erosion is a measure of the volume of dune sediment lost per
146 alongshore metre of beach (also in m^3/m , refer Fig. 22.2b). This value is typically defined by the volume of
147 dune sediment above the dune toe, whose elevation is site-specific and related to the tidal range, local wind
148 and wave conditions as well as sediment characteristics. The practical difficulties in accurately identifying the
149 dune toe means that this threshold is often defined by a representative elevation contour (e.g. the 3 m
150 contour as per Fig. 22.2b). While the dune volume change usually represents only a small fraction of the
151 overall subaerial beach volume change, it is of critical importance to evaluating coastal storm impacts, as
152 volume lost from the dunes due to storms can undermine nearby infrastructure (e.g. roads and residential
153 properties) and lead to major inundation of the coastal hinterland. Dune volume losses also take many times
154 longer to recover naturally compared to the rest of the subaerial beach profile (refer Section 22. 3.2 below).
155 Dune erosion also supplies sand to the surf zone and consequently reduces the demand and erosion of the
156 subaerial beach.

157 **22.2.2 Recovery**

158 In general terms, the word 'recovery' is defined as "a return to a normal state of health, mind or strength"
159 or "the action or process of regaining possession or control of something stolen or lost" (Oxford Dictionary
160 of English, 2010). This broad definition masks a considerable variability when it comes specifically to *beach*
161 *recovery*. A wealth of morphological indicators of beach recovery can be considered depending on
162 morphological settings and the cross-shore region of interest. Philipps (2018) reviewed all existing
163 morphological indicators. Amongst 16 different morphological indicators, the most commonly used are:
164 subaerial beach volume in the subaerial region (e.g. Birkemeier, 1979; Morton et al., 1994); shoreline
165 position, of which definition can vary, in the foreshore region (e.g. Philipps et al., 2017); dune volume (e.g.
166 Suanez et al., 2012) or dune crest height (Houser et al., 2015) in the backshore. Due to the difficulties in
167 monitoring the nearshore region, only a small number of studies have used a subaqueous morphological
168 indicator (e.g. offshore sandbar position) for beach recovery. Consideration of subaqueous indicators can
169 however, provide a broader description of beach recovery in addition to the more commonly observed
170 subaerial indicators (see Section 22.5).

171 Storm erosion and the post-storm recovery signal is affected by other natural modes of variability, such as
172 longer-term trends due to additional sediment sources or sinks (e.g., river mouth, inner shelf) or gradients in
173 longshore sediment transport. Building on observations along a 30-km stretch of coast in Texas, USA, Morton
174 et al. (1994) proposed four different post-storm recovery scenarios (Fig. 22.3): complete recovery (Fig.
175 22.3a); no recovery with continued erosion (Fig. 22.3b); partial recovery (Fig. 22.3c) and excess recovery (Fig.
176 22.3d) relative to pre-storm conditions. It is important to note that, within cycles of erosion/recovery, which
177 can be sub- to multi-annual, intermediate storm activity during recovery can result in smaller sub-cycles of
178 erosion and recovery in the subaerial beach. Erosion/recovery cycles can also be multi-decadal or even
179 longer, for instance on beaches adjacent to estuaries and tidal inlets (Castelle et al., 2018a). These systems

180 disrupt the longshore drift and sediment supply and exhibit cyclic and/or migrating behaviour (Ridderinkhof
 181 et al., 2016). Coastal lagoon openings are additional processes implying more complex and longer
 182 erosion/recovery cycles (e.g., Costas et al., 2005; Baldock et al., 2008).

183

184

185 *Figure 22.3. Scenarios of post-storm recovery from a given morphological indicator: (a) complete recovery;*
 186 *(b) no recovery; (c) partial recovery; (d) excess recovery. Modified after Morton et al. (1994).*

187 22.3 Characterisation of storm impact and recovery

188 22.3.1 Storm impacts

189 22.3.1.1 Storm impact regimes

190 The impacts of coastal storms are multi-faceted and span a range of socio-economic indicators (Ciavola et
 191 al., 2018). Focusing specifically on the hydro/morphological impacts of coastal storms, Sallenger (2000)
 192 recognised a gradation of impacts depending on the vertical (2D) extent of storm influence on the localised
 193 beach profile. Specifically, Sallenger (2000) proposed four *storm impact regimes* according to the maximum
 194 (R_{high}) and minimum (R_{low}) wave runup levels of the storm relative to the elevations of the localised dune
 195 crest (D_{high}) and dune toe (D_{low}). Maximum and minimum wave runup levels are defined in this impact scale
 196 by the 2% and 98% exceedance threshold levels, which can be calculated using the empirical wave runup
 197 equations of Stockdon et al. (2006) or others (e.g. Holman, 1986). Because of its ability to capture a range of
 198 hydro/morphodynamic processes related to storm impacts, the Sallenger storm impact scale has since been
 199 adopted as a standard measure of storm impacts on a range of coastlines (e.g. Armaroli et al., 2012; Plant et
 200 al., 2017). The four regimes – namely, *swash*, *collision*, *overwash* and *inundation*, are described in more detail
 201 below and presented schematically in Fig. 22.4.

202 1) *Swash regime* ($R_{high} < D_{low}$): this regime, representing the lowest of the four impacts regimes, occurs when
 203 storm wave runup is confined to the berm and beachface and does not interact with the dune profile. Since
 204 waves do not reach the dunes, the potential for erosion and inundation in this regime is theoretically limited
 205 to the lower subaerial beach (i.e., below the dune toe) and the storm impacts are considered relatively minor.

206 2) *Collision regime* ($R_{high} > D_{low}$, $R_{low} < D_{low}$): the collision regime occurs when storm wave energy and/or ocean
 207 water levels are sufficiently elevated that some waves collide with the seaward face of the dune. Depending
 208 on the duration and force of these collisions, as well as dune structure (e.g. geometry, composition and
 209 vegetation), this can lead to significant dune recession (i.e., a landwards displacement of the dune toe) and
 210 dune erosion (Larson et al., 2004; Palmsten and Holman, 2012).

211 3) *Overwash regime* ($R_{high} > D_{high}$, $R_{low} < D_{high}$): in storm conditions with particularly elevated water levels (e.g.
 212 large storm surge combined with spring high tides) and/or on low-lying coastlines, storm wave runup might
 213 exceed the dune crest and flow down the lee slope of the dune. This overwash process typically results in a
 214 major lowering of the dune crest, significant dune erosion and minor-moderate inundation of the coastal
 215 hinterland (Matias and Masselink, 2017). Overwash can also deposit sand many 10s to 100s of metres
 216 landwards of the dune crest as overwash fans and is responsible for the landward migration, or *rollover*, of
 217 barrier islands (topographic sedimentary barriers protecting the mainland coast from storm impacts) over
 218 time (Plant et al., 2017).

219 4) *Inundation regime* ($R_{low} > D_{high}$): the inundation regime represents the most extreme of the four impact
 220 regimes. In this regime, even the lower vertical bound of storm wave runup exceeds the localised dune crest,
 221 such that the entire beach profile is submerged during the storm. In such instances, the entire subaerial
 222 beach and dune system is exposed to energetic surf-zone processes, which can completely remove the dune
 223 system and result in catastrophic flooding of the coastal hinterland.

225 *Figure 22.4. Examples of the four storm impact regimes according to Sallenger (2000). R_{low} and R_{high} denote*
 226 *the minimum (98% exceedance) and maximum (2% exceedance) runup levels reached during each regime in*

227 *these examples. D_{low} and D_{high} represent the elevations of the dune toe and dune crest, respectively.*

228 *22.3.1.2 Localised three-dimensional impacts*

229 The storm impact scale described above represents a 2D framework based on the assumption that storm
230 impacts are uniform alongshore. However, storm impacts can be localised and/or show large alongshore
231 variability (Fig. 22.5), which can greatly complicate field assessment of storm impacts. For instance, localised
232 storm impacts are often observed along barrier islands that show a large range of geomorphic form (Otvos,
233 2012) and exposure to storms. As a result, barrier islands can undergo a wide variety of responses to storms
234 (see the review by Plant et al., 2017), which differ from that observed along other coastal settings. Storm-
235 driven overwash processes (Matias and Masselink, 2017), inlet opening and barrier breaching (Sherwood et
236 al., 2014) can have dramatic localised impact on the overall barrier morphology (Fig. 22.4a), and are key to
237 the long-term change in position and shape of the entire barrier island system (Plant et al., 2017). Storm
238 impact can also be locally increased (Fig. 5b) as a result of the presence of hard structure through scouring
239 (end) effects and/or because of downdrift sediment deficit causing the coast to become set-back (e.g. Brown
240 et al., 2011). Localised erosion is also almost systematically observed along eroding cliffs (Fig. 22.5c) although
241 in this case erosion is often not caused by storms but rather by the progressive cumulative effects of marine
242 and continental (physical and chemical weathering of cliff material) processes. Sometimes, beach and dune
243 erosion can show striking patterns of alongshore periodicity (refer Chapter 13 Rhythmic patterns in the surf
244 zone), with large megacusp embayments resulting in rhythmic cusped-type erosion scarps (Fig. 22.4d).
245 Megacusp embayments can be considered as the erosive signature at the shoreline of the presence of rip
246 channel currents (see Chapter 11 Rip Currents), which are reasonably regularly spaced self-organised
247 patterns in the sand (Falquès et al., 2000). These megacusp embayments can form in lee of major stationary
248 rips during severe storms when, during high tides, dune erosion can occur in the embayments where the
249 beach is the narrowest, as it is more vulnerable to undercutting by swash (Thornton et al., 2007; Castelle et
250 al., 2015).

251

252 *Figure 22.5. Examples of localised storm-driven erosion patterns. (a) Barrier breaching at Chandeleur Island,*
 253 *Louisiana, USA (Ph. Jim Flocks, USGS, src. www.earthsky.org); (b) accelerated beach and dune erosion next to*
 254 *a hard coastal structure at the end of the 2013/2014 winter at Lacanau, southwest France (Ph. J. Lestage); (c)*
 255 *Localised cliff erosion near Veulettes sur Mer in north France (Ph. S. L'Hôte); (d) rhythmic cusped-type erosion*
 256 *scarp along the Gironde coast, southwest France, at the end of the 2013/2014 winter (Ph. J. Lestage).*

257 **22.3.2 Beach and dune recovery**

258 Once destructive storm conditions subside, a period of constructive beach and dune recovery ensues. In this
 259 recovery period, the various hydro/morphological impacts related to the storm event (summarised above)
 260 are reversed to varying degrees, depending on the amount and spatial extent of sediment redistribution due
 261 to the storm as well as the prevailing recovery conditions. The temporal progression of beach and dune
 262 recovery follows a number of semi-discrete modes, each characterised by common morphodynamic
 263 processes. These recovery modes have been described piecewise by various authors (e.g. Wright and Short,
 264 1984; Morton et al., 1994; Hesp, 2002; Phillips et al., 2019), although to-date no holistic morphological model
 265 of the complete beach and dune recovery cycle exists (Phillips, 2018).

266 Starting with the recovery of storm deposits from the subaqueous and alongshore-uniform storm bar (e.g.
 267 Fig. 22.2), Wright and Short (1984) describe six morphodynamic beach states (Chapters 16 and 18) that the
 268 beach system transitions through as sediment is returned to the beachface. These six beach states, in order
 269 of 'least' to 'most recovered', are: the *dissipative* beach state; four intermediate beach states (the *longshore-*
 270 *bar-trough*, the *rhythmic-bar and beach*, the *transverse-bar and rip* and the *low tide terrace*); and, the

271 *reflective* beach state. This recovery cycle has been found (Wright and Short, 1985; Short 1999; Davidson et
272 al., 2013) to be strongly controlled by the progressive lowering of incident wave energy following the storm,
273 described in particular by the dimensionless fall velocity Ω (also known as the Dean parameter or Gourlay
274 number). Subsequent temporary increases in wave energy (e.g. during a successive, more moderate, storm
275 event) can interrupt this cycle, potentially resulting in a backwards transition towards a less-recovered beach
276 state and an overall prolonging of the entire beach recovery process. The end member of the Wright and
277 Short (1984) recovery cycle, i.e. the *reflective* beach state, is characterised by a steep beachface, a well-
278 established berm and minimal surf zone morphology. This end member however is rarely obtained on many
279 beaches, particularly those of finer grain sizes and/or in higher-energy wave climates. In these cases,
280 sediment may be returned to the beachface and berm from the complex coupling of multiple sandbars and
281 the shoreline.

282 Continuing the beach and dune recovery cycle, Morton et al. (1994) focused specifically on recovery of the
283 subaerial beach and dunes and developed a conceptual model based on only four recovery modes. The first
284 of these modes, described by Morton et al. as *berm reconstruction and forebeach accretion*, essentially
285 captures in an overarching manner the same wave-driven return of subaqueous storm deposits to the
286 beachface described in the morphodynamic beach state model of Wright and Short (1984). The second mode
287 in this Morton et al. recovery model represents a more advanced stage of recovery, whereby the backbeach
288 aggrades (i.e., grows vertically) as a result of wave overtopping of the recently re-established berm crest and
289 occasional aeolian deposition. The final two modes describe the ultimate stages of beach recovery, namely
290 the complete re-establishment of the dune system (Lynch et al., 2008) previously eroded by storm forces
291 (i.e., during the *collision*, *overwash* or *inundation* storm impact regimes described in Section 3.1.1 above).
292 These final modes are separated into an initial *dune formation* phase and a later *dune expansion and*
293 *vegetation recolonization* phase. A more comprehensive morphological description of foredune initiation
294 following storm removal is presented by Hesp (2002).

295 Rates of beach and dune recovery reported in the literature vary significantly and depend on the stage of
296 recovery over the complete cycle as well as the morphological indicator used to track recovery progression
297 (refer Phillips, 2018 for a comprehensive review). Houser et al. (2015) analysed recovery of barrier island
298 morphology following several large hurricanes and mathematically characterised recovery rates over the
299 complete recovery cycle by a sigmoid curve. This sigmoid curve describes a slow rate of recovery in the initial
300 stages, before rapidly increasing in the middle stages and slowing down again in the most advanced stages
301 of recovery. Morphologically, this curve reflects the initially slow transport of deepwater storm deposits from
302 the shoreface back to the surf zone, followed by a more rapid transfer of sediment from the inner surf zone
303 to the beachface and berm and, finally, the slow growth and revegetation of the foredune(s). Recent high-
304 resolution observations of beach recovery rates largely agree with the sigmoid curve model of Houser et al.
305 (2015). Scott et al. (2016), for example, found that the initial recovery of deep-water storm deposits was

multi-annual and relied on moderate storm waves from successive winters to return sediment onshore towards the beachface. Using 10 years of daily shoreline observations from a coastal imaging system, Phillips et al. (2016) also showed that recovery of the shoreline was initially slow on average (approximately 0.1 m/day for the study site) when storm deposits were detached from the beachface, but rapidly increased (~0.4 m/day) as sediment welded to the beachface and promoting berm progradation. Shoreline recovery was then observed to slow considerably again to approximately 0.05 m/day as the more advanced stages of recovery (berm aggradation and dune re-establishment) took hold (Morton et al., 1994).

22.4 Example observations from Europe and Australia

22.4.1 Impact and recovery from the 2013-14 winter along the Atlantic coast of Europe

During the 2013/2014 winter, the combination of a very intense polar vortex and unusually strong North Atlantic jet stream caused a succession of deep, low pressure systems to cross the North Atlantic (Davies, 2015) and reach western European coasts. The combination of high cyclone frequency and above-average cyclone intensity resulted in considerable impacts across the entire European Atlantic seaboard, down to Morocco (e.g., Thorne et al., 2014; Castelle et al., 2015; Blaise et al., 2015; Autret et al., 2016; Masselink et al., 2016a, 2016b; Burvingt et al., 2018; Cox et al., 2018). Although a positive phase of the North Atlantic Oscillation (NAO), which reflects an intensification of the latitudinal pressure gradient between the Azores High and the Icelandic Low (Hurrell, 1995), is supposed to enhance storminess in the North Atlantic, that particular winter was associated with an average positive NAO. Instead, the latitudinal atmospheric dipole of pressure anomaly strengthened but more importantly shifted 15° southward, which is represented by the West Europe Pressure Anomaly index (WEPA, Castelle et al., 2017b). That winter was characterised by the highest winter average wave height since at least 1948 (Masselink et al., 2016a), and likewise the highest WEPA. Large waves were generated along the entire coast of Europe (Fig. 22.6a), down to northwest Africa, with the anomaly peaking at +1.62 m at approximately 50°N, which corresponds to an approximately 40% winter mean wave height increase along the entire Bay of Biscay (Fig. 22.6b). There was no particularly exceptional storm occurrence during that winter, instead the clustering was itself outstanding, and particularly from mid-February 2015 to early March 2014 (see the zoom onto the period January 20 – March 6, 2014 in Fig. 22.1b). The number of storms and the total storm duration during the 2013/2014 winter was, depending on definition, generally at least 100% larger than the second most energetic winter since at least 1948 (Masselink et al., 2016a).

Figure 22.6. Winter (DJFM) 2013/2014 distribution of (a) winter-averaged significant wave height and (b) percentage relative to the long-term (68 years) winter average. (c) Location of the Atlantic coast beaches studied in Masselink et al. (2016a) and Dodet et al. (2019) with PT = Portrush (Northern Ireland); PP = Perranporth (southwest England); SP = Slapton Sands (southwest England); VG = Vougot (northwest France); PM= Porsmilin (northwest France) and TV = Truc Vert (southwest France).

Fig. 22.8 shows the time series of beach volume and dune toe position at the Atlantic coast study sites shown in Fig. 22.7c, with outstanding storm erosion impacts during the 2013/2014 winter leaving the majority of the sites in their most depleted state since measurements began. The most exposed sites Perranporth and Truc Vert lost in excess of 80 and 200 m³/m of subaerial beach volume, respectively, and such storm response was observed to be typical of most exposed beaches along the coast of southwest England and France. Contrasting responses occurred at the more sheltered sites such as Porsmilin and Vougot. At Slapton Sands, the middle profile (SP₁₀) experienced a subaerial beach volume loss of 100 m³/m, whereas accretion of a similar amount occurred at the north profile (SP₁₈) owing to unprecedented beach rotation (Masselink et al., 2016b). Portrush did not experience any erosion, as it was reasonably sheltered from the storm waves during that winter.

Fig. 22.7 shows that the recovery signature is site-specific and multi-annual. Only one of the studied beaches fully recovered after two years in terms of beach-dune volume (Truc Vert beach, Fig. 22.7g). However, it is important to highlight that shoreline recovery was not complete, as most of the sand that came back primarily fed the beach rather than the incipient foredune after 2 years (Castelle et al., 2017a), therefore embodying the first *berm reconstruction and forebeach accretion* mode of beach recovery according to the model of Morton et al. (1994). During the subsequent 2 years incipient foredune formation was observed, suggesting that Truc Vert is about to enter the *dune expansion and vegetation recolonization* phase. Three sites only partially recovered with large difference in magnitude after two years (60% and 90% at Perranporth

359 and Porsmilin, respectively). Patterns are also different, with the eastern profile at Slapton (SP₁₈) showing no
 360 subaerial volume recovery, like at Vougot although partial shoreline recovery is observed at this latter site. A
 361 more detail assessment of the recovery process (Dodet et al., 2019) shows that beaches recover during the
 362 spring–summer–autumn period at modest and relatively steady rates (not much inter-annual variability).
 363 However, it is the energetic winter conditions that primarily control the time it takes for beaches to recover
 364 from extreme erosion as energetic winter conditions stall the recovery process whereas moderate winter
 365 conditions accelerate it, which further emphasise the strong link between the dominant wave climate indices
 366 (WEPA and NAO) and coastal response (severe erosion and multi-annual recovery).

367
 368 *Figure 22.7. Time series of beach volume at the six European beach study sites (with two profiles shown for*

369 *Slapton Sands) with the winter of 2013/2014 indicated by the light grey area. For Vougot and Truc Vert the*
370 *evolution of the location of the dune toe (grey line) is also shown. Modified after Dodet et al. (2019).*

371 **22.4.2 Impact and recovery from the 2016 east coast low in SE Australia**

372 In June 2016, a severe extratropical storm known as an east coast low or cyclone, impacted the coastline of
373 SE Australia. While storms of this type are not unusual for this coastline and occur on average 55.5 days per
374 year (Pepler et al., 2013), the June 2016 event was notable for its unique synoptic characteristics whereby
375 the low pressure system combined with a blocking high in the South Tasman Sea to result in a large (>2000
376 km) and relatively stable north-easterly wind fetch directed at the coastline for several days. Waves
377 generated from this system peaked at a deepwater significant wave height between 4.5-8.5 m along the
378 coastline (Mortlock et al., 2017), which is equivalent only to a modest average recurrence interval on this
379 coastline of between 1-in-2 to 1-in-5 years (Shand et al., 2011). Crucially though, the unusual fetch created
380 by this synoptic pattern meant that these waves were from an anomalous easterly wave direction relative to
381 the modal southerly swell and storm waves that dominate this coastline. This anomalous wave direction had
382 the effect of significantly enhancing the exposure sections of beaches to anomalously high incident storm
383 wave energy, since southerly waves are usually attenuated to a significant degree by large rocky headlands
384 that dominate the southern extremities of beaches on this embayed coastline (Short, 2007). The impacts of
385 this event were also exacerbated by the fact that the storm coincided with winter solstice spring tides.

386 A rapid-response coastal monitoring program captured high-resolution Airborne Lidar measurements both
387 immediately prior to and following the June 2016 storm along 178 km of embayed sandy coastline (Harley et
388 al., 2017a). These measurements recorded an impressive 11.5 M m³ of subaerial beach erosion due to the
389 threeday event over this survey region, equivalent to an average subaerial volume loss of 65 m³/m (maximum
390 recorded localised volume loss = 228 m³/m). Beach erosion was enhanced approximately fourfold on sections
391 of sandy beach embayments directly exposed to the incident easterly waves, relative to more sheltered
392 southerly and northerly-oriented beach sections (Harley et al., 2017a). At the Narrabeen-Collaroy long-term
393 coastal monitoring station, where beach measurements have continued uninterrupted at monthly frequency
394 since 1976 (Turner et al., 2016), beach erosion due to this event was found to be the largest in four decades
395 (Harley et al., 2017a). This monitoring period included storm events of much larger deep-water significant
396 wave heights, but from more usual southerly wave directions, highlighting the critical importance of storm
397 wave direction on embayed coastlines such as those of SE Australia.

398 Monitoring using Airborne Lidar over the same 178 km coastal stretch in the 12 months following the storm
399 found that recovery was initially rapid, with 49% and 66% of the total subaerial volume lost during the storm
400 returning in the first three and six months alone. More detailed inspection of this recovery observations
401 reveal that this return of sediment was confined to the berm and beachface, embodying the first *berm*
402 *reconstruction and forebeach accretion* mode of beach recovery according to the model of Morton et al.
403 (1994). Similar to the erosion response of the storm itself, incident wave exposure was also found to

404 significantly control rates of recovery on this embayed coastline. In the first six months of beach recovery,
 405 the most rapid recovery rates were observed at sites directly exposed to the milder southerly waves that
 406 characterised wave conditions over the initial recovery period. At the same time however, these more
 407 exposed locations also made them more vulnerable to a subsequent southerly storm event typical of this
 408 coastline, causing a reversal in beach recovery in these regions (Harley et al., 2017b). The culmination of
 409 these effects was that despite the rapid initial recovery, most beaches exhibited only partial recovery over
 410 the 12-month period following the storm. More-sheltered embayed beach stretches generally indicated
 411 more gradual, but steady, beach recovery, whereas more-exposed locations exhibited very rapid initial
 412 recovery, but also significant interruptions and reversals in recovery due to a subsequent storm event.
 413 An example of the beach erosion and recovery cycle due to the June 2016 storm in SE Australia is presented
 414 in Fig. 22.8. This figure charts the evolution of beach erosion and recovery in the vicinity of the Narrabeen-
 415 Collaroy coastal imaging station (Harley et al., 2011; Turner et al., 2016). In this example, the subaerial beach
 416 is observed to recover in just 11 months (Fig. 22.8i,j) and subsequently indicates excess recovery that reaches
 417 a maximum at the 13.5 month mark.

418
 419 *Figure 22.8. Temporal evolution of beach erosion and recovery at Narrabeen-Collaroy Beach (SE Australia)*
 420 *due to the June 2016 east coast low storm. a) immediately pre-storm; b) mid storm (day 1); c) mid storm (day*
 421 *2); d) immediately post-storm (day 3); e) +3 month recovery; f) +6 month recovery; g) +11 month recovery; h)*
 422 *+13.5 month recovery; i) surveyed beach profiles at PF6 during different stages of the erosion/recovery cycle;*
 423 *j) time-series of monthly subaerial volume evolution at profile PF6 following the storm. All images are taken*

424 *at mid-tide stages.*

425 **22.5 Future perspectives and knowledge gaps**

426 Deficits in large-scale sediment budgets (see Chapter 23 Coastal sediment compartments, wave climate and
427 sediment budget) combined with increases in extreme coastal water levels (e.g. Marcos et al., 2019) and/or
428 extreme wave heights (e.g. Castelle et al., 2018b; Young and Ribal, 2019) result in increased coastal erosion
429 hazards. In addition, the *risk* of coastal erosion has been observed to increase in some regions primarily
430 because of increased exposure of assets to coastal storms (Lazarus et al., 2018). Better understanding and
431 further mitigation of coastal risk is therefore one of the greatest challenges for future coastal scientists,
432 engineers and managers. Such advances in understanding require the development of a better, holistic,
433 definition of storm events, storm-driven erosion and subsequent recovery. As we have seen in this chapter,
434 existing definitions cover a variety of concepts, thresholds and coastal compartments, which often limit the
435 broad scale application of findings regarding storms and storm impacts. Generic qualitative definitions of
436 storm events such as that proposed in Masselink et al. (2014) must be encouraged. Such an objective
437 definition based on the wave climate (the probability distribution of the significant wave height using the 5%
438 and 25% exceedance levels) is recommended for generic use. Similar definitions based on, for instance, the
439 magnitude of storm impact to the subaerial beach volume, must also be developed to build a suite of generic
440 quantitative definitions which will allow for robust inter-site comparisons.

441 The large natural variability in coastal response and recovery presented in this chapter and through the two
442 examples in Europe and Australia demonstrates the value of coastal monitoring programs implemented
443 across a wide range of representative sites with different geological settings, tidal regimes and degrees of
444 wave exposure (Guisado-Pintado and Jackson, 2018). These observational records are crucial to
445 understanding and further developing the necessary techniques to manage and predict coastal behaviour
446 across the full range of coastal systems which vary considerably across the globe, regionally and even locally.
447 However, the cost of implementing such programs has meant that long-term, continuous records of coastal
448 change are rare and have tended to focus more on the subaerial domain (Turner et al., 2016). There is a need
449 therefore to monitor the complete erosion/recovery cycle from the shoreface to the beachface - and up to
450 the dune where present. This will enable accurate quantification of the coastal sediment budget and further
451 examine inter- and multi-annual variability including extreme storm erosion and post-storm recovery (e.g.
452 Scott et al., 2016; Ruiz de Alegria-Arzaburu and Vidal-Ruiz, 2018). Remote sensing techniques, such as depth
453 inversion from video images (Holman et al., 2012), satellite-derived bathymetry (Pacheco et al., 2015) and
454 large-scale data inferred from satellites (Luijendijk et al., 2018; Vos et al., 2019) are starting to provide more
455 insight into long-term coastal changes (see Chapter 26). Such large-scale data must be combined with short-
456 term shoreface and surfzone measurements (Aagaard, 2014; see Chapter 27) and seabed characteristics and
457 sediments to better understand the sediment exchanges and pathways between the different compartments
458 (dune, subaerial beach and lower shoreface) and, in turn, the processes controlling storm erosion and

459 subsequent recovery (Kinsela et al., 2017).

460 Numerical modelling is a promising avenue to hindcast, understand and ultimately predict the response of
461 beaches to storm and their subsequent recovery. These models can be classified into three categories,
462 namely process-based, hybrid and data-driven models. Process-based models, which rely on a detailed
463 description of the dominant hydrodynamics and sediment transport processes, are powerful tools to describe
464 storm-driven erosion on small scales (<km) pending calibration (e.g. Roelvink et al., 2009). However, these
465 models show limited skill to simulate beach changes on larger temporal and spatial scales, typically
466 associated with recovery periods. In contrast, simpler hybrid models which are based on general principles
467 (e.g. behavioural law) can be successfully applied over larger temporal and spatial scales and, in turn, be used
468 to investigate shoreline erosion and recovery processes in more detail. Within this hybrid model category,
469 Dodet et al. (2019) showed that equilibrium models (e.g., Yates et al., 2009; Davidson et al., 2013; Splinter et
470 al., 2014a) have significant skill in reproducing the shoreline erosion during the 2013/2014 winter described
471 above and the subsequent (partial or complete) recovery on the most exposed, cross-shore transport
472 dominated sites. At other sites, such as in coastal embayments where the rotation signal may dominate (refer
473 Chapter 24 Beach rotation), more complex hybrid shoreline models must be developed and applied. Such
474 models have been recently developed (Vitousek et al., 2017; Robinet et al., 2018; Antonilez et al., 2019) which
475 open up new perspectives to improve our understanding and predictive ability of shoreline response to storm
476 and recovery in complex hydrodynamic and geological settings. A third model category, referred to as data-
477 driven models range from simple autoregressive models to machine learning techniques (e.g. Beuzen et al.,
478 2018; refer to Chapter 28 Machine learning and coastal processes). In a recent effort of the coastal research
479 community (Montaño et al., 2019), a “blind” comparison between approximately 20 state-of-the-art hybrid
480 and machine learning shoreline models has been performed. Most of these models indicated good skill in
481 capturing seasonal changes (seasonal recovery cycles) but often failed to predict large changes or changes
482 occurring at shorter timescales (storm-driven erosion). Multi-model ensemble (i.e., aggregating predictions
483 across different models) is also shown to improve the reliability of the predictions, providing additional
484 information on model-related uncertainty. Hybrid and data-driven models therefore need to be further
485 improved and tested against other datasets with different hydrodynamics and geological settings, which once
486 again motivate the development of ambitious coastal monitoring programs.

487 Over the last decade, our understanding and predictive ability of storm-driven erosion and subsequent multi-
488 day to multi-annual recovery has greatly improved. Although there is still considerable room for further
489 improvement, there is also a need for a better coordination between output from the research community
490 and what is translated to end-users/practitioners (refer Chapter 29 Applying beach morphodynamics to
491 management). This is reflected by the recent development of early-warning systems for coastal flooding and
492 erosion hazards, typically providing warning a few days ahead (e.g. Vousdoukas et al., 2012; Harley et al.,
493 2016). In addition, it is now well-established that extreme coastal wave climate is strongly affected by large-

494 scale climate patterns of atmospheric variability (refer Chapter 3 Wave climates: deep water to shoaling
495 zone). Given the strong correlation of certain climate indices on winter wave climate and coastal response in
496 different regions of the world (Barnard et al., 2015; Dodet et al., 2019), the ability of climate models to predict
497 these dominant climate indices a few months ahead will also be crucial to anticipate potential impacts
498 months or years in advance (e.g. Davidson et al., 2017). It will also provide supplementary information for
499 predicting the potential for recovery and thus ultimately help coastal management approaches.

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