GEOLOGICAL RECORDS OF TSUNAMIS AND OTHER EXTREME WAVES

EDITED BY
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Preface

This edited volume compiles the state of the art in research on the geological record of tsunamis and other extreme-wave events and guides the reader in designing goal- and site-specific research. It has evolved from an initial idea, first explored by the editors in early 2016, to final publication online and in print in mid-2020. The motivation for developing a handbook-type compendium on this topic was driven by the observation that such a unifying volume devoted to this particular discipline, which lies at the crossroads between sedimentology and tsunami science, was missed by the scientific community. What we had in mind was an exhaustive work that enables the broader dissemination and transfer of ideas, methods and concepts associated with identifying tsunami and other extreme-wave deposits. By doing so, we seek to promote their application to a wide range of different coastal sedimentary environments and their enhanced use for coastal hazard assessment.

The great success of our first thematic session "Geological records of extreme wave events" organized at the European Geosciences Union (EGU) General Assembly in 2016 was a clear demonstration that there was an active community of researchers who were enthusiastically pushing the tsunami geoscience field forward. A special issue of the journal *Marine Geology* related to this EGU session followed in 2018 (Vol. 396, edited by Ed Garrett, Jessica Pilarczyk, and Dominik Brill) compiling 16 papers on paleo- and modern tsunami and storm records. With a wide range of exciting new research being presented at subsequent editions of the EGU session, we felt that a detailed compendium would be of significant interest for the continuously growing community. Consequently, this work represents a true community effort: leading experts were invited to contribute chapters, while each chapter was peer-reviewed by at least one external reviewer and a minimum of one of the editors. It is great to see the substantial overlap between the authors and reviewers of this compendium, and the contributors to the thematic sessions at the annual EGU General Assemblies.

Two existing edited books, both well established in their scientific communities and regarded as benchmark literature resources, have inspired and guided the concept of the present work. *Tsunamiites* (Elsevier/Amsterdam) of 2008, edited by Tsunemasa Shiki and colleagues, provides an exhaustive overview on the aspect of tsunami sedimentology. It also gathers some of the most prominent figures in this field as authors, but in contrast to the present book, it combines textbook-type chapters with case studies and has a clear emphasis on the older, pre-Quaternary geological record. The *Handbook of Sea-Level Research* (Wiley/Chichester) of 2015, edited by Ian Shennan and colleagues, follows a proxy-by-proxy structure, with detailed methodological information to guide research on reconstructing relative sea-level histories. Such a structure focusing on operational workflows, methodological details, opportunities and limitations associated with specific proxies has been adopted and built upon in the present compendium.

A special merit is dedicated to the reviewers of the individual chapters of this book. The external evaluations by expert members of the community were essential to the final quality of this volume and are, therefore, gratefully acknowledged. We sincerely thank the following persons:

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Above all, we thank the authors of the individual chapters for contributing their tremendous expertise and precious time. To fulfill the purpose of this book, each chapter combines expert knowledge on a wide range of different aspects of paleotsunami research with systematic guidelines and recommendations for their implementation in research. The very positive responses that we received to invitations sent to potential authors around the globe may be an additional indication of the timeliness of this book and is hopefully a signal that it will provide a useful addition to the body of literature on tsunami geoscience.

Furthermore, we are thankful to everyone at Elsevier, who accompanied the entire process from the initial idea to printing and distribution of the book in a professional and efficient way.

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Max Engel, Jessica Pilarczyk, Simon Matthias May, Dominik Brill, Ed Garrett
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CHAPTER

Erosive impact of tsunami and storm waves on rocky coasts and post-depositional weathering of coarse-clast deposits

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Abstract

The spatial distribution of boulder deposits along rocky coastlines provides important implications for estimating the hazard of extreme waves (storms or tsunamis). However, rocky coasts are highly dynamic environments, and their changes through time have to be considered when analyzing the coarse-clast record and inferring characteristics of past events. This chapter reviews the impact of extreme waves on the erosion of rocky coasts and the formation of coarse-clast deposits, evaluates the potential of weathering and erosion features for relative dating of coastal erosion and boulder transport, and investigates the relevance of boulder "shrinking" through post-depositional erosion processes.

Keywords: Boulders; Cliff-top deposit; Coastal geomorphology; Relative age dating; Tsunami deposit.

Introduction

Rocky coasts are shaped by both gradual, long-term, as well as episodic high-energy wave action. The intensity of coastal transformation and the rate of coastal retreat are a function of wave climate, bathymetry and coastal topography, rock type and structure, as well as epi- and endolithic fauna and flora (Benumof and Griggs, 1999). Additionally, relative sea-level variations shift the level of highest energy transfer between sea and land. The erosion of cliffs and coastal platforms of all latitudes produces clasts of different size and shape, which are potentially transported

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and deposited onshore in distinct patterns during high-energy waves of severe storms or tsunamis (Williams and Hall, 2004; Scheffers, 2005; Goto et al., 2010; May et al., 2010, 2015; Etienne et al., 2011; Fichaut and Suanez, 2011; Nandasena et al., 2011; Engel and May, 2012; Terry et al., 2013; Scheffers et al., 2014; Erdmann et al., 2017, 2018b; Piscitelli et al., 2017; Cox et al., 2018; Biolchi et al., 2019; Fig. 26.1A–D). Thus, such coastal coarse-clast deposits are spectacular testimony to the hazard of extreme waves, and their size and spatial distribution may provide pivotal information on long-term magnitude–frequency relationships, in particular on maximum levels of marine flooding over millennial timescales (Etienne et al., 2011; Engel and May, 2012; Terry et al., 2013). On the one hand, investigations on cliff-top blocks and boulders require the consideration of cliff transformation and cliff retreat, in particular on Pleistocene timescales (e.g., Rovere et al., 2017; Mylroie, 2018; Erdmann et al., 2018b), since transport distance and local topographic conditions during the event are crucial for their interpretation. On the other hand, as the rocky shoreline is constantly subjected to weathering and erosion, the boulders are also modified and tend to reduce their size over centuries and millennia of subaerial exposure. Since most approaches of inversely modeling the extent and characteristics of coastal flooding based on the coarse-clast record rely on the size, shape, and density of individual boulders, their modification may directly influence the modeling outcome and, thus, bias hazard assessment. In addition to direct dating approaches using common dating techniques such as ^{14}C-AMS, U-series, and electron spin resonance (ESR) (e.g., Terry et al., 2013; Scheffers et al., 2014; Biolchi et al., 2016; Rixhon et al., 2018), important relative information on the timing of clast emplacement can also be derived from weathering patterns of the coarse clasts, the intensity of which is generally a function of age, supratidal position, and lithology. This is particularly the case for limestone coasts, since post-depositional geomorphological indicators for boulder transport such as secondary rock-pool formation or karst features are best developed in the carbonate realm (Matsukura et al., 2007; Engel and May, 2012; Terry et al., 2013, Erdmann et al., 2017; Fig. 26.1E).

In this chapter, we (i) review the impact of extreme storm waves and tsunamis on the erosion of rocky coasts and the formation of coarse-clast deposits; (ii) evaluate the potential of weathering and erosion features for relative dating of coastal erosion and boulder transport; and (iii) try to quantify the erosive modification of subaerial coastal boulders through time.

Erosive impact of tsunamis on rocky coasts

Rocky coasts by their very nature are destructive environments with a comparably low preservation potential of geomorphological evidence for extreme-wave (and other) events. The main agents of rocky shoreline disintegration are weathering as well as gradual (long-term, normal wave conditions), periodic (e.g., storm waves), and episodic (e.g., strong tsunamis) extreme-wave action. Given their low frequency, the contribution of tsunamis appears negligible as it is mostly overprinted by gradual

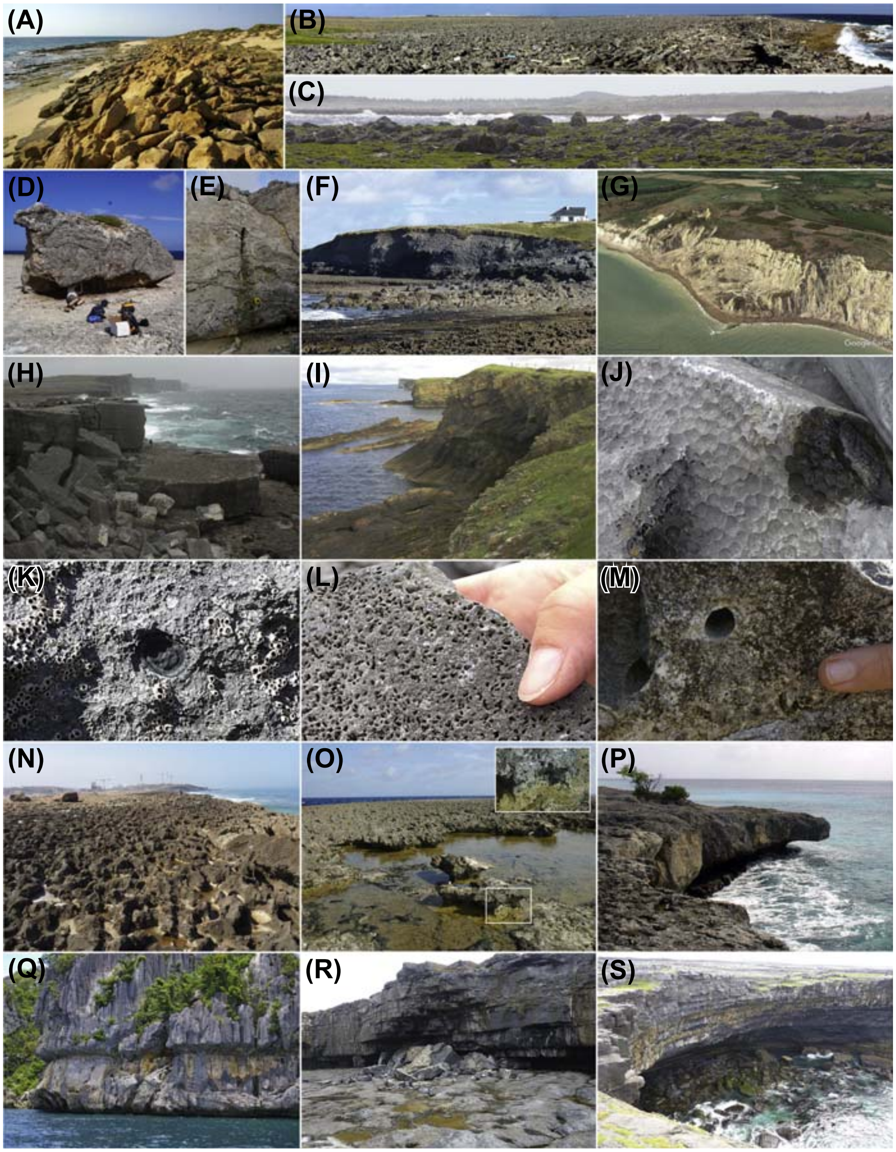

FIGURE 26.1

Selection of coastal coarse-clast deposits as well as weathering and erosion features discussed in the text. (A) Boulder ridge in low supratidal position dominated by flat, imbricated clasts, resting on an inclined shore at Cape Range, Western Australia. (B) Up to 400 m wide, multimodal rampart of *Halimeda* sands, other skeletal debris, as well as massive slabs of *Acropora palmata* along the windward coast of Bonaire (Washikemba area) (cf. Scheffers, 2005; Scheffers et al., 2014). (C) Boulder field on top of a reefal and periodic wave action. Unsurprisingly, studies on tsunami erosion along rocky coasts are very rare and only few sources of wider information and discussion exist (e.g., Bryant and Young, 1996; Bryant et al., 1996; Aalto et al., 1999). These studies associate potholes, keels, flutes, and grooves along rock coasts with megatsunami impacts, comparable to similar landforms known from extreme flood events like meltwater-lake outbreaks. However, they are mostly descriptive and conceptional and have been challenged by several authors (e.g., Felton and Crook, 2003; Courtney et al., 2012), bringing up convincing arguments against a tsunami origin of these erosional features.

The recent megatsunamis in the Indian Ocean in 2004 and along the coast of Tōhoku, Japan, in 2011 provided some limited insights into tsunami erosion on rocky coasts. Paris et al. (2009) report erosional signatures around Lhok Nga Bay in northern Sumatra, such as slabs and boulders broken off from the cliff, regolith

limestone platform, c. 4–5 m above mean sea level (windward coast, Bonaire) (cf. Scheffers, 2005; Engel and May, 2012; Scheffers et al., 2014). (D) Isolated, overturned boulder resting on top of an elevated limestone platform along the northern coast of Bonaire (Boka Onima area). Rainwater seeping through the boulder has led to the growth of microbialites on the surface of the downward-facing, inactive rock pools, which were sampled for ^{230}U/Th dating to derive minimum ages for boulder transport (Rixhon et al., 2018). (E) Reefal limestone boulder on Bonaire (Boka Onima area) developing karst pipes and other post-depositional weathering features as relative age indicators (Engel and May, 2012). (F) Cliff formation in an LGM (Last Glacial Maximum) glacial till deposit, associated with a gently inclined shore platform in Galway Bay, western Ireland. (G) Subaerial slump along a chalk cliff, England (50°52′08″N, 0°38′48″E). The image shows an 85 m-high cliff with protalus bulge and strong back-cutting with a slide at an elevation of +127 m in chalk rock. The scene is 1 km wide. (H) Receding cliff (18 m high) at the wave-dominated limestone coast of Inishmore, Aran Islands, western Ireland, with very large cliff-top blocks in the background. (I) Cliff shape influenced by folding of sedimentary Paleozoic rocks along Shetland Mainland (Scotland). (J) Sea-urchin indentations on a dislocated limestone boulder in Galway Bay, western Ireland. (K) Limpet (*Patella* sp.) indentations on a dislocated limestone boulder in Galway Bay, western Ireland. (L) Borings of the sponge *Cliona* sp. on a limestone boulder in Galway Bay, western Ireland. (M) Boreholes of *Lithophaga* sp. in massive reefal limestone of a cliff-top boulder on Bonaire (Rixhon et al., 2018). (N) Rock-pool belt along a low calcareous sandstone cliff south of Rabat (Atlantic coast of Morocco). (O) Very wide rock pool on the windward coast of Bonaire, actively shaped by *Littorina* gastropods (see insert). (P) 1.5 m-deep Holocene bioerosive notch in reefal limestone at the leeward coast of Bonaire. (Q) Holocene and last interglacial bioerosive notches at the limestone coast of Gigante Sur Island, Central Visayas, Philippines. (R) Collapse of an overhanging limestone cliff, undermined along a shale stratum, at the south coast of Inishmore (Aran Islands, western Ireland). (S) Overhanging cliff in Carboniferous limestone resulting from higher resistivity compared to the intertidal shale rocks (Inishmore, Aran Islands, western Ireland).

(G) Credit: Google Earth 4/2015. All photographs owned by the authors.

stripped from bedrock slopes, and widened natural trenches. Generally, the lack of high-resolution pre-tsunami topographic data prevents the quantification of erosion, as also exemplified by a laser scanning-based study of surface change after the 2011 Tōhoku Tsunami along the northern ria-type Sanriku coast (Hayakawa et al., 2015). Extreme runup of up to 40 m in the narrow drowned valleys removed the regolith cover of bedrock slopes and is estimated to have eroded exposed bedrock surfaces in the order of millimeters to centimeters. Taking into account the frequency of strong tsunamis in the region over Holocene timescales, these effects add up to meters or even tens of meters and are considered to be a significant landscape-shaping factor in the lower ria valleys of the Sanriku coast (Komatsu et al., 2014; Hayakawa et al., 2015). In the same area, Nandasena et al. (2013) document erosional scars on steep slopes along with numerous locations where boulders were quarried from exposed bedrock surfaces.

However, based on our currently limited understanding of the effects of tsunamis on the rocky coastal realm, major parts of this contribution cover storm-wave impacts and their geomorphological consequences, which have been subject to a considerably larger number of investigations.

Cliff destruction: episodic versus long-term effects

Mechanisms of cliff retreat: the significance of lithology, gravity and marine forcing

The formation of cliffs mainly depends on the resistance of the cliff rock, which is in continuous competition with the hydraulic forces of approaching waves (Sunamura, 1992). Coastal erosion and cliff recession is of particular concern along coastlines prone to mass wasting (Allison, 1989; Budetta et al., 2000; Lee, 2008; Kline et al., 2014), and most examples stem from those of less resistant rocks such as chalk, marls, Neogene marine sediments, or glacial till (Foote et al., 2006; Hénaff et al., 2006; Dornbusch and Robinson, 2011; Moses and Robinson, 2011; Hurst et al., 2016; Fig. 26.1F and G). In more resistant rocks, folding, faulting, or bedding patterns determine pathways of percolating rainwater and/or wave-injected seawater, ultimately controlling the effects of freeze and thaw cycles, rock dissolution, and weathering and disintegration intensities (Allison, 1989; Hampton et al., 2004; Kennedy and Dickson, 2007; Biolchi et al., 2019; Fig. 26.1H and I).

Limestone coasts (Lace and Mylroie, 2013, and contributions therein) are particularly prone to bioerosive rock-surface sculpturing of mostly centimeter to decimeter scale (rarely meter scale), which occurs at inter- to supratidal elevations (Fig. 26.1J–Q) and ultimately contributes to cliff transformation. Local collapse of bioerosive notches may occur, which, however, is only of minor importance in cliff recession. Rock-surface sculpturing results from grazing by gastropods (littorinids, limpets) and chitons on endolithic blue-green algae (Cyanobacteria, Chlorophyceae), and occurs only at the wave- and splash-affected sections of rock surfaces (Schneider, 1976; Spencer, 1988; Kelletat, 1997). The most significant bioerosive forms along limestone coasts are rock pools on gently sloping or horizontal rock platforms (Fig. 26.1N and O). Rock pools develop through lateral enlargement of small initial pits by littorinid gastropods, when pools are large enough to contain water over longer periods of time. Notches occur along steeper cliff sections, with larger limpets (*Patella* sp.) or chitons typically involved in their formation (Kelletat, 1988, 1997; Kogure and Matsukura, 2010; Fig. 26.1P and Q.).

Where gravity overcomes internal rock cohesion, cliff collapses occur. The intensity and characteristics of these gravitational processes depend on the angle of the cliff face, cliff-base morphology (plunging cliff or rock platform), the coherence of the cliff rock, and the kinetic energy of the moving rock mass, but also on climate and hydrology that drive weathering processes. Cliff recession is generally cyclic, with intermittent slides, slumps, or rock falls, followed by the removal of rock debris or sediment aprons from the rock platform. Periods of wave erosion at the cliff base generally alternate with periods where talus sediments protect the cliff toe. For the British chalk coasts, the residence time of cliff-base talus has been found to be 10–40 years, and for the more resistant chalk in the environs of Étretat along the coast of Normandy about 25–50 years (Foote et al., 2006; Hénaff et al., 2006; Dornbusch and Robinson, 2011; Kline et al., 2014; Fig. 26.1G). More resistant limestone (e.g., along the western Ireland cliffs, Aran Islands, Galway Bay) is typically affected by the collapse of upper cliff sections, where joints are widened by subaerial dissolution or where undermining of limestone units by erosion of underlying less resistant rock types, e.g., shales, occurs (Erdmann et al., 2018b; Fig. 26.1R and S). In addition, cave systems along limestone coasts may represent locations of intensified cliff collapse (Hampton et al., 2004).

The efficiency of marine impact in terms of erosion depends on bathymetry, wave climate, and the exposure of a site, which determine the character and form of waves and the impact of wave action. Continuous wave impact with cyclic build-up and release of wave pressure reduces the strength of cliff rocks over time, even if no sediments for mechanical abrasion are available. At cliffs with very resistant rocks, where no sediments are entrained in the wave, hydraulic action—both swell waves and turbulent bores—may be the only destructive force, supported by compressed air in cracks and joints (Noormets et al., 2004). These cracks are gradually widened and predefine clasts, which are either quarried and transported toward the cliff top during extreme-wave conditions (e.g., Noormets et al., 2004; Fichaut and Suanez, 2011; Engel and May, 2012; Erdmann et al., 2017, 2018b; Kennedy et al., 2017) or end up as debris at the cliff toe through gravitational processes (Bird, 2016; Figs. 26.1A–D and 26.2A–C'Figs. 26.2A). Noormets et al. (2004) consider a minimum of 60% of fracturing necessary to enable detachment of large clasts from the cliff edge by the impact of breaking swell waves, even though their power is in most cases not sufficient to transport the clasts onto the cliff top. Herterich et al. (2018) identify bending stress imposed by the wave impact and filling of pre-existing cracks to result in the propagation of microcracks up to complete fracture and quarrying. Maximum dynamic shock pressures applied to the edge of the shore platform may only last 0.1 s (Earlie et al., 2015) and result

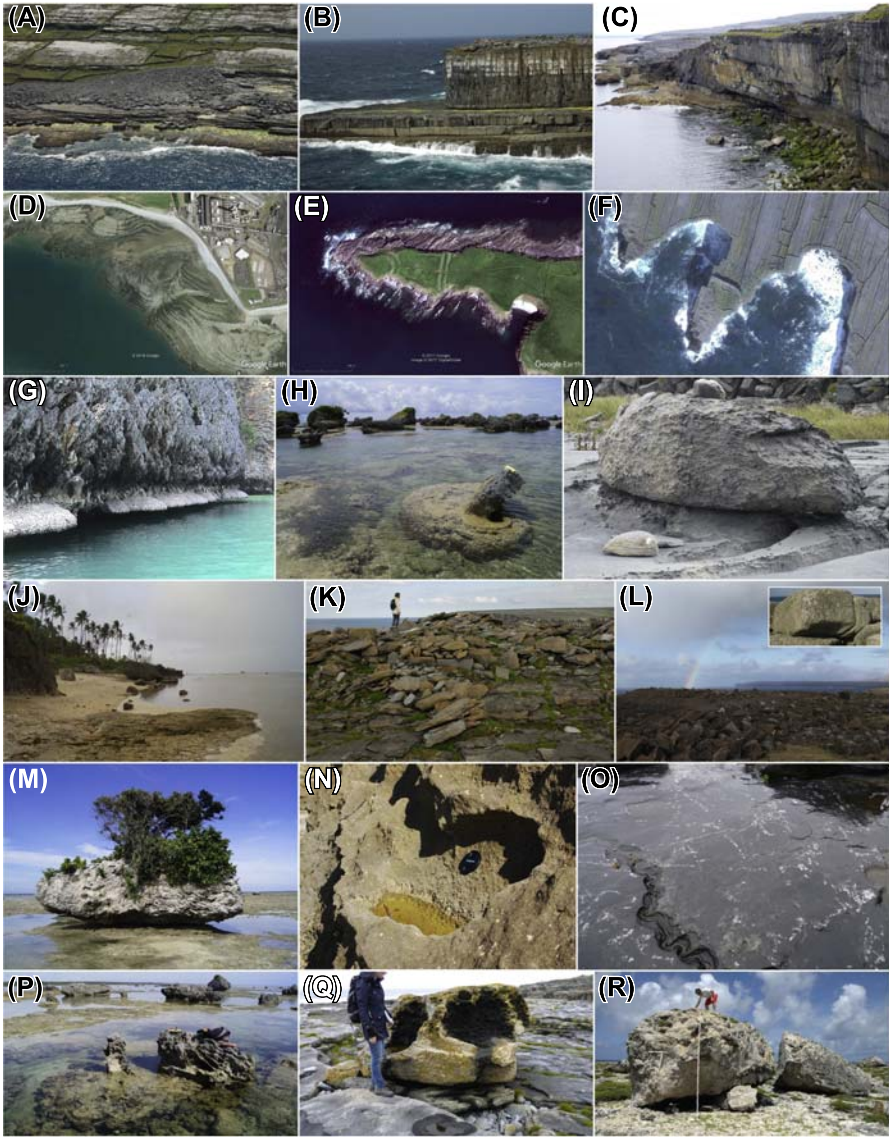

FIGURE 26.2

Selection of coastal coarse-clast deposits as well as weathering and erosion features discussed in the text. (A) Pattern of cliff retreat in horizontally bedded Carboniferous limestone, with onshore dislocated boulders (Inishmaan, Aran Islands, western Ireland). (B) Steep cliff formation in highly resistant limestone along the wave-dominated coast of Inishmore (Aran Islands, western Ireland). (C) Well-rounded large talus boulders at a cliff base as remnant of a past rockfall (limestone, inside Galway Bay, western Ireland). (D) Wave-cut platform in folded limestone rocks near Southerndown, Wales, UK. (E) from relatively short-period (∼8 s) and steep deep-water waves (height/length = 0.06–0.04). While such shorter waves possess higher quarrying capacity due to the higher pressure exerted on the face of the cliff as compared to those of longer waves, the capacity to transport these boulders landward increases with wavelength and is highest in a tsunami bore (Noormets et al., 2004).

Over time, rock platforms may develop in front of cliff toes due to wave-driven back-cutting of the cliffs (Stephenson, 2001; Foote et al., 2006; Hénaff et al., 2006; Dornbusch and Robinson, 2011; Regard et al., 2012; Fig. 26.2D and E). Since the nearshore water depth directly influences height, power and inundation depth of waves, rock platforms in turn diminish wave energy and wave erosion. Furthermore, sea-level fluctuations and tectonics play a crucial role in the evolution of wave-impact patterns, cliff morphologies, and their retreat.

Ballineden, County Sligo, Ireland, at 54°′N and 8°41′W; this promontory fort is now still 160 m long. The remains at its tip date from Neolithic times, whilst it was enlarged during Iron Age (500 BCE to turn of eras). (F) Iron Age promontory Fort Dun Duchathair ("Black Fort") at +26 m on a limestone promontory of Inishmore, Aran Islands, western Ireland (Google Earth 5/2010). (G) Bioerosive notch with an up to 40 cm-wide bench constructed by the rock oyster *Saccostrea cucullata*, exposed during low water along the tower karst islands of southwest Thailand. (H) A wave-emplaced fossil microatoll became the nucleus of an active fringing microreef inside the moat of an elevated intertidal reef platform of Eastern Samar, Philippines, representing a rare example of post-depositional size increase. (I) Pedestal resulting from protection of an erratic boulder on limestone and denudation of the cliff top since deglaciation after the LGM (Inishmore, Aran Islands, western Ireland). (J) Low pedestal formed beneath a massive limestone block (a-axis 9 m, weight c. 180 t) before it was shifted to its present position during Typhoon Haiyan in 2013 (Eastern Samar, Philippines) (May et al., 2015). (K) Dense cover of lichen (*Caloplaca marina*) on an old limestone boulder ridge inside Galway Bay, western Ireland. (L) Landward-increasing lichen growth (light spots) on a cliff-top boulder ridge near Eshaness, Shetland Islands' Mainland, and dense lichen cover on wave-emplaced boulders in Ireland (see insert). (M) Limestone boulder on top of an uplifted (emerged) intertidal reef platform in Eastern Samar (Philippines), heavily overgrown by bushes and small trees. (N) Two rock pool generations in a tilted, wave-emplaced limestone boulder at Cape Bon, Tunisia. The bottom of the original, inactive rock pool is steeply inclined, while since deposition another horizontal one has formed, indicating that quarrying and transport occurred stepwise some time ago (May et al., 2010). (O) Striations (scratch marks) in different directions indicating contact of boulders with limestone bedrock during saltating or rolling transport (Erdmann et al., 2018a). (P) Extreme post-depositional karstic weathering pattern of a wave-emplaced limestone boulder on top of an uplifted (emerged) intertidal reef platform in Eastern Samar (Philippines). (Q) Post-depositional solutional cavities and dense lichen cover on a wave-emplaced limestone boulder at the south coast of Inishmaan, Aran Islands, western Ireland. (R) Wave-emplaced boulder split into two pieces during or after deposition on an elevated reefal platform on the windward coast of Bonaire (Scheffers, 2005; Engel and May, 2012).

(D) Credit: Google Earth. (E) Credit: Google Earth 5/2010. All photographs owned by the authors, apart from map data (D–F): Google Earth, Digital Globe.

Intensities of cliff development and recession

In general, the spacing of joint patterns best correlates with erosion rates, i.e., a narrow joint pattern typically results in fast erosion and cliff retreat. Cliff recession and rock platform development occurs due to the continuous impact of both normal, fair-weather waves, and more episodic events such as high-energy storm waves or tsunamis. By measuring the episodic changes in cliff-top failure and calculating related mass-wasting cubatures using terrestrial laser scanning or drone-derived orthophotos and digital elevation models, cliff retreat during single events may be estimated, which can be significantly higher than the long-term values (Table 26.1)

Table 26.1 Comparison of calculated cliff-retreat rates from different rock types and exposures, measured with different methods.

Region	Rock type	Methods (survey period)/data source	Rate (cm/year)	Reference
England	Hard rock	Terrestrial Laser Scanning/LiDAR (16 months)	1–7	Rosser et al. (2005)
Worldwide	Soft rock	Not specified	>100	Bird (2016)
Worldwide	Hard rock	Not specified	<1	Bird (2016)
Mediterranean, Black Sea	Limestone	Not specified	6–80	Furlani et al. (2014)
Mediterranean, Black Sea	Marble	Not specified	0.3	Furlani et al. (2014)
England	Varying	Long term	2–45	Earlie et al. (2015)
England	Varying	LiDAR	1–29	Earlie et al. (2015)
New Zealand	Volcanic	Rock platform width	1–2	Kennedy and Dickson (2007)
England	Chalk	Long term, ^{10}Be	2–6	Hurst et al. (2016)
England	Chalk	Long term, historical	5–133	May (2005)
Aran Islands	Limestone	Boulder–ridge mass	0.15–3	Erdmann et al. (2017)
World	Hard rock	Rock platform width	0.9–10	Erdmann et al. (2018b)
England	Chalk	Cliffs	30–50	Moses and Robinson (2011)
England	Chalk	Rock platform width	0.3–0.4	Moses and Robinson (2011)

An exhaustive compilation of older, pre-1990 data can be found in the Appendix of Sunamura (1992).

(e.g., Rosser et al., 2005; Regard et al., 2012; Katz and Mushkin, 2013; Earlie et al., 2015; Bird, 2016; Cullen et al., 2018). For instance, based on multi-temporal LiDAR measurements, Katz and Mushkin (2013) found that over a 13-month period with one exceptional winter storm, 70% of the total erosion of a weakly cemented aeolianite cliff at the coast of Israel occurred during the storm event and the four subsequent months.

However, extrapolations from such monitoring studies covering the range of years to a maximum of a decade may only be representative if local environmental conditions remain similar for the extrapolated period. Consequently, short-term direct measurements of changes along the cliff face may contrast with cliff-retreat rates over centennial to millennial time scales derived by extrapolating mass-wasting volumes and frequencies. Using the volume of onshore cliff-sourced coarse-clast deposits is also inappropriate, since only a variable fraction of cliff-detached material is transported onshore, while most of it ends up in the foreshore by gravitational processes (e.g., Allison, 1989; Budetta et al., 2000; Rosser et al., 2005; Hall et al., 2008; Regard et al., 2012; Earlie et al., 2015). The application of terrestrial cosmogenic radionuclides such as ^{10}Be may bridge this gap, as it allows for estimating long-term denudation rates as well as spatial and temporal changes directly at the cliff face (Hurst et al., 2016).

Likewise, the width of (abrasive) rock platforms in front of cliffs may be used to determine cliff retreat rates, which generally depend on wave climate and nearshore bathymetry, exposure, coastal topography, and rock characteristics (e.g., Stephenson, 2001; Foote et al., 2006; Hénaff et al., 2006; Dornbusch and Robinson, 2011). Rock platforms are relatively wide in coastal regions with frequent and strong storms, but narrow or absent at tropical latitudes with extreme but only occasional storms such as tropical cyclones. A 200 m-wide rock platform—if developed over 6000 years of a relative sea-level highstand—may indicate an average rate of cliff retreat of about 3.5 cm/year, which is in the same order of magnitude as rates directly measured at sedimentary rocks (Table 26.1). However, since friction and shoaling diminish wave energy on larger rock platforms, cliff recession will become slower over time, if sea level and tidal range remain similar, challenging their indicative value for cliff-recession rates.

Archaeological hints for coastal and cliff-retreat rates

Besides geochronological or high-resolution monitoring approaches, (pre-)historic man-made deposits and/or buildings such as shell middens, earthworks, forts, castles, or lighthouses (archaeologically, historically, or radiometrically datable; e.g., Bromhead and Ibsen, 2006), placed in delicate positions along cliffs, may provide clues to cliff-erosion patterns and rates (Fig. 26.2E and F). Some may even date back to the Neolithic and offer constraints on the position of coastlines over the entire period of the postglacial sea-level highstand. On the Aran Islands (western Ireland), however, it is debated if an Iron Age fort located on a promontory and close to the cliff edge indicates either low rates of cliff retreat (Scheffers et al., 2009;

Erdmann et al., 2018b) or fast cliff erosion if it was once founded as a ringfort further inland (Williams, 2004; Hansom, 2005; Fig. 26.2F). As more than 100 constructions of this age (about 2500–2000 BP) still exist in strongly exposed promontory settings along western Irish shorelines characterized by resistant sedimentary and igneous rocks, at least over this time span cliff retreat was probably on the order of only a few cm/year (Table 26.2). The same conclusions can be drawn from the preservation of glacial erosive landforms in the littoral environment and slope-over-wall cliff types (see examples in Bird, 2016).

Table 26.2 Rates of cliff retreat as measured from small-scale karst morphologies (Crete), size and preservation of notches (Crete, Thailand), and (pre-)historical promontory forts (W Ireland, N Ireland).

Region	Rock type	Time frame (years)	Max. total distance (m)	Total average rate (cm/year)	Reference
Inishmaan (Ireland)	Limestone	16,000	100	0.6	Erdmann et al. (2018b)
Aonghasa (Ireland)	Limestone	6000	5	<0.1	Erdmann et al. (2018b)
Inishmore (Ireland)	Limestone	6000	40	0.7	Erdmann et al. (2018b)
Crete (Greece)	Limestone	1650	0.05	0.003	Kelletat (1996)
Crete (Greece)	Limestone	1650	0.1	0.006	Kelletat (1996)
Crete (Greece)	Limestone	120,000	10	0.008	Kelletat (1996)
Crete (Greece)	Aeolianite	1650	0.2	0.01	Kelletat (1996)
Phang-nga (Thailand)	Limestone	6000	1	Max. 0.02	Scheffers et al. (2012)
Phang-nga (Thailand)	Limestone	120,000	2	Max. 0.02	Scheffers et al. (2012)
Western Ireland	Iron Age fort; limestone, greywacke	2500	Max. 25-50	Max. 1-2	Scheffers et al., unpublished data
North Ireland	Medieval castle; basalt	800	5	0.6	Scheffers et al., unpublished data

Rates of rock weathering and dissolution

The intensity of bioerosion in pure limestone is comparably high, often occurring at a rate of ≥1 mm/year (Kelletat, 1988; Spencer, 1988). The formation of 2 m-wide rock pools and 1 m-deep notches from bioerosion thus requires >1000 years, indicating quasi-stability of the coast for a similar time span.

On the other hand, calcareous algae like *Lithophyllum* sp., *Lithothamnium* sp., *Neogoniolithon notarisii*, vermetids (tubeworms) like *Vermetus* sp., *Dendropoma petraeum* or *Galeolaria caespitosa*, or rock oysters (e.g., *Crassostrea amasa*) may construct hard and protective covers (crusts, rims, or trottoirs) or even real biohermata on coastal rocks with growth rates reaching >1 mm/year (Kelletat, 1997; Fig. 26.2G and H). Their strict vertical zonation forms a sharp boundary between the levels of (maximum) bioerosion and bioconstruction. If dated by ^{14}C or $^{230}Th/U$, such incrustations also allow for quantifying the intensity of erosion over time.

The rate of surface lowering of massive limestone by dissolution in a terrestrial environment is rather similar in different climates and amounts to about 0.02–0.03 mm/year on average (Häuselmann, 2008; Table 26.3). In inter- to supratidal coastal environments, where downwearing is usually measured by using standardized carbonate weight-loss tablets (Spencer, 1988), micro-erosion meters (MEM) (over months or a few years) (Kelletat, 1988; Furlani et al., 2009; Cullen et al., 2018), structure-from-motion photogrammetry (Cullen et al., 2018), or surface exposure dating by cosmogenic nuclides (in limestone: ^{36}Cl) (Rixhon et al., 2018), the contribution of bio-agents leads to significant variations in downwearing rates, ranging from low values similar to terrestrial environments up to several mm per year (Spencer, 1988). By setting the height of rock pedestals below glacial erratics (150–300 mm) in relation to the timing of deglaciation in the Aran Islands of western Ireland (~16 kyrs ago), Erdmann et al. (unpublished) found a rather low rate of 0.01–0.02 mm/year, ranging slightly below those found elsewhere in similar environments (Goldie, 2005) (Fig. 26.2.I).

For Kikai-jima, SW Japan, a mean lowering rate of 0.205 mm/year was inferred over the last 6000 years for the reefal limestone platform around pedestals protected by large wave-transported boulders (Matsukura et al., 2007) (cf. Fig. 26.2J). In the Caribbean, such pedestals of up to 0.7 m have formed on Bonaire (Engel and May, 2012) and Barbados (Scheffers and Kelletat, 2006). Denudation rates of exposed supralittoral limestone terraces in the Caribbean were estimated to be in the range of 0.01–0.02 mm/year on Curaçao (Focke, 1978) and 0.021 ± 0.005 mm/year for the same landforms on the neighboring island of Bonaire (Rixhon et al., 2018). Donn and Boardman (1988) report remarkably high rates of surface lowering in supralittoral limestone of Andros Island, Bahamas, of up to 1.4 mm/year. In a soft siltstone environment along the more temperate Portuguese coast, Oliveira (2017) found rates between 0.03 and 0.2 mm/year by multi-year micro-erosion meter measurements. Furlani et al. (2014) compiled values of intertidal to supratidal limestone bedrock lowering in the Mediterranean basin, ranging from 0.001 to 2 mm/year. In general, young coral limestones with a high porosity should result in higher lowering rates (and pedestal dimensions) compared to more dense and resistant Mesozoic or Paleozoic limestones.

Table 26.3 Compilation of terrestrial dissolution rates on exposed limestone (MEM = micro-erosion meter).

Region	Rock type	Method	Time frame (kyrs)	Dissolution rate (mm/year)	Reference
England	Mesozoic limestone	Erratic pedestals	18	0.027	Sweeting (1966)
England	Mesozoic limestone	Erratic pedestals	18	0.002–0.0086	Goldie (2005)
Japan	Pleistocene reef rock	Erratic pedestals	Few	0.1–0.3	Matsukura et al. (2007)
Ireland	Paleocene limestone	Erratic pedestals	16	0.0093–0.0186	Erdmann et al., unpublished
Italy	Eocene limestone	MEM	Few	0.01–0.035	Forti (1984)
Italy	Tertiary limestone	MEM	Few	0.02–0.03	Cucchi et al. (1995)
Australia	Eocene limestone	MEM	Many	0.006–0.013	Smith et al. (1995)
Alaska	Mesozoic limestone	MEM	Many	0.04	Allred (2004)
Switzerland	Mesozoic limestone	MEM	Many	0.007–0.021	Häuselmann (2008)
Italy/ Croatia	Tertiary limestone	MEM	Many	0.009–0.018	Furlani et al. (2009)
Croatia	Tertiary limestone	MEM	Many	0.01–0.03	Taboroši and Kázmér (2013)
Norway	Paleozoic limestone	Tablets	Few	0.03	Lauritzen (1990)
Austria	Mesozoic limestone	Tablets	Many	0.01	Plan (2005)
Svalbard	Paleozoic limestone	Tablets	Few	0.004–0.035	Krawczyk and Pettersson, 2007
Japan	Mesozoic limestone	^{36}Cl	Many	0.02–0.06	Matsushi et al. (2010)
England	Mesozoic limestone	^{36}Cl	Many	0.0333	Vincent et al. (2010)
China	Mesozoic limestone	^{36}Cl	Many	0.017–0.047	Xu et al. (2013)
France	Mesozoic limestone	^{36}Cl	Many	0.035	Zerathe et al. (2013)
France	Mesozoic limestone	^{36}Cl	Many	0.03–0.04	Godard et al. (2016)

Data from different continents, elevations, and climates are rather congruent on dense old limestones and significantly higher only in reef rocks with a different texture and higher porosity.

Relative age estimation for boulder transport

While cliff retreat in less resistant rocks may be too fast for the long-term preservation of cliff-top deposits, rocks with high resistivity tend to develop rather steep cliffs that are stable over long time periods and erode with very low rates (Tables 26.1 and 26.2). In such settings, the accumulation of coarse clasts has led to the formation of a variety of deposits since the deceleration of global eustatic sea-level rise in the mid-Holocene. Most of these coarse-clast records—mostly ridges, ridge sequences, ramparts, or boulder fields (Figs. 26.1A–D and 26.2A and B)—can be found along rocky shorelines with low to moderately high and particularly stepped cliffs, or along coastlines with medium slope angles or elevated intertidal reef platforms (Chapter 24). However, whether their formation results from continuous accumulation throughout the Holocene or from a few extreme-wave events often remains under debate.

Wave-emplaced (cliff-top) boulders are exposed to various exogenic processes after deposition. Where direct dating of marine organisms attached to the clasts (e.g., boring bivalves, vermetids; Fig. 26.1K–M) is not possible or problematic, relative age indicators are at least as valuable and important. These relative age indicators are directly related to exposure time (i.e., age of clast dislocation) and include geomorphological (rock-pool generation, karst features, impact marks) as well as biological (soil or lichen and other vegetation cover) evidence (Fig. 26.2K–N).

However, weathering patterns are rock- and site-specific. For instance, fragmentation by frost weathering may refresh forms of clasts over rather short time intervals. On the other hand, granites and similar crystalline rocks are subjected to a more rapid surface disintegration in warm and humid climates compared to cold ones.

Vegetation, lichen cover and microbialites

Few studies have systematically used lichen growth (Fig. 26.2K and L) as a function of time to infer the age of a boulder deposit. Oliveira (2017), who provides a detailed explanation on the methodological approach, collected lichen structures (*Opegrapha durieui*) from rock surfaces of known age along the Portuguese coast to establish a local growth curve applicable to lichen findings on boulder deposits associated with extreme-wave impacts. In her study, boulder transport is dated from years to centuries, even though error margins may exceed ±30%.

In an area with rapidly growing lichens (*Verrucaria maura*, *Caloplaca marina*, *Lecanora* sp.) along the high-energy coasts of Ireland and Scotland, Hall et al. (2008) choose a more qualitative approach and estimated boulder-ridge inactivity on decadal scales based on lichen cover. For instance, on old volcanic rocks of the Shetland Islands, age estimates are based on observations that a 50–100% rock-surface cover of black *V. maura* takes a minimum of 70 years (Fig. 26.2L). Similarly, Scheffers et al. (2009) and Erdmann et al. (2017, 2018b) use dense lichen carpets on limestone (*Caloplaca marina*) as evidence for decadal to centennial inactivity of single boulders or boulder ridges on the Aran Islands, Ireland.

In a pioneering attempt to cross-date boulder deposition on Bonaire, Dutch Antilles, Rixhon et al. (2018) identified carbonate microbialite inside a former rock pool at the bottom side of an overturned boulder, which owes its formation to carbonate-enriched rainwater seeping through the boulder and a moist and warm microclimate at the shielded underside. Despite impurities of gypsum coatings and organic remains, the Mg-calcite-, and aragonite-dominated microbialite was conclusively dated to c. 1230 years ago by ^{230}Th/U dating, representing a useful alternative approach in dating coarse-clast transport of carbonates (see also Chapter 31).

Where debris lines or even boulder belts from singular extreme runup events cover existing higher vegetation or their remnants (e.g., along Lituya Bay, Alaska; Miller, 1960), this may also provide age constraints. Inactive boulders may become heavily overgrown by vegetation (e.g., boulder #1 in Frohlich et al., 2009), providing minimum ages for transport (Fig. 26.2M).

Rock pools and other bioerosive indicators

Among the most evident geomorphological relative age indicators are pre- and post-depositional rock-pool generations, which are best recorded in clasts composed of massive limestone (Fig. 26.2N). In many cases, cliff-top clasts derive from the (splash- and spraywater-influenced) inter- to supratidal cliff-edge sections along rather low-lying cliffs, where rock pools shaped by littorinids are present. Where limestone boulders are detached and removed from these locations, their biogenic formation immediately stops, and the rock-pool surface is then subjected to dissolution by rainwater (see examples in May et al., 2010; Engel and May, 2012). Although much less intensive compared to bioerosion (0.02–0.03 mm/year on average), carbonate dissolution changes the characteristic round and closed form of bioerosive rock pools, e.g., by dissolving a new horizontal basin (second generation) inside the tilted rock pool (first generation), or by the formation of an overflow channel draining the old rock pool (Kelletat and Schellmann, 2002; Erdmann et al., 2018b; Fig. 26.2N).

The cross-dating approach of Rixhon et al. (2018) on Bonaire allows for a comparison of dimensions of post-depositional karst features with radiometric data. For instance, a vertical dissolution pipe of a width of several cm (Fig. 26.1E; see also Fig. 26. in Engel and May, 2012) or a second rock pool generation, a few cm deep and c. 20–30 cm wide (see Fig. 26. in Engel and May, 2012), has evolved in boulders likely inactive for at least 1500 years (Rixhon et al., 2018).

Striations on rock may be used to identify impacts or changes over one to twodecades, in particular during post-event inspections (Fig. 26.2O). Observations over several years additionally allow classifying these intensities, site- and rock-specifically (Erdmann et al., 2018a).

Long-term modification of coastal boulders

Weathering of onshore clasts is difficult to quantify as it depends on structure, texture and mineral composition, exposure to precipitation, radiation and wind, stability of setting, age, and other factors. Theoretically, non-carbonates can be tested with a Schmidt Hammer or similar devices for the resistivity of their outer parts (Goudie, 2006), while carbonates are subjected to rainwater dissolution dominating all other weathering processes. Its intensity mostly depends on individual facies composition and density.

Erosive processes summarized earlier taking their toll on coastal boulder deposits have to be considered if the size of a boulder is used to infer flooding characteristics of past high-energy wave events (Nandasena et al., 2011; May et al., 2015; Biolchi et al., 2016; see Chapters 28 and 29). Even though these effects can only be roughly quantified for reefal limestone boulders exposed to inter- or supratidal conditions over centuries to millennia, as possibly in various Caribbean cases (Scheffers, 2005; Scheffers and Kelletat, 2006; Scheffers et al., 2014; Rixhon et al., 2018), size reduction of boulders may amount to decimeter scales in all three main axes, if denudation rates previously cited are taken into account. In temperate environments, where boulders have been identified to remain stable over longer periods, such as the Aran Islands (Erdmann et al., 2018b), these losses are expected to be smaller.

Recent observations, however, made on an elevated reef flat in Eastern Samar, Philippines (Boesl et al., 2019), reveal examples where late Holocene boulders in intertidal position have almost entirely lost their subaerial part to (microbio- or abrasive) erosion (Fig. 26.2P and Q), while others seem relatively unaffected. The degree of weathering in this setting is likely a function of time, vertical position, distance to the platform edge, texture of reefal lithofacies, availability of sediments, and local density of the boulder cluster. In some cases, coral growth even leads to a size increase of clasts (Fig. 26.2H). In addition, boulders may break up during transport or deposition (Fig. 26.2R) (Scheffers, 2005; Engel and May, 2012; Piscitelli et al., 2017).

Conclusions

After decades of research on coastal coarse-clast deposits, methods and approaches in data acquisition are still somewhat heterogeneous. Boulder properties (rock type, form, size, mass, spatial distribution etc.) are mostly in the focus, accompanied by statistical analyses and interpretations, as well as inverse and forward numerical models of the transport mode and process. In many cases, however, information on the dynamic landscape surrounding the coastal boulders (relief, surface characteristics, exposure, nearshore bathymetry, and many more) is neglected. The hunt for the largest boulder may remain as the most important result, and speculations on transport processes, especially in the case of "megaboulders" from "superstorms" find their way even into new interpretations of climate change and coastal hazards in the future. Another deficit in coastal boulder research may be its concentration on a time frame of the younger Holocene, while reconstructions of much older processes and deposits with a much longer history of transformation are in their infancy (see discussion in Hansen et al., 2016; Rovere et al., 2017; Mylroie, 2018).

Identifying boulder transport inland and against gravity by waves in the first order is a question of geomorphology rather than sedimentology. Elevation and relief including nearshore bathymetry, rock type and age, wave climate, sea-level history, cliff-retreat processes and rates, hints to the age of boulder deposits; post-depositional erosion and other factors have to be considered for conclusions on past wave processes. Through investigations of dislocated coarse clasts alone it is not possible to reconstruct the dominant processes and their intensities and frequencies, as the dislocation itself depends on a wide range of environmental factors in their multiple and mutual relationships. Additionally—depending on the scientific question—the spatial and temporal dimensions are important for such analyses.

Cliff retreat through mostly gravitational processes provides an important source for coastal boulders and is mostly a cyclic process on decadal scales. It is driven by the interplay of both long-term impacts of regular short waves and periodic to episodic, much more powerful extreme waves of severe storms or even tsunamis. While the former are more efficient in quarrying boulders, the latter have higher capacities to shift them. Deciphering and quantifying the contribution of extreme-wave events to rocky-shoreline disintegration, however, is hardly possible at present due to the lack of specific case studies.

Rates of (bio-)mechanical and (bio-)chemical downwearing and erosion along highly resistant rocky coasts are highest and best studied in tropical carbonates. Rates vary along vertical and lateral subzones of the coast but reach very high rates of >1 mm/year in the supratidal zone. Having these site-specific differences in mind, the size of post-depositional erosive features associated with boulders such as rock pools, karren, or karst pipes may provide relative age estimates of boulder transport; likewise, post-depositional lichen cover may act as a chronometer. For boulders from subaerial pre-transport settings, different generations of such features need to be considered in this case.

The post-depositional changes of individual boulders related to erosion may—on mid- to late-Holocene time scales—amount to several decimeters for the main boulder axes. Likewise, boulders may have broken up during transport or deposition. This certainly needs to be taken into consideration if boulder size is used to infer hydraulic characteristics of the transport event.

As in other landscapes, the importance of one extreme event depends on the energy of former events and the time span between events: coastal or cliff retreat by storm waves may be extreme, if weathering has weakened rock structure and texture. If this material is dislocated, the next events (even if much stronger) will be less efficient. Good examples for these conditions are rock falls at cliffs, which may produce talus deposits. These often comprise large boulders and debris protecting the cliff toe and part of the cliff face. A sudden cliff retreat may follow decades or even centuries of stable conditions at cliff faces and cliff tops, before the talus is worn away by marine forces (see examples in Erdmann et al., 2017).

Several potential processes may contribute to boulder transport onshore, in particular since we mostly discuss a time frame of thousands of years. None of these processes (frequent waves, rare severe storm waves and bores, infragravity waves, tsunamis) can be excluded at the beginning of any investigation. They may only be eliminated step by step in the course of research at a specific site.

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