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Quarrying, transport and deposition of cliff-top storm deposits during extreme events: Banneg Island, Brittany

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A B S T R A C T

On 10 March 2008, a particularly high energy storm hit the French Atlantic coast and the western part of the English Channel. This storm did not generate exceptionally high waves: significant and maximum wave heights recorded in the Iroise Sea (Brittany) reached 10.85 and 18.17 m respectively, whereas during the 1989–90 winter storms, they had reached 12 and 20 m, respectively. The exceptional character of the March 2008 storm event arises from the fact that it occurred during an exceptional spring tide. From a morphogenetic point of view, the effects of this storm in terms of block quarrying, transport and deposition on Banneg Island (Brittany) were significant. This study shows that the weight of the blocks displaced during the event was between 0.07 and 42.64 t, with a median value of 0.72 t. More than 60% of the blocks were quarried from the wave-scoured cliff-top platform, demonstrating that the favoured zone for supplying material was located inland of the cliff. Two dominant modes of transport were involved depending on the relationship between extreme water levels and cliff height. At the centre of the island, the height of the waves breaking over the top of the cliffs on the western coast resulted in a torrential surge that flowed out towards the eastern coast of the island over a landward-sloping platform. On the western coast of the island, blocks weighing between 0.3 and 1.4 t were displaced between 50 and 90 m from the cliff edge by this flow. At the flow outlet on the eastern coast (90 m from the western cliff edge), a pit 1.6 m deep was excavated at Porz ar Bagou cove and some of the mobilised blocks were deposited in two parallel lobes about 40 m seaward of the eastern shoreline. Elsewhere on the island, block transport occurred by airborne projection although wave heights were lower than the altitude of the cliffs. The pressure exerted by breaking waves on the bedrock was sufficient to quarry and displace blocks. A temporal comparison of the changes recorded in the double and triple boulder ridges showed that the most seaward ridges were practically untouched with the most important changes occurring in the second and third ridges. These observations allow us to propose a spatio-temporal model for the accretion of cliff-top storm deposits (CTSDs), with the various stages of CTSD formation being directly related to the morphological evolution of the cliff.

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1. Introduction

Cliff-top storm deposits (CTSDs) in coastal areas have been reported from various places in the North Atlantic, such as Orkney and Shetland Islands in northern Scotland (Hall et al., 2006, 2008), Aran Islands off the western seaboard of Ireland (Hall et al., 2006; Williams and Hall, 2004), Reykjanes Peninsula in Iceland (Etienne and Paris, 2010), and Banneg Island in the Molène archipelago in Brittany (Fichaut and Suanez, 2008; Suanez et al., 2009). To develop, CTSDs require full exposure to storm waves coupled with deep water offshore so that shoaling of waves is minimal. During extreme storm events, blocks can thus be quarried from the upper part of a cliff, transported by wave run-up or overtopping across cliff-top platforms or ramps and deposited further inland. CTSDs are typically characterised by large, angular blocks that reach 277 m^3 and a lack of sorting (Hansom et al., 2008). The lack of sorting demonstrates that CTSDs are repeatedly re-worked by storm waves and tends to distinguish them from tsunami deposits that are generally dumped during a single event and have finer clast [grain] sizes on their landward sides (Williams and Hall, 2004). Nevertheless as demonstrated by Paris et al. (2009), after the 2004 tsunami on Sumatra Island, lack of sorting may also occur in tsunami deposits. CTSDs typically accumulate at some distance back from the cliff edge behind cliff-top rock surfaces (wave-scoured platforms or ramps) swept clear of debris by storm activity. CTSDs are found as different morphological features such as steep-faced ridges composed of seaward-dipping imbricated clasts, or more simply, clusters, spreads and individual clasts. As observed by

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Etienne and Paris (2010) and Williams and Hall (2004), organisation into ridges may require repeated re-working by storm waves rather than the much rarer impact of a train of tsunami waves. This layout does not occur in case of tsunamis (Paris et al., 2009).

Several studies have attempted to analyse and model the processes responsible for quarrying, transport and deposition of large clasts. Using a theoretical approach, Noormets et al. (2004) assumed that the rock body must have substantial initial fracturing for swell waves to quarry large clasts from the cliff edge. Their modelling study demonstrated that dislodgement, entrainment, transport and deposition most likely occurred in sequence during the impact of a single wave on the cliff edge. Nevertheless, maximum dynamic shock pressures applied to the cliff edge are the result of steep deep-water waves ($H_0/L_0 = 0.06$–0.04), with a relatively short period ($T = 8$ s). On wave-scoured platforms, Noormets et al. (2004) showed that sliding without rotation is the main mechanism of transport of the large, irregular-shaped megaclasts, whereas tabular clasts are rolled and occasionally turned over during transport. More recently, Hansom et al. (2008) modelled the morphogenetic context of CTSDs using wave-tank laboratory experiments. The results from these experiments suggest that steep waves of 10 m and greater that hit a 15 m cliff will result in impact pressures sufficient to promote crack propagation, block detachment and lifting of large blocks. Large, but not necessarily steep, waves of the same height as the cliff edge produce sufficient impact pressures and water flow over the cliff edge to the cliff-top platform to detach, transport and deposit blocks on the cliff top. When incident, non-breaking wave crests overtop a cliff edge, bore flow occurs, with sufficient force to rotate or lift blocks from the cliff-top and adjacent wave-scoured platform. High flow velocities rapidly accelerate and transport blocks inland until flow attenuates, resulting in deposition of blocks at the limit of run-up. Field observations on the Grind of the Navir (in the Shetland Islands) were used to help clarify erosion and deposition patterns produced by high-energy wave processes (Hall et al., 2008). These authors showed that in most years, the cliff-top is washed with flows due to incident storm wave crests that are higher than the cliff-top. This upward moving water flow is capable of removing blocks from both vertical faces and steep overhangs. It is also capable of quarrying blocks from the cliff face close to the edge and from rock steps on the cliff top, promoting further rock fracturing. Socket sizes and block characteristics indicate that blocks of >1 m^3 are rotated from sockets on the cliff-top platform, transported landwards for up to 60 m and deposited in a series of boulder ridges inland of the cliff top. Nevertheless, fresh sockets low on the cliff face also indicate removal of fracture-bounded blocks that are lost seaward. Nott (2003) developed hydrodynamic equations that relate the forces required to transport boulders from their pre-transport position to where they are currently located. These equations assumed that boulders resting in a sub-aerial position on a wave-scoured platform are subject to the force of inertia as well as that of drag and lift, whereas submerged boulders only experience the latter two forces. In contrast, boulders derived from joint-bounded blocks on a wave-scoured platform are influenced by lift force alone, requiring much higher waves to initiate transport.

With regard to these analyses and models, this paper seeks to analyse the processes of quarrying, transport and deposition of CTSDs that occurred during the 10 March 2008 extreme storm event on Banneg Island (Molène archipelago, Brittany). Based on a study of the morphological changes that occurred, the first objective of this research was to analyse the dynamics of quarrying, transport and the deposition that result in the particular block accumulations found on the island. The second objective was to improve our understanding of the processes behind the construction of boulder ridges and clusters. Finally, a comparison with the 1989–90 winter storms was carried out to gain insights into how block accumulations may evolve in the long term.

2. Regional setting

2.1. Morphological setting of CTSDs on Banneg Island

The distribution and morphology of CTSDs on Banneg Island have previously been described in detail (Fichaut and Suanez, 2008; Suanez et al., 2009). Here, we give only brief outline of the salient features of the CTSDs on Banneg Island. Banneg Island lies in the north of the Molène archipelago off the north-western tip of Brittany (Fig. 1). The island is a granite batholith, oriented north–south, 0.8 km long and 0.15 to 0.35 km wide; the western coast is flanked by 50 m deep water at 2 km from the shore and thus exposed to storm waves from the Atlantic Ocean. The coast is cut into high, sub-vertical cliffs at headlands with lower, less steep cliffs in embayments. More than 1000 m^3 of blocks, whose individual weights vary from several tens of kilogrammes to several tens of tons, have been deposited at the rear of the cliff top edge backing the embayments on a slope that gently dips towards the eastern coast of the island. The furthest inland accumulation is 100 m from the western cliff edge. CTSDs can locally form a spread of isolated blocks that can be found all the way down to the eastern coast such as in the centre and the centre-north areas of the island (Fig. 2). However, CTSDs more often form clusters or ridges at altitudes between 7 and 14 m above sea level (asl) or French datum (3.5 to 11.5 m above the highest spring tide) and up to 75 m from the edge of the cliff (cluster #74 in the centre-north area, see Fig. 2). Over the entire island, 61 accumulations have been inventoried; the largest ridge is located at the centre of the island (ridge #28). It reaches a height of 2.5 m, stretches over 60 m long and 20 m large, and it has a volume of around 350 m^3. In general, the accumulations are in rectilinear ridges parallel to the cliff edge (e.g. accumulations in the centre and north areas), or in arcuate trains at the rear of embayments, mirroring the area's coastline (e.g. accumulation in the centre-south and south areas). Locally these ridges are deposited in several parallel lines, separated by areas of turf or rock surfaces. There are double ridges in the south, with clusters #36, #37 and #38 lying seaward to clusters #35, #39 and #40. There are triple ridges in the central part of the island, such as the one where clusters #28, #25 and #22 compose the most seaward ridge, ridge #24 makes up the middle ridge and clusters #23 and #19 constitute the most inland ridge. Another triple ridge can be found in the centre-south area and is made up of ridges #31 and #30 with cluster #23d at the rear. In triple ridges, the accumulations closest to the cliff edge are the largest ones. Comparable patterns have been described in the Shetland Islands (Hall et al., 2006). In only one case, in the centre-north area, clusters are scattered at the rear of the cliff, with a seemingly random distribution.

2.2. Lithostructural setting: predisposition to block quarrying

As demonstrated by Naylor and Stephenson (2010) geological properties and differences in joint spacing play a fundamental role in rock resistance and erosion. Banneg Island is composed entirely of Paleozoic leucogranite. The morphological characteristics and evolution of the cliffed west coast depend in part on the structure and joint system of the bedrock. Roughly rectilinear, the west coast is actually made up of a series of more or less pronounced headlands and embayments (Fig. 2). At the headlands, massive jointing organized in a loose orthogonal pattern makes the rock body more resistant to marine erosion. The headlands thus constitute the summits of the island usually in the form of towering rocky outcrops when vertical joints are more numerous than horizontal ones. The cliff tops of headlands have altitudes in excess of 12 m and frequently reach 14 m and the slope of the cliff face is always greater than 50%. At the rear of the headland cliffs, no clusters or ridges occur. All headlands exhibit this type of morphology, except the headland situated between the south and centre-south areas, called Skloum Maurice. At Skloum

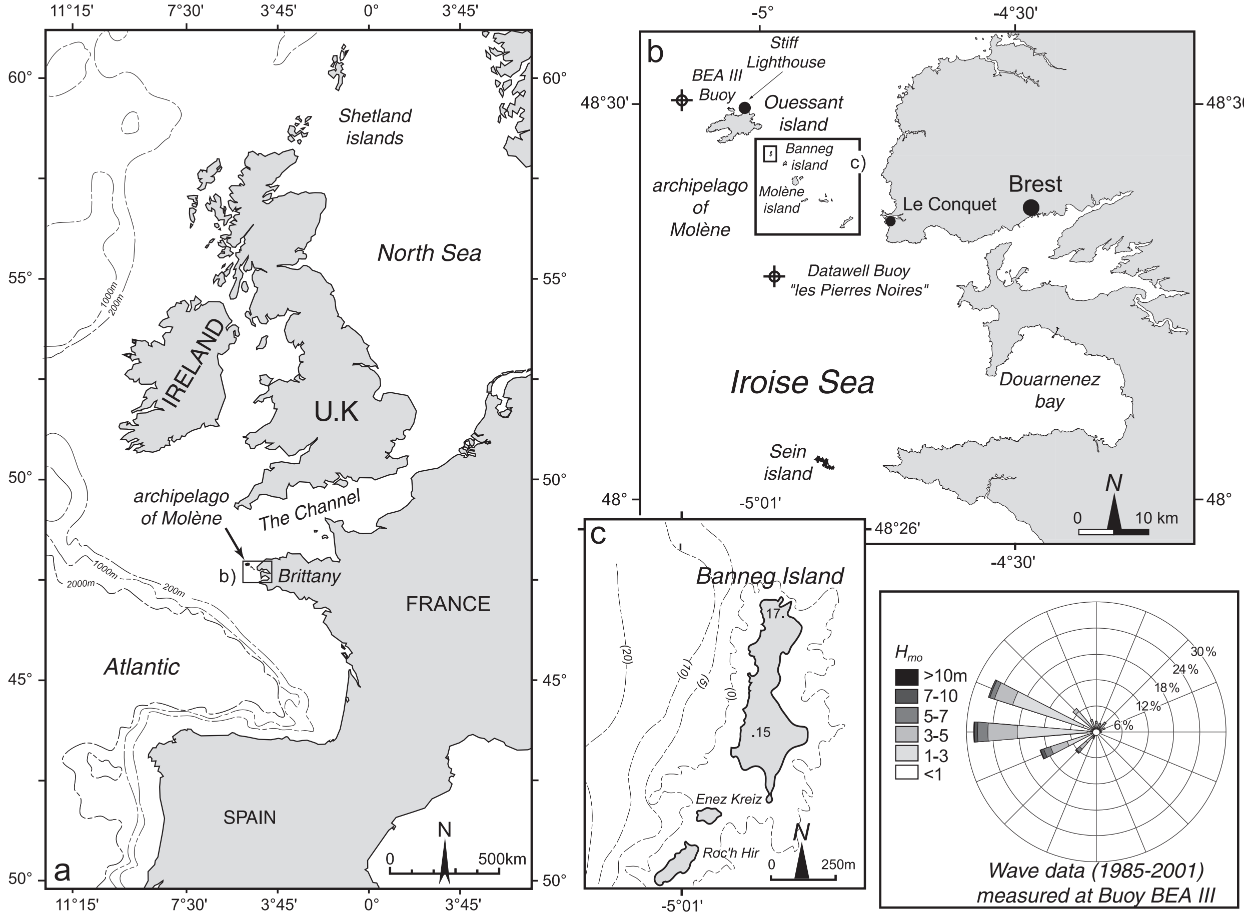


Fig. 1. Location map.

Maurice, the upper part of the cliff shows many horizontal joints called sheet joints that cut the rock face into large horizontal slabs (Fig. 3b). The rounded dome shape of this headland, which contrasts with the above-described towering outcrops, is due to these sheet joints. Likewise, due to the dense jointing in the upper part of the cliff face, some isolated clasts can be found at the rear of Skloum Maurice.

At the base of cliffs along the embayments, massive, orthogonal jointing engenders enormous quadrangular clasts. When these clasts are dislodged, they accumulate on the shore as an armor layer of rounded boulders or quadrilateral-shaped boulders with smoothed edges, weighing from one to several tens of tons (Fig. 3b). In contrast, from the middle to the upper cliff face, the granite has very dense sheet jointing, promoting large horizontal or box-shaped clasts of smaller size. Similar patterns have been identified on other coastlines (Knight and Burningham, 2011; Stephenson and Naylor, 2011). These lithostructural characteristics differentiate the embayments from the headlands and make the embayments much more sensitive to marine erosion. In embayments, cliffs, which generally have gentler slopes (15 to 35%), culminate at an average altitude of 8 and 9 m asl. Most boulder clusters and ridges are found at the rear of embayments (Fig. 2).

2.3. Hydrodynamic setting

In the Molène archipelago, the directional wave spectrum shows that waves are predominantly westerly to north-westerly (270° to 310°); these waves are the most energetic and represent roughly 60% of the annual regime. Based on buoy BEA III measurements off Ouessant Island (Fig. 1) between 1985 and 2001, the significant wave height is roughly 1.5 m (30% of the annual regime). The most frequent wave heights (between 25% and 30% of the annual regime) reach 2.5 m, but may exceed about 16 to 18 m (Suanez et al., 2009). Banneg Island, located on the western edge of the Molène archipelago is directly exposed to Atlantic swell. The 20 m and 50 m isobaths are respectively 250 m and 2200 m from the cliffs on the western side of the island and so these cliffs receive the full force of the dominant waves that are only slightly attenuated.

3. Characteristics of the 10 March 2008 storm

On 10 March 2008 extra-tropical cyclone Johanna was caused by a low pressure system that had originated two days earlier over Newfoundland, crossing the Atlantic in a longitudinal track (Cariolet et al., 2010). On the morning of 10 March, the storm hit the European coasts generating a pressure gradient of 30 hPa over 600 km, stretching from Ireland to southern Brittany. The arrival of the cold front at 0:00 on the morning of 10 March generated a barometric pressure drop inducing and strong south-westerly winds (Fig. 4a). Once the cold front passed, atmospheric pressure gradually increased and the wind shifted west and then north-west, with increasing velocity. On the evening of 10 March, the cyclone eye moved to southeastern England, 500 km from northern Brittany, but the pressure gradient remained high (25 hPa), inducing gale force west-north-westerly winds. During the 11th of March, the depression gradually dissipated over the North Sea.

During the morning of 10th of March, a 0.80 m storm surge was recorded at Le Conquet; in the evening, values were lower (<0.43 m) because wind speeds decreased and atmospheric pressure increased (Fig. 4b). From a statistical point of view, comparing the extreme water levels recorded at Le Conquet on 10 March with the tidal

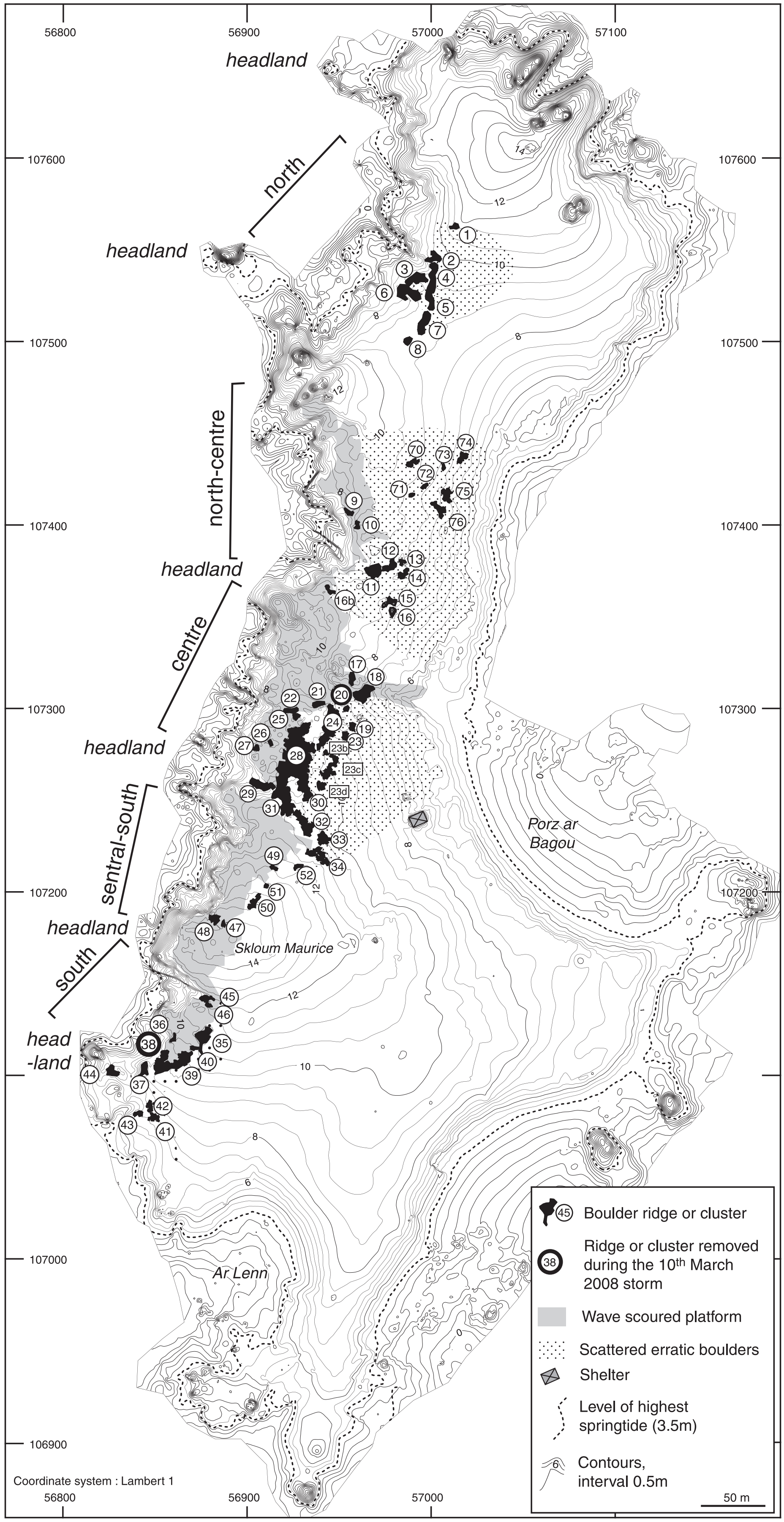


Fig. 2. Location of CTSDs on Banneg Island (after Suanez et al., 2009).

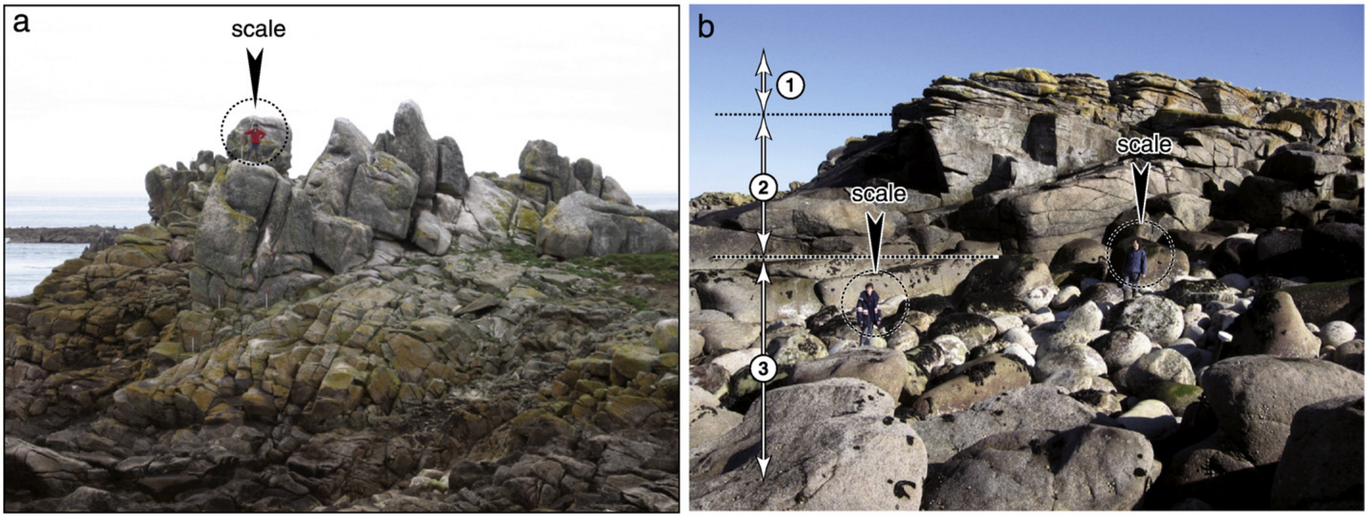


Fig. 3. Lithostructural components of the rocky cliffs on the west coast of Banneg Island. a) Granitic headland between north and centre-north area. b) Granitic headland between south and centre-south sectors on Banneg Island. 1: High density of horizontal joints bounding rock clasts. 2: Massive joint system bounding huge cuboidal blocks. 3: Lichen and algae free edge-rounded clasts abrased during the 10 March 2008 storm.

predictions of the French Naval Hydrographic and Oceanographic Service (Simon, 1996), shows that these water levels have a return period of 10 years. The only wave measurements available for Brittany coasts for this storm event are those that were recorded by the directional waverider located near Les Pierres Noires south of the Molène archipelago (Fig. 1). The significant and maximum wave heights reached 10.85 m and 18.17 m respectively, in the mid-afternoon (Fig. 4a). The steepness of deep-water waves was $0.01 < H_0/L_0 < 0.06$. From a statistical point of view, these wave heights were not remarkable (a return period of two years). For example, off Ouessant Island during the 1989–90 winter storms, the significant and maximum wave heights were more than 12 m and 20 m, respectively (Fichaut and Suanez, 2008).

The severe morphogenetic impact of the 10 March storm stems from the fact its landfall occurred (on neighbouring islands, Fichaut pers. obs.) during a high spring tide at 3.9 m asl. The co-occurrence of these two parameters during high-magnitude events is exceptional and has not been observed since the middle of the 19th century (Pirazzoli, 2000). The most dramatic erosional effects of the storm occurred mainly during the evening high tide on 10th of March as will be demonstrated with wave data later in the paper.

4. Methods

4.1. Analysis of the hydrodynamic conditions during the storm

The analysis of hydrodynamic conditions focussed on quantifying the extreme water levels on the coast in order to determine the conditions under which the height of run-up from breaking waves reached or overtopped the upper part of the cliffs. It has been repeatedly shown that it is under these conditions that CTSD quarrying and transportation occur (Hall et al., 2008; Hansom and Hall, 2009; Hansom et al., 2008; Noormets et al., 2004).

The estimation of these extreme water levels on the coast was based on the combination of two parameters acting on the deformation of the water surface: observed tide levels and run-up, calculated using the equation established by Mase (1989). As mentioned in a previous study (Suanez et al., 2009) the Mase's formula is used in case of irregular waves (wave steepness, H_0/L_0 of $0.007 \leq H_0/L_0 \leq 0.07$) breaking on slopes of $0.03 \leq \tan\beta \leq 0.2$. The offshore slope on the west coast of Banneg is $0.02 \leq \tan\beta \leq 0.05$, and during the 10th of March storm the wave steepness was H_0/L_0 of $0.04 \leq H_0/L_0 \leq 0.09$. The values of steepness and slopes do not fit exactly with the intervals of Mase's formula but, the results show that the theoretical estimation of run-up heights closely fit the observations made in the field. The wave data used to calculate run-up heights come from the records of the waverider at Les Pierres Noires; tidal data were extracted from the tide gauge records at Le Conquet.

4.2. Analysis of the quarrying, transport and deposition of CTSDs

The analysis of quarrying processes was based on an inventory of the sources of blocks that were transported during the storm. Sockets in the bedrock were thus recorded, an easy task because these areas lack lichen cover and show sharp edges at their fracture points (Fig. 5a). Regarding the displacement of isolated blocks, already deposited on the cliff top or in pre-existing clusters, the absence of obvious source scars made identification more difficult. In such instances identification of sources was based on the absence of lichen cover on the bedrock and the absence of vascular plants in soils lodged within joints and cracks, both of which indicate that a clast was present just prior to the storm. Direction and carry distance were thus determined by comparing the shapes of the deposited boulders and the shapes of sockets or scars. The latter and the current resting places of corresponding transported boulders were recorded by Differential Global Positioning System (DGPS) measurements (Fig. 5b). These data, issued from this semi-quantitative method, were used to determine and analyse how the blocks had been transported. The dimensions and the weight of these blocks were measured to analyse their sedimentological characteristics.

4.3. Analysis of the morphological changes in ridges and clusters

This study also included a time-series analysis of recent ground photographs. Every accumulation had been photographed on purpose in 2005 and 2006, in anticipation of the potential occurrence of a storm that would re-work the CTSDs. After the storm, the accumulations were photographed again from the same angle. This analysis made it possible to (i) determine the direction and displacement distance for some blocks; (ii) estimate the degree to which each cluster in each of the different areas of the island had been re-worked or (iii) indeed had completely disappeared (Fig. 5c and d). Finally, a similar comparison with older, oblique aerial photographs taken in

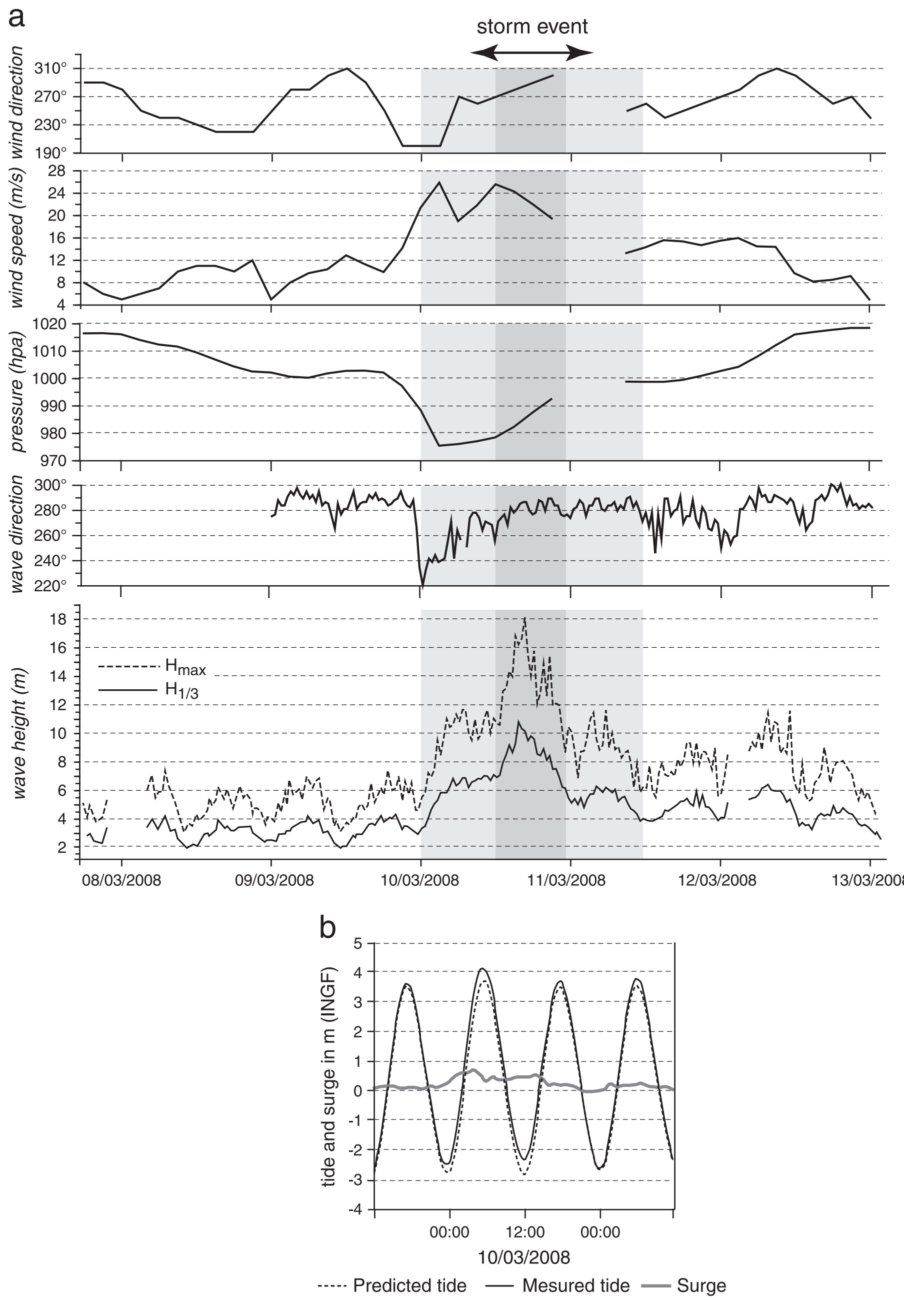


Fig. 4. Hydrodynamic and marine weather conditions during the 10 March 2008 storm event (modified from Cariolet et al., 2010). a) Meteorological and oceanographic data (wind, pressure and wave) recorded between 8 and 13 March 2008 in the Molène archipelago. Wind and pressure data obtained from meteorological station Météo France at Ouessant Island (Stiff Lighthouse, see Fig. 1); wave data obtained from the French Naval Hydrolographic and Oceanographic Service (SHOM) Datawell buoy at Les Pierres Noires (see Fig. 1). b) Tide data obtained from the SHOM tide gauge station at Le Conquet (see Fig. 1).

1979 and 1990 helped reconstitute the sequences of changes that have occurred in the clusters over the past three decades.

5. Results

5.1. Submersion of the central area of Banneg Island

According to the tide gauge at Le Conquet, the maximum tide level was reached at 19:00; maximum wave heights reached nearly 15 m and storm surge was around 0.80 m. The theoretical estimation of run-up height indicates that when swells were at their maximum heights, the waves breaking on the west coast of the island were as high as 12.5 m. Combining these wave heights with the observed tide levels, the extreme water levels may have been greater than 15.5 m at the centre-south area of the island, and may have overtopped the cliff edge by 2 to 7 m (Fig. 6b). In the centre-north and north sectors of Banneg Island, where the slopes of the foreshore are not as steep, run-up may have only reached 9.5 m, a height equivalent to or slightly greater than that of the cliffs (Fig. 6a). As detailed below, our theoretical estimations of run-up heights closely fit the observations made in the field, particularly those that identify the trash line marking the limit of wave wash (Fig 7a).

At the northern part of the island, run-up heights reached the top of the cliff where some blocks were displaced; however, the vegetation on the cliff-top platform at the rear did not show any signs of wave wash. Here, no breaking waves overtopped the edge of the cliffs, confirming the low run-up height values calculated for this area. In the southern part of the island, uprooted turf and debris

Fig. 5. a) Fresh socket on a cliff edge. b) Quarrying and transport of clasts on cliff tops at Banneg Island during the 10 March 2008 storm. 1a and 2a: quarrying socket with sharp edges and absence of lichen cover. 1b and 2b: deposition of quarried blocks and transport direction. c) Cluster #20 in 2005 before the 10 March 2008 storm event. d) Cluster #20 eroded during the 10 March 2008 storm event. 1: Location of cluster #20 in April 2008.

deposited behind clusters indicated roughly the inland limit of wave wash. Waves overtopped the cliff up to 12 m asl. and flowed out through Ar Lenn Cove following the slope of the island (Fig. 7b). The central part of the island, with altitudes of 9 to 14 m, was completely submerged. The southern limit of wave wash, marked by a continuous trash line (seaweed and drift debris), was recorded using DGPS. This limit of wave wash runs behind the cliff at 14 m altitude – matching the calculated run-up height – and continues down to the eastern coast. The northern limit of wave wash was not as marked, but could nevertheless be estimated (Fig. 7a). In this area, gravel eroded from the decomposed granite at the cliff top, north of cluster #9, was deposited in a thin and discontinuous spread on top of the turf located at the back of the cliff. The level of the water that inundated this part of the island can be estimated from field observations. As during the 1989–90 winter storm (Fichaut and Hallégouët, 1989), the overtopping flow was locally torrential, especially in the centre-south area and near the gully at the centre of the island. At the centre-south area, the cliff-top platform behind clusters #48 and #50 culminates between 12 and 14 m at the cliff tops and slopes to the east coast at an average slope of 6%. On this slope, patches of turf and soil of up to 4 m^2 were torn up at several places (Fig. 8a). The interior of the building located in this area was flooded, although its door was on the opposite side of the flow (Fig. 7). Just north of this area, behind the two to three parallel ridges of the island (#28, #30 and #31), the turf remained intact. It is likely that the water flow had been channelled between the ridges and flowed out simultaneously to the south and towards the central gully (Fig. 7). Submersion levels reached their maximum in the gully and the surrounding area. The low altitude of this part of the island (around 9.5 m) and the eastward slope facilitated the passage of particularly high velocity flows that considerably changed the morphology of the gully and its surrounding area (Fig. 8). As shown in Fig. 9, cluster #18, lying in the axis of the gully and comprising a volume of 12 m^3, was almost completely swept away by the water flow. Similarly, cluster #20 positioned 2 m above the thalweg of the gully completely disappeared as well as half of cluster #17 (Figs. 8b, c and 9). In the eastern part of the centre area of the island located between the gully and the shelter, five blocks weighing from 0.3 to 1.4 t were found upside down on the turf at distances of 50 to 90 m from the edge of the western cliff (Fig. 10c). These blocks were not overturned in place because there were no scars in the surrounding turf where they would have previously sat. They could not have come from Porz ar Bagou cove either, because it is composed of rounded pebbles and quarrying and transport of blocks from this cove would have had to occur in the opposite direction of the storm waves. The blocks thus come from the western cliff-top platform of the island and were carried over unknown distances eastward. At the head of the gully located on the eastern coast, torrential flow dug a 1.6 m deep pit (Fig. 8d and e). Some of the blocks that were displaced eastward were deposited in two parallel lobes on the beach of Porz ar Bagou cove, 40 m from the shoreline (Fig. 9b).

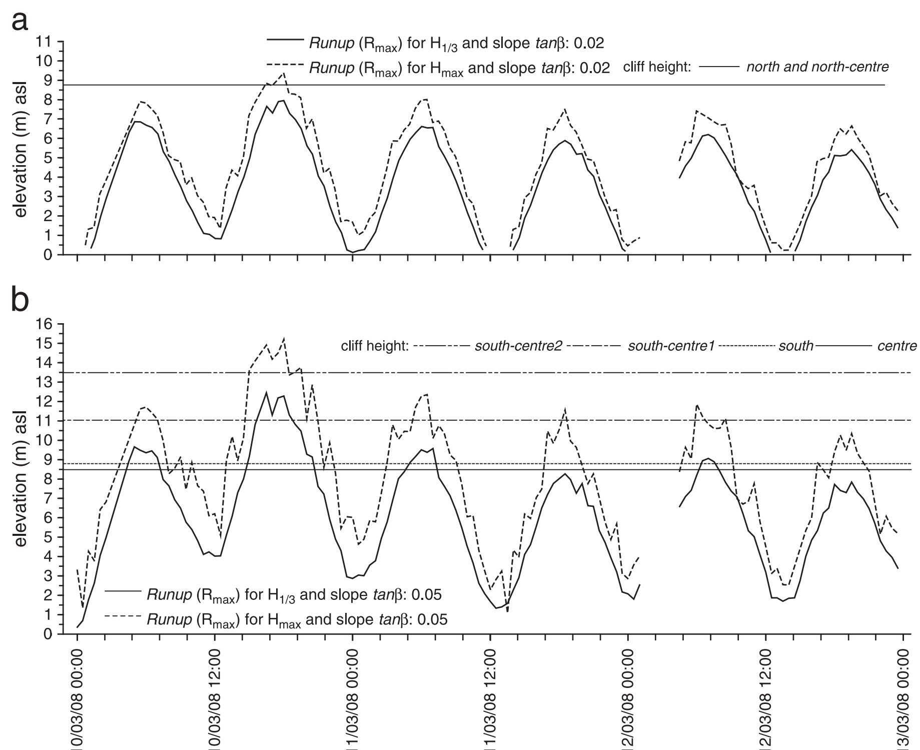


Fig. 6. Heights of extreme water levels (measured tide and runup) during the 10 March 2008 storm event compared to cliff heights.

Given these data, it is likely that when submersion was at its maximum, water levels were higher than 3 m in the gully and at least 1 m on the slope behind the clusters of the centre-south area. North of the gully, water levels were probably lower and the flow was weaker because the turf was not scoured and, except for the immediate edges of the cliff tops, no blocks were displaced.

5.2. Quarrying processes and sources of blocks

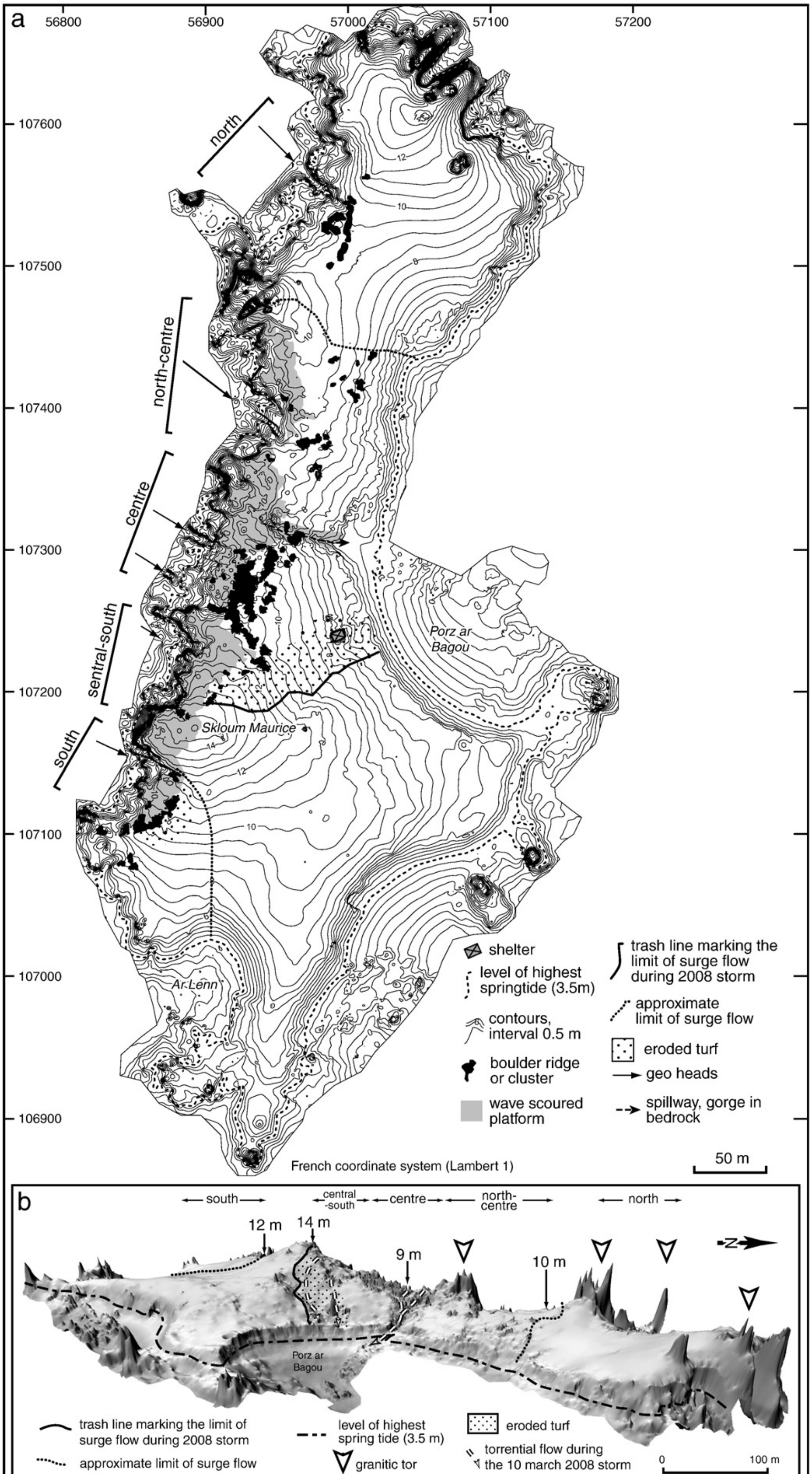
Based on the field work carried out between April 22 and April 25 2008, after the 10 March 2008 storm, we identified and located, using DGPS, 180 sockets in the bedrock and/or scars indicating that isolated blocks had moved (Fig. 11). This figure probably does not reflect the actual number of blocks that were quarried or displaced. While the inventory of sockets was practically exhaustive, that of scars left by displaced blocks was incomplete. Sometimes, entire clusters were erased by erosion (Fig. 10c and e).

However, in spite of these gaps in our inventory, the sources of transported blocks could be determined with a high degree of precision. The lowest socket was located at 0.4 m asl and the highest at 11 m asl (Fig. 11). In the centre-south sector, all sockets and remobilisation scars were found at levels greater than 6.5 m asl (Fig. 10d), whereas in the centre and in the centre-north, they were all below 6.5 m asl (Fig. 10a, b and c). This distinction reflects the difference in cliff heights in these zones and indicates only that storm waves can quarry blocks at 11 m above the highest spring tides. The analysis of the relative position of sources is more informative: it shows that 60% of the blocks were quarried not from the cliff itself but from the cliff-top platform at the rear (Fig. 11 and Table 1).

Closer inspection of the data shows that the location of block sources varies. In the north, erosion primarily affected the upper part of the cliff. Furthermore, given that the run-up barely overtopped the cliff, these morphological observations corroborate previous results obtained based on hydrodynamic conditions (which was also the case during the 1989–90 winter storms, see Fichaut and Hallégouët, 1989). This is the only place on the island where the rocky cliff-top platform is not scoured and the cliff top extends directly into a turfed platform (Figs. 10a and 11). In the south, the high proportion of blocks quarried from the cliff results from the fact that the southern coast trends parallel to the direction of incident waves; waves thus do not break directly in front of the cliff face, but alongside it (Fig. 10e). In the centre area, the lowest part of the island, erosion was greatest landward of the cliffs, on the wave scoured platform. This morphological feature means that action of breaking waves extends far beyond the edge of the cliff; it also facilitates the passage of overtopping waves and a flow regime capable of moving boulder ridges or isolated clasts, particularly near the gully. This water flow displaced blocks of up to 1 t and transported them all the way to the sheltered east coast. Clearly, hydrodynamic conditions also play a significant role during these storm events, particularly in the spectacular effects of run-up. In the centre-south where the cliffs culminate from 10 to 14 m asl and have a seaward gradient of up to 60%, tabular clasts of more than 2 t were quarried (Fig. 10d). In this specific area, block quarrying is most likely to be produced by waves breaking at the upper part of the cliff face and rebounding several meters beyond the cliff edge, where quarrying blocks in place or displacing of loose blocks already present occurs.

5.3. Block transport and deposition processes

The tracks of 63 blocks displaced on 10th of March could be reconstructed based on the identification of their sources and resting places (Fig. 10). However, this sample does not represent all the block displacements, the ground photos taken before and after the storm indicate that several hundreds of blocks had been moved. Analysis of the identified tracks shows that the predominant direction trended WNW–ESE, the main orientation of the incident waves (Fig. 1). The weight of the 63 blocks varied from 0.07 to 42 t (Table 2). There is no significant correlation between weight, carry distance, vertical lift or the socket position and cliff height. However, morphometric data gave an indication of the intensity of the morphodynamic processes that occurred during the storm event. The largest block (42 t) was quarried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Ar Lenn
shelter
level of highest springtide (3.5m)
contours, interval 0.5 m
boulder ridge or cluster
wave scoured platform
trash line marking the limit of surge flow during 2008 storm
approximate limit of surge flow
eroded turf
geo heads
spillway, gorge in bedrock
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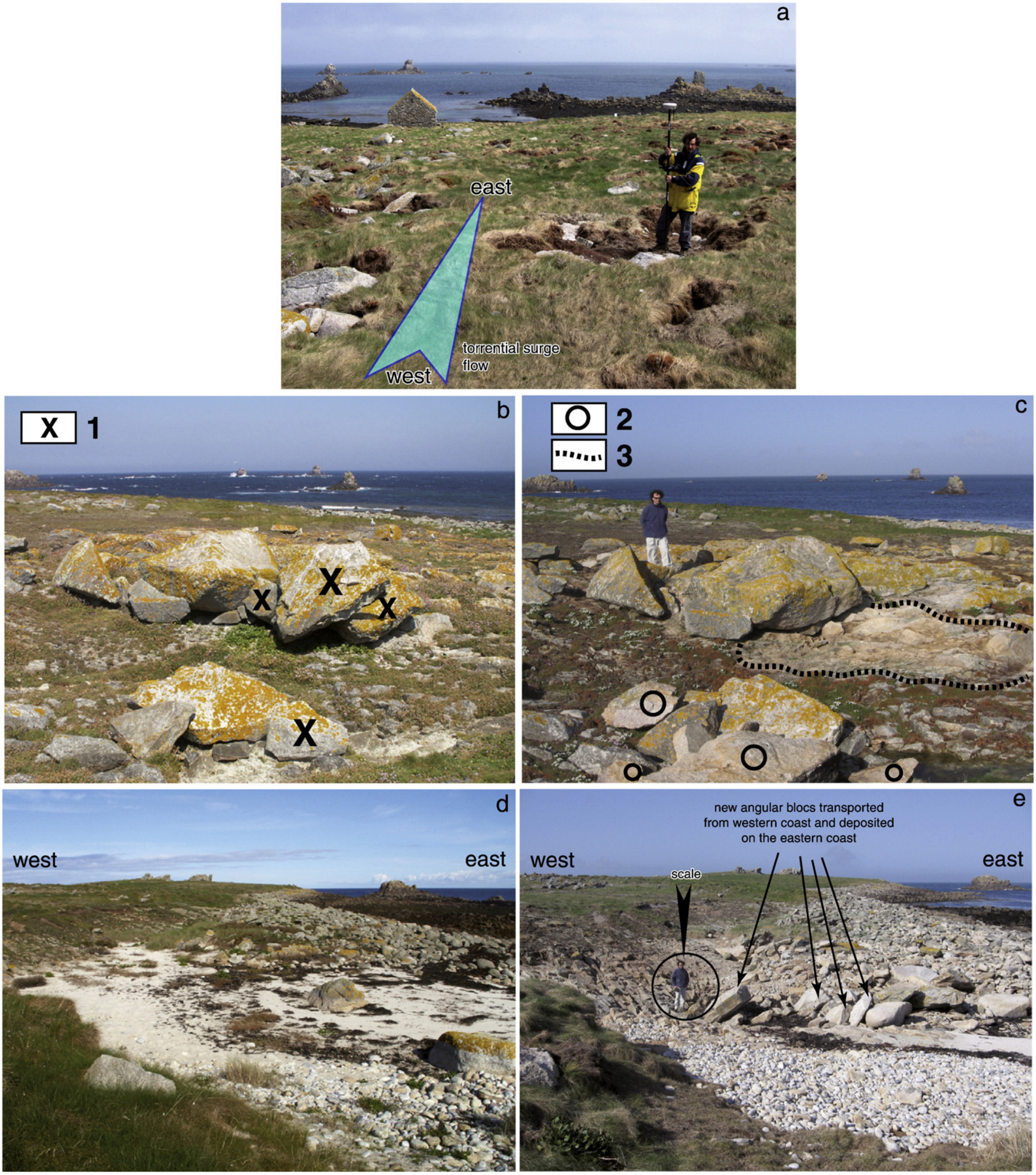

Fig. 8. Morphological changes in the central part of Banneg Island. a) Patches of turf removed by torrential surge flow during the 10 March 2008 storm; b) and c) comparison between (b) June 2005 and (c) April 2008 showing cluster #17 partially removed during the 10 March 2008 storm event. 1: Clast removed. 2: Clast deposited. 3: Socket. d) and e) Comparison between (d) June 2005 and (e) April 2008 of the area located at the mouth of the gully on the eastern coast of Banneg Island. Note the 1.6 m deep pit dug in the upper sandy beach and deposition of blocks during the storm.

Fig. 7. Location of submerged areas on Banneg Island during the 10 March 2008 storm event. a) Planar view showing extension of submerged area. b) Digital Elevation Model showing the lowest part of Banneg that favoured submersion.

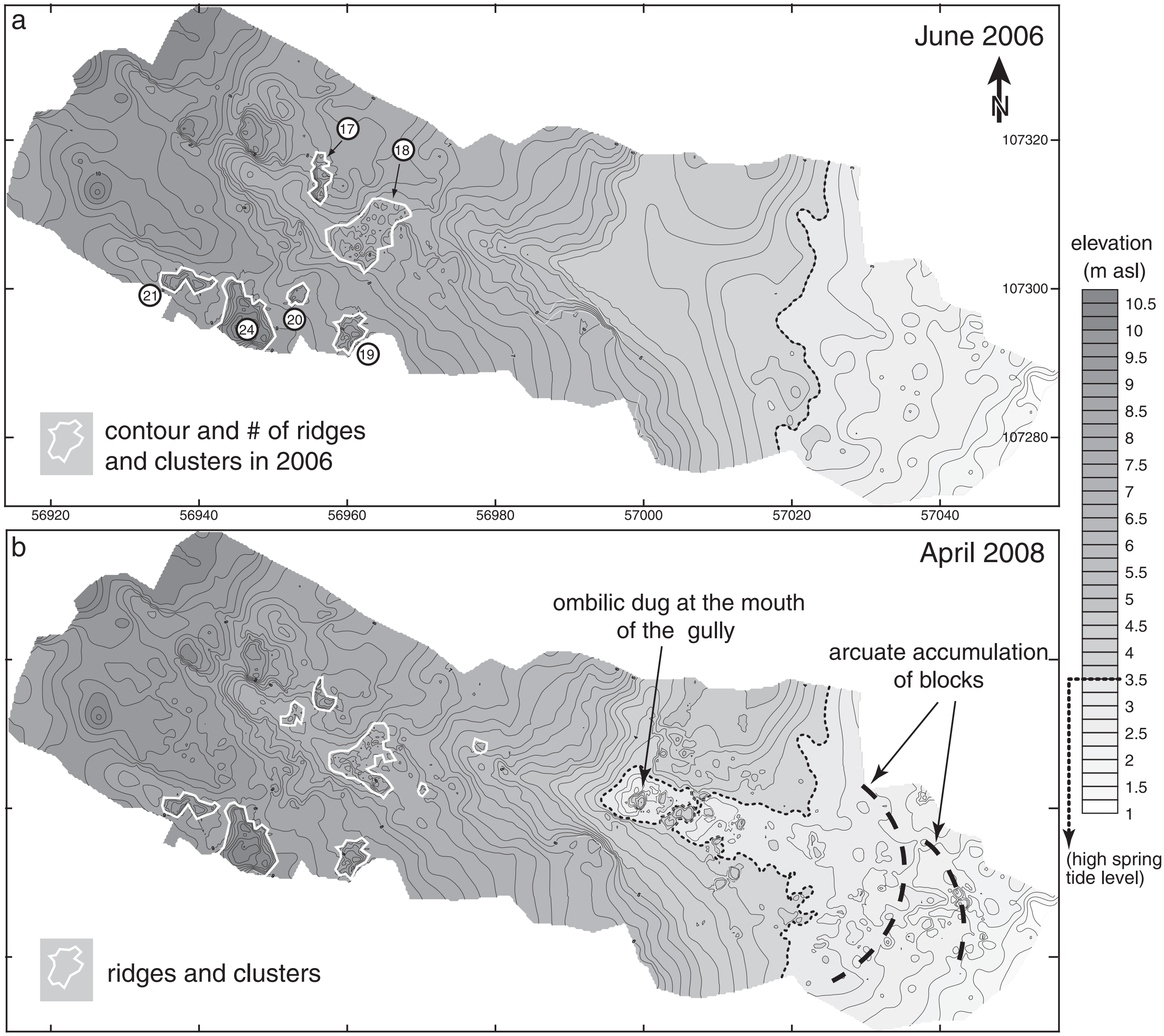

Fig. 9. Topo-morphological changes of the gully and its surrounding area at the centre of Banneg Island between (a) June 2006 and (b) April 2008 due to the 10 March 2008 storm event.

from the bedrock at 3 m above the highest spring tides and was deposited upside down 7 m away (Fig. 12a). In its passage, it swept away part of cluster #44 (Fig. 10e). In the centre area, a 7 t clast, deposited 40 m landward of the cliff edge on cluster #24, was upturned (Fig. 12b). In the same area, another block weighing 2 t was deposited at 45 m from its starting point (Fig. 10c). In the centre-south area, a clast of 2.4 t resting at 14.5 m in altitude, was displaced (Fig. 10d).

5.4. Mechanisms of block transport within clusters

5.4.1. During the 10 March 2008 storm

The following analysis is based on the comparison of ground photographs taken in 2005–06 and after the 10 March 2008 storm on 4th of April. As the wave data indicates that no morphogenic storm occurred in the meantime (Suanez et al., 2009), it can be assumed that the morphological changes took place on the 10 of March 2008. The analysis concerns a double ridge located in the southern part of the island, composed of cluster #36 on top of the cliff edge and ridge #39 located 13 m further inland. The analysis was also based on a triple ridge located in the centre area and composed of clusters #28, #24 and #23 (Fig. 10c and e).

In the southern part of the island, ridge #39 was drastically reworked during the storm (Fig. 12c and d). Five blocks in the northern part of this cluster were moved, two of which weighed 0.26 and 0.72 t and were found 6 to 8 m further inland (Fig. 12e and f). They were displaced in the same direction as the incident waves (Fig. 10e). In contrast, cluster #36 located at the forefront and the adjacent blocks lying in the axis of the oncoming wave did not move (Fig. 12e and f). It is thus clear that a wave or a mass of water from a breaking wave arched over the cliff edge at 5.5 m asl and landed at distance behind the cliff edge on cluster #39.

The triple ridge at the centre of the island consists of ridge #28, the first, most seaward ridge with a height of 2.5 m and lying 15 m beyond the edge of the cliff, followed by clusters #24 and #23 further inland. On 10 March, the exposed face of cluster #28 changed only slightly. A few blocks were removed, and others were deposited at the foot of the ridge. The ridge had the same topography as before the

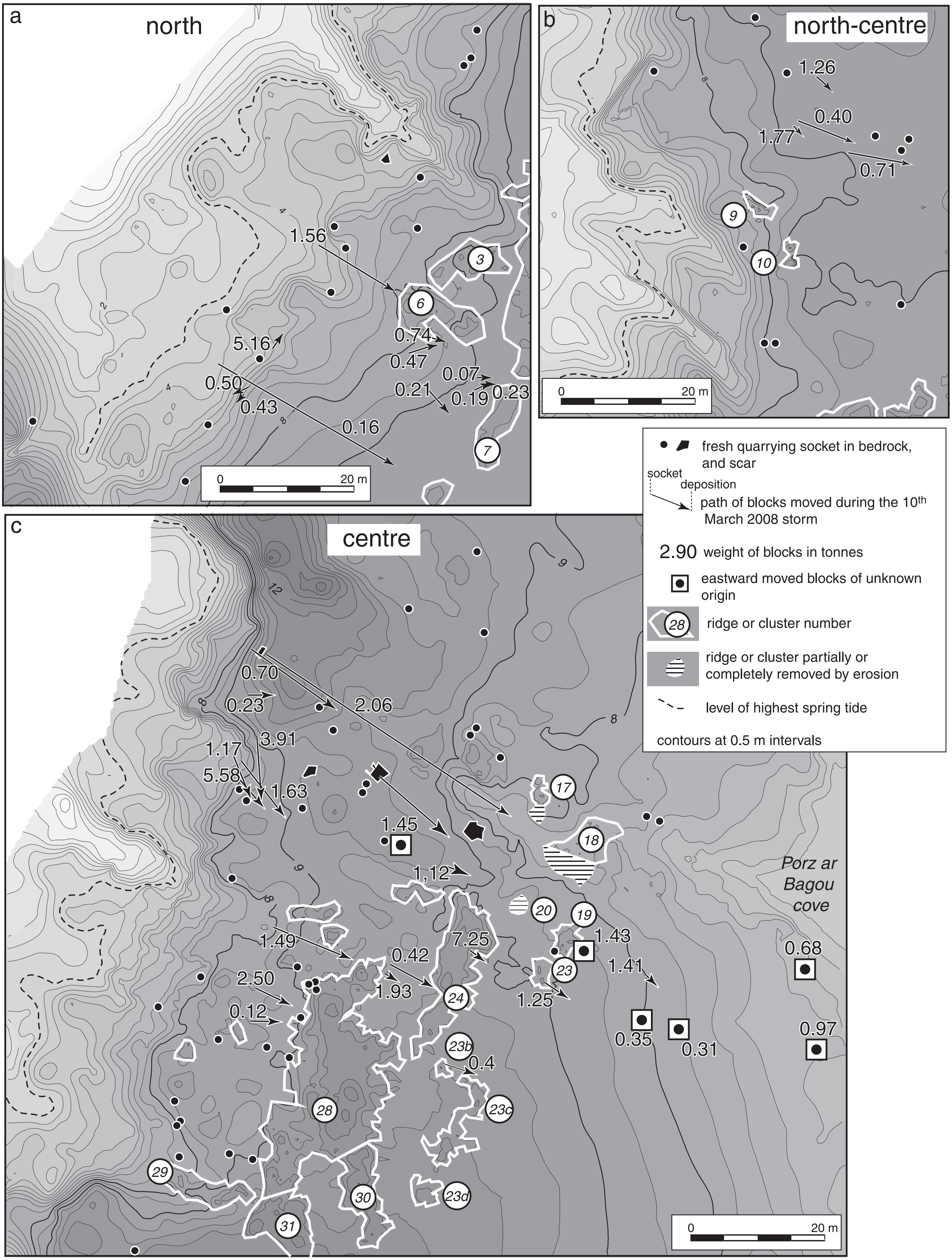

Fig. 10. Erosion, transport and deposition of CTSDs at the cliff-top and on the wave scoured platform on Banneg Island during the 10 March 2008 storm event.

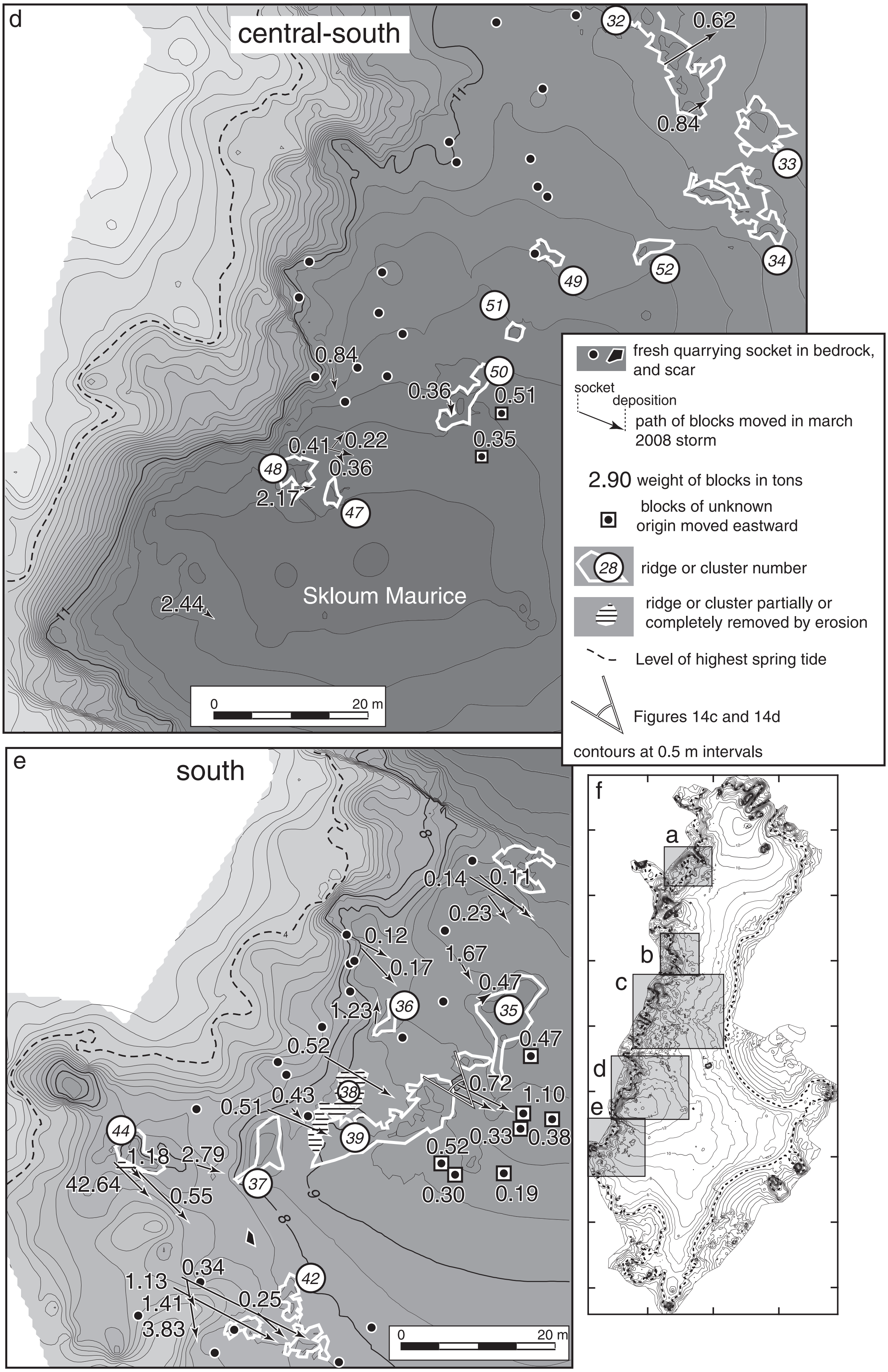

Fig. 10 (continued).

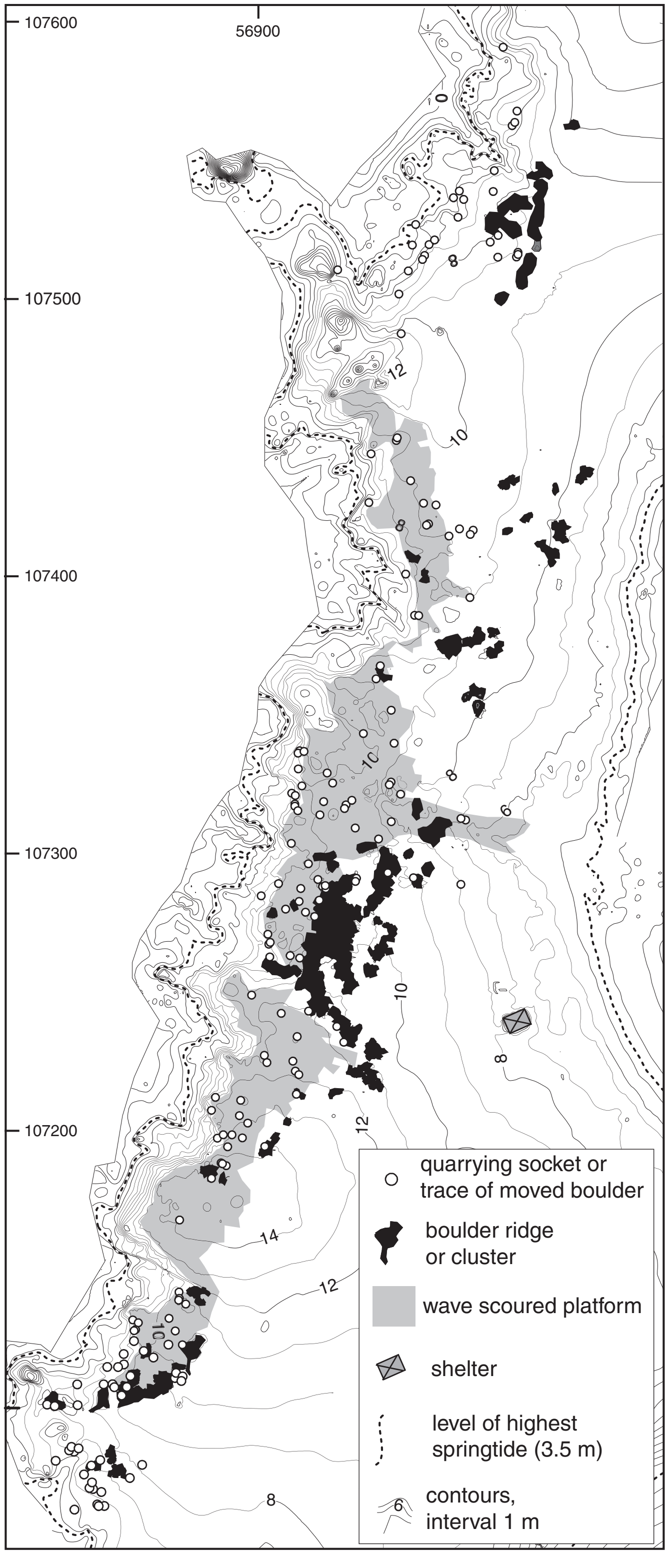

Fig. 11. Location of quarrying sockets and scars of transported CTSDs occurred during the 10 March 2008 storm on Banneg Island.

storm (Fig. 13a and b). At the rear, however, there was a considerable change in topography. Blocks 1 (1.93 t) and 2 (0.42 t), originally lying on the lee side of ridge #28 were moved, whereas blocks X, X1 and X2, of which X is on the exposed side of this accumulation, were not moved (Figs. 10c and 14). The exposed face of cluster #24 located in the middle ridge lost some smaller blocks and clast #4 (7.25 t) was taken from the cliff edge and upturned (Figs. 12b and 14c). At the same time, this accumulation grew with the addition of blocks of all sizes, and several others were transported and scattered on the turf in front of it (Fig. 13). The volume of this accumulation thus increased. Finally, on cluster #23, the third ridge, which was sheltered by the two seaward ridges, a block weighing 1.2 t was lifted and thrown 5 m from its original position. In all cases, the blocks were displaced in the same direction as the incident waves (WNW–ESE). In this area, because the blocks were selectively removed from lee-side positions, it is unlikely that they were moved by a torrential flow. It is likely that water from one or several breaking waves were thrown and projected from the cliff top, then became airborne before landing behind ridge #28 located in front line. They then rebound once, lifting one clast from the top of cluster #24 in the second ridge, then crashed again on cluster #23, located 55 m from the edge of the cliff.

These two examples show that during the March 2008 storm event, the front ridges were locally left unscathed whereas ridges further inland were re-worked.

Table 1
Location of quarrying sockets with respect to the edge of the cliff (see Fig. 12).

Sector	Number of sockets on the cliff slope	Number of sockets on the wave-scoured platform	% of sockets above cliff-top edge
South	26	26	50
South-centre	3	25	90
Centre	15	41	73
North-centre	4	13	76
North	20	7	25
Total	68	112	62

5.4.2. During the 1989–90 winter storms

Naturalists surveys (ornithology, biology, etc.) and a weekly visit by a ranger certify that only two truly morphogenetic events have occurred over the past 50 years: the 1989–90 winter storms (Fichaut and Hallégouët, 1989) and the 10 March 2008 storm. For the centre area of the island, photographs from 1979, 1990, 2005 and 2008 are available and provide before and after images of both storm events.

During the 1989–90 storms, cluster #28 was the first ridge back from the edge of the cliff and was, once again, barely affected. As shown in Fig. 13a, taken after the storm, most blocks were covered with lichens, indicating that they had not moved and had not been upturned (this does not preclude the possibility that some blocks could have fallen right-side up after having been displaced). At the back of the cluster, blocks X, X1 and X2 remained in place but blocks strewn on the turf were displaced and cluster #24 in the second ridge underwent a profound morphological change (Fig. 14a and b). On this second ridge, many blocks disappeared or were upturned, the many lichen-free clasts – thus deposited upside down – suggest that they arrived recently (Fig. 13c). In parallel, this cluster greatly increased in volume (Fig. 14a and b). As for the March 2008 storm, the observations made on accumulations after the 1989–90 storms illustrate that when accumulations form parallel series, the most seaward ridges were only slightly affected. At the same time, morphological changes and changes in volume were great in the second ridge and locally, as in the centre of the island, a third, [more inland] ridge was constructed. Using these observations, it was possible to formulate a spatio-temporal model of cluster formation and development on Banneg Island.

6. Spatio-temporal model of cluster formation

The model proposed here is based on an analysis carried out in two steps. The analysis was first performed for the centre area of the island where the most detailed data were available and then extended to the rest of the island.

Table 2
Size and weight of quarried and transported blocks during the 10 March 2008 storm event (mean values and standard deviations noted with an asterisk (*) were calculated without the largest clast that weighs 42.64 t).

Sector	Number of measured clasts	Clast size	a-Axis	b-Axis	c-Axis	Volume (m^3)	Weight (t)
North	11	Largest	1.6	1.3	0.8	1.66	5.16
		Smallest	0.4	0.3	0.2	0.02	0.07
Centre-north	4	Largest	1.1	0.8	0.65	0.57	1.77
		Smallest	0.8	0.8	0.2	0.13	0.40
Centre	17	Largest	3.6	1.3	0.5	2.34	7.25
		Smallest	1	0.5	0.08	0.04	0.12
Centre-south	9	Largest	1.5	1.5	0.35	0.79	2.44
		Smallest	0.6	0.6	0.2	0.07	0.22
South	22	Largest	4.6	2.3	1.3	13.75	42.64
		Smallest	0.6	0.3	0.2	0.04	0.11
Total	63	Largest	4.6	2.3	1.3	13.75	42.64
		Smallest	0.4	0.3	0.2	0.02	0.07
		Median	1.1	0.7	0.3	0.23	0.72
		Average	1.26/1.21*	0.77/0.75*	0.35/0.33*	0.6/0.39*	1.87/1.22*
		Standard deviation	0.44/0.38*	0.28/0.26*	0.13/0.12*	0.61/0.31*	1.9/0.98*

In the centre of the island, clusters and ridges make up a volume of 560 m^3. The isolated blocks scattered on the turf inland as well as those at the head of the gully further increase the volume of CTSDs. This lithic material comes from repeated quarrying from the cliff, over a linear distance of 80 m and almost exclusively at the upper part of the cliff. The formation of CTSDs is part of the ongoing morphological evolution whose first stage occurred when the cliff was steeper than it is today, with a higher cliff top and more seaward. At this stage extreme water levels did not overtop the cliff (Fig. 15a). It is only once the cliff retreat reached the eastern slope of the island that the waves were able to overtop the cliffs and quarry blocks from the cliff-top platform. Then, erosion created and augmented cluster #28 that was initially further from the edge of the cliff than it is currently (Fig. 15b). The upper part of the cliff face then started to recede, causing a gradual decrease in the steepness of the cliff face. This change in steepness causes breaking at a lower angle than before dislodging clasts and allowing water to surge up to overtop the cliff edge as a bore or as a partially airborne flow and pass over the first ridge, dropping some of the quarried blocks on the seaward side of the ridge. More re-working occurs inland where the overtopping water falls and a second ridge is formed (Fig. 15c). As the steepness of the cliff face continually decreases, eventually more and more wave water passes over the first ridge. This model thus explains the re-working and the increase in volume of cluster #24, as well as the formation and evolution of cluster #23 in the third ridge, born of a second wave rebound (Fig. 15d).

In various points around the island, the position and the morphology of the clusters reflect the different stages of this evolution. Initial stage is found at Skloum Maurice were the cliff-top is over 14 m asl. The first step in construction of CTSDs exists in centre-south area 2 (Fig. 10d). Here, the cliff has a slope that exceeds 50%. The cliff edge is always higher than 11.5 m asl and can locally reach 14 m. Behind the edge of the cliff, clusters make up a single string of blocks (clusters #32 to #48). At the highest points of the cliff, clusters are small and close to the edge (cluster #48); at the lower points, accumulation is larger, further inland and may form a ridge (ridge #32).

A double ridge, such as that found in the southern part of the island, is the next step. As described above, this type of structure is represented by clusters #36 and #39, which are located at the rear of a cliff rising 9 m with a slope of between 30 and 40%. During the March 2008 storm, cluster #36 located in the front, seaward ridge did not undergo any morphological changes whereas cluster #39 was greatly modified. Some of the blocks that were mobilised accumulated at the rear or form the outline of a third ridge.

The three-ridge formation that we described in the centre of the island is the third and final step in this model. This type of formation is also found in the centre-south 1 area. In this area, the slope of the cliff is of the same order as at the centre (30 to 40%) but the cliff face is higher (11 m). On 10 March, only some blocks were displaced in this formation.

Beyond a double- or triple-ridge structure, an even more evolved stage can be observed as at the southern tip of the island (clusters #38 and #39). Here, and particularly at cluster #38 (Fig. 11e), the slope of the cliff is very regular and reaches only 30% and the barely distinguishable cliff edge extends directly into the sloping platform at the rear of the cliff. Cluster #38, which made up the most seaward ridge, was completely destroyed on 10 March and the blocks from this cluster either joined cluster #39, located behind #38, or passed over it. It is likely that foremost ridges will eventually be eroded and thereby supply the clusters located more inland (Fig. 15c and d).

Finally, as the platform at the rear of the cliffs slopes to the east throughout the island except in the north, cliff height decreases as the cliff retreats due to erosion of the platform as well as to the general sagging topography. Eventually, as in the gully, a growing number of waves will overtop the island without breaking and generate flows that break up the existing clusters and entrain them towards the east coast (Fig. 15E and F).

7. Discussion

7.1. Quarrying zones and clast production

The position of 180 sockets of clasts quarried during the 10 March 2008 event supports the hypothesis that only the upper part of the cliff face and the rocky cliff-top platform at the rear supply the clast material for the accretion of clusters and boulder ridges. This distribution also shows that the supply zone is located above the highest spring tide levels, corroborated by the absence of rounded or worn blocks in the higher accumulations. CTSDs do not come from the intertidal area. This is also confirmed by the granulometry of the displaced and/or deposited blocks, which are mainly angular clasts. The spatial distribution of quarrying zones and clast production on Banneg Island differs slightly from what has been observed at other sites. Noormets et al. (2002, 2004) showed that quarrying blocks can occur on shore platforms near sea level. Similarly, Hall et al. (2006, 2008) indicate that the whole cliff face could supply clasts under some circumstances, but that the bulk is quarried from the cliff top and cliff-top platform. Quarrying processes are different on Banneg Island due to the structural characteristics of the leucogranite bedrock and the variations in the joint pattern. The role of jointing in the quarrying process is not specific to granite; it also occurs in calcareous rock, limestone, sandstone, ignimbrites or other volcanic rocks (Etienne and Paris, 2010; Hall et al., 2008; Jones and Hunter, 1992; Naylor and Stephenson, 2010; Noormets et al., 2004). As demonstrated for

Fig. 12. Largest eroded CTSDs during the 10 March 2008 storm event. a) In the south of Banneg Island, a clast weighing 42 t was deposited upside down 7 m inland. b) On the landward side of ridge #24 located at about 50 m inland from the cliff edge, a tabular clast weighing up to 7.2 t was turned over. Morphological changes of clusters due to the 10 March 2008 storm event. c) Ridges #35 and #39 in the southern area in 2005. d) Ridges #35 and #39 in the southern area in April 2008. Note the deposition of new clasts. e) and f) Morphological changes of ridge #39 between June 2006 (e) and April 2008 (f). 1: Removed clasts. 2: Deposited clast.

Banneg Island, at the headlands on which granite outcrops are found, or at the base of the cliffs everywhere else, massive jointing results in individual, large, box-shaped clasts (Fig. 3a and b). When these clasts are dislodged, they lie on the shore platform where they form a pavement of more or less rounded boulders, each weighing several tons (Fig. 3b). However, these large rounded boulders do not contribute at all to the supply of boulder clusters for CTSDs because waves cannot project them over the cliff. As the observations made after the 10 March 2008 storm show, the absence of lichen on their surfaces indicate that these boulders undergo only slight displacement during highly dynamic events.

7.2. The dynamics of cliff retreat

Jointing in the granite also controls the outline of the coastline (because it determines the rate at which the cliff retreats) and the morphology of boulder accumulations. Thus, embayments are characterised by dense jointing of the rock that weakens the cliff face at a

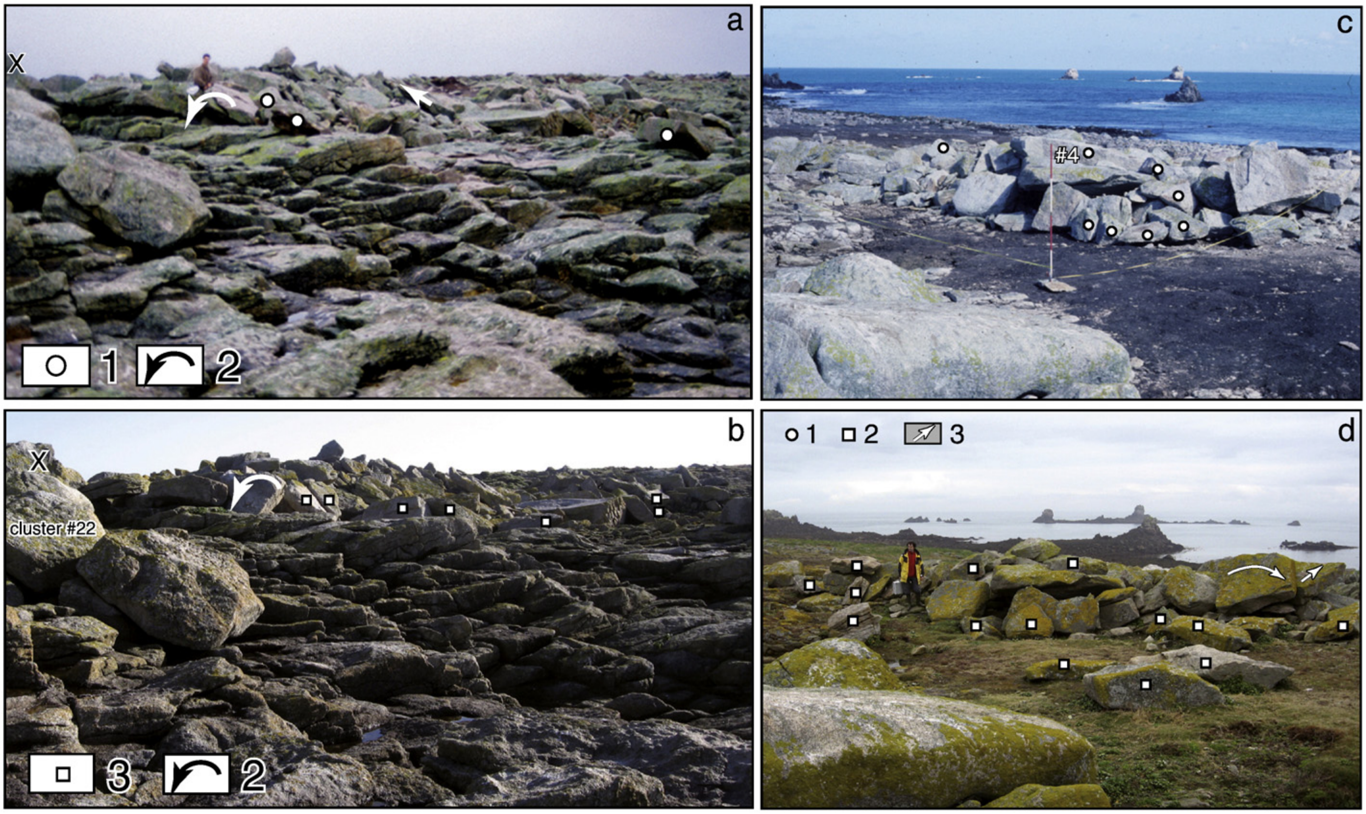

Fig. 13. Morphological changes of the exposed face of ridge #28 between April 1990 (a) and April 2008 (b). 1: Clasts removed during the storm. 2: Clasts that were displaced during the storm. 3: Clasts that were deposited during the storm. Morphological changes of the exposed face of ridge #24 between 1990 (c) and 2008 (d). 1: Blocks removed during the 10 March 2008 storm event. 2: Blocks deposited during the 10 March storm event. 3: Blocks displaced during the 10 March 2008 storm event.

height much lower than at the headlands. Retreat of the cliff-top is thus greater than at the headlands and is accompanied by a higher clast production rate. The largest boulder clusters are thus found at the rear of these [embayment] areas. In contrast, the headlands correspond to areas where the fractures are organised in a loose orthogonal system, rendering the rock more resistant to marine erosion. In these areas, the boulder accumulations (when there are any) are rather small. Thus, erosion – and thereby cliff retreat – occurs primarily around high tide periods due to the high water levels. In these conditions, the morphogenetic action of waves is much more efficient at the top of the cliff, which is weakened by the denser joint system. Similar types of processes have been described in the Shetland Islands for the Grind of the Navir cliffs (Hall et al., 2008). In this case, the erosive action of waves is at its greatest from the middle of the cliff face up to the cliff-top platform. However, the cliff maintains its sub-vertical or vertical slope. In the Shetland Islands, the energy of the breaking waves on the upper portion of the cliffs leads to the widening and propagation of the joints down to the base of the cliff that retreats as well (Hall et al., 2008). This process maintains the sub-verticality of the cliffs. On Banneg Island, while the cliff top erodes, the base seems to remain stable. This results in a decrease in the verticality of the cliff face, which is then characterised by a slope that reaches a maximum of 30%. Cliff retreat on Banneg Island contrasts with classic models whereby erosion of rock material at the lower levels causes the failure of material above, maintaining the overall verticality of the cliff. As our spatio-temporal model demonstrates, the morphology of the cliff face helps explain the development of boulder clusters and understand the dynamic processes behind their formation.

7.3. Processes of block quarrying and transport

The morphodynamic observations carried out after the March 2008 storm show that quarrying and transport processes on Banneg Island followed the two models constructed by Hansom et al. (2008). In the centre of the island, the height of breaking waves was greater at the top of the cliffs that are located between 9 and 14 m asl. The centre of the island was completely submerged by bore flow that flowed out towards the eastern coast of the island following the inland slope. On either side of the gully that forms the lowest point in this area, the highest waves resulted in bore flow, several meters thick, that was strong enough to remove pre-existing deposits, excavate a pit in the Porz ar Bagou cove and deposit several tons of blocks 40 m seaward of the beach in this cove. Similar erosion related to the observed action of cliff-top bores was reported by Hall et al. (2008) and modelled by Hansom et al. (2008).

South of the gully, although submerged, particularly in the centre-south as indicated by scoured turf at more than 150 m landward of the western cliff (Fig. 7), the waters were not that high. In this area, blocks were not transported by bore flow, but were thrown after being quarried. This second mechanism of transport has been described by Hansom et al. (2008) and Noormets et al. (2004) as being related to the pressure exerted by the breaking waves on the bedrock when the height of the waves is lower or equal to the height of the cliff. With the exception of the gully area at the centre of the island, this block projection process occurred at all other locations on Banneg Island.

The formulas given by Nott (2003) to characterise the hydrodynamic context that promotes block displacement are not applicable in this study. The "submerged boulder scenario" is not applicable here because none of the boulders was removed from the submerged zone. Regarding the "the joint-bounded block scenario", the formula can only be applied when a block is bounded by joints on all sides. On Banneg Island, blocks were quarried from the cliff-top ramp or platform, and therefore were not joint-bounded on the seaward side (Fig. 5a and b). In this situation, one side of the blocks experiences the impact of the wave and drag force is therefore involved. For the third

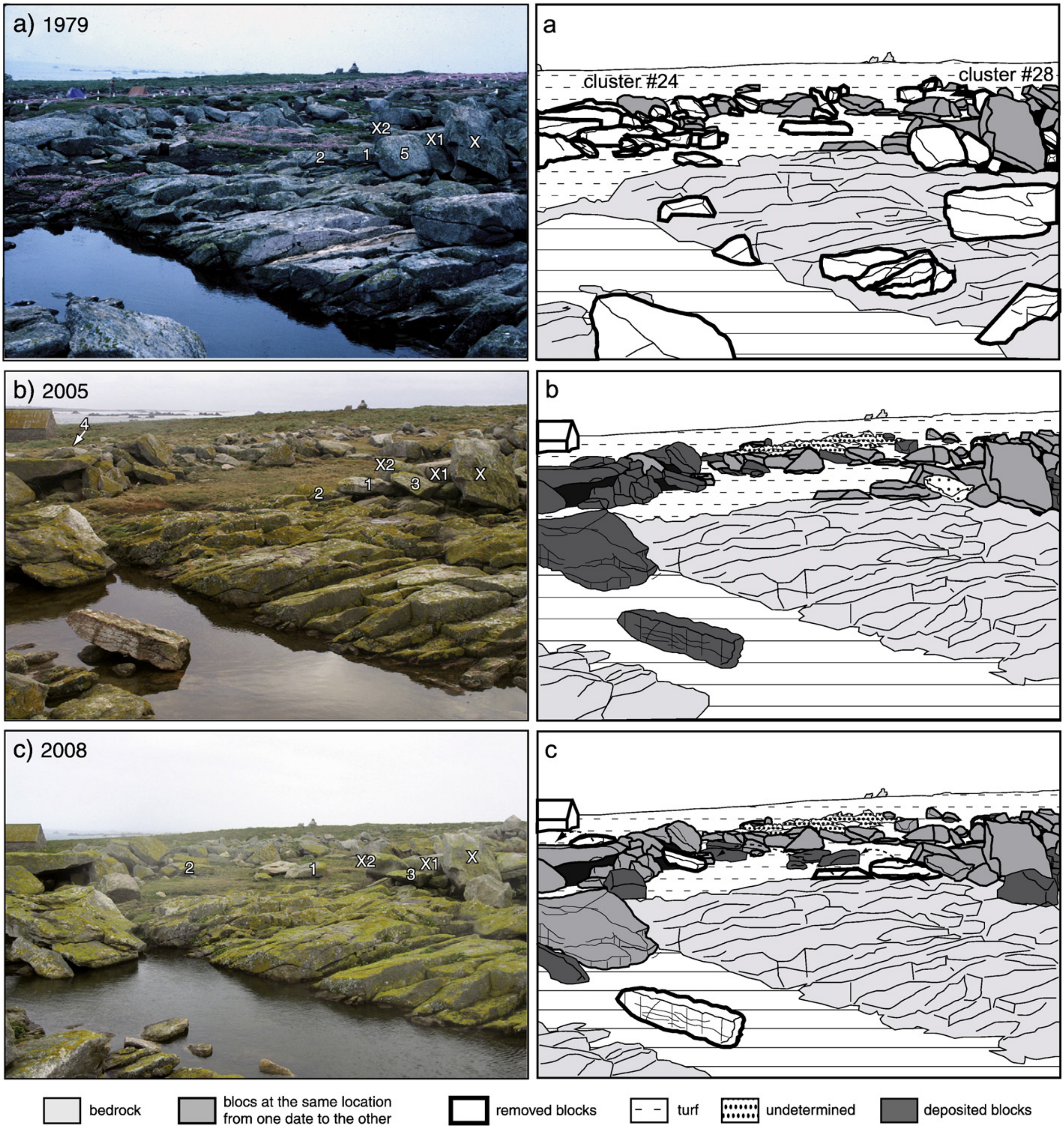

Fig. 14. Morphological changes of the lee side of ridge #28 and of the exposed side of ridge #24 in the centre of Banneg Island, between 1979 and 2008. a) Blocks removed during the 1989–90 winter storms. b) Blocks deposited during the 1989–90 winter storms. c) Changes that occurred during the 10 March 2008 storm event.

case, “sub-aerial block scenario”, the formula only refers to boulders moved by waves breaking at their pre-transport location. It cannot be applied to boulders that have *(i)* slid, *(ii)* been tossed or thrown, *(iii)* or transported in a bore. It happens that these are the types of transport that prevail during storm events on Banneg Island.

7.4. Model for the temporal evolution of deposits

Field data obtained after the 10 March 2008 storm on the position and the number of clusters and ridges helped to determine a relative chronological sequence of their formation. This sequence is different from those proposed previously. Nott (2000) analysed a series of nine ridges and showed that there was no ordered chronological sequence for their deposition. In our case, the number and temporal sequence of ridge deposition depend on the stage of cliff retreat. When the slope of the cliff face is nearly vertical, there is generally only one ridge or one train of clusters at the cliff top and it is usually well to the rear of the cliff. As the slope of the cliff decreases with time, waves tend to be projected against the cliff face at an angle and break progressively further to the rear of the cliff where they may develop a second or

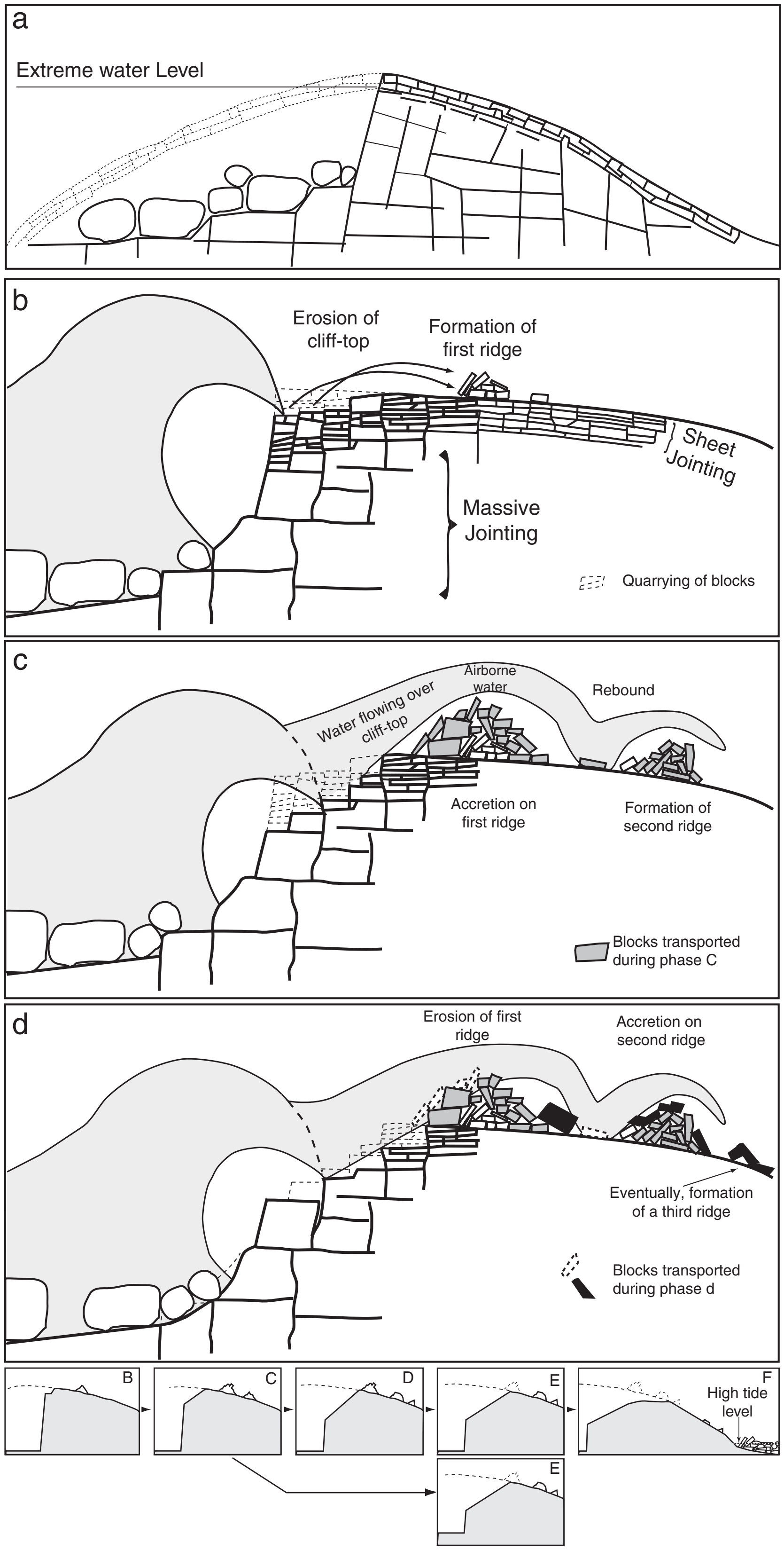

Fig. 15. Spatio-chronological model of CTSDs evolution on Banneg Island.

even a third series of clusters. This notion of wave rebound has also been proposed by Hall et al. (2008) to explain boulder spreads and airthrow during the 1992 storm in the Shetland Islands. On Banneg, the accumulations that are the closest to the coast are the oldest but are eventually destroyed to supply new ridges formed at the rear. When the upper part of the cliff becomes too low, such as in the gully, the clusters are gradually eroded and blocks are deposited on the lee side of the island.

8. Conclusion

The quantitative analysis carried out in this study confirms the qualitative observations made by Fichaut and Hallégouët (1989) after the 1989–90 winter storms. The analysis also helps to improve our knowledge of the morphodynamic processes of deposition and evolution of CTSDs.

From a morphodynamic point of view, the analysis of the process of block removal showed that lithostructural conditions play an important role. These conditions promote the quarrying of slabs where the joint system is particularly dense, i.e. at the top of the cliff. The joint system in the rock influences the shape of the coastline (because it determines the rate at which the cliff retreats) and the morphology of boulder accumulations. Thus, embayments are characterised by a dense joint system in the rock that weakens the cliff face at a height much lower than for headlands and so cliff retreat is greater and clast production is higher. In contrast, the headlands correspond to areas where the fractures are organised in a less dense orthogonal joint pattern, and render the rock more resistant to marine erosion. In these areas, if any boulder accumulations occur, then they are small. Lithostructural conditions are also involved in the erosion process and in the morphology of the cliff face. Cliff retreat is more rapid near the top of the cliff than at the base and results in a decrease in the slope of the cliff face, reducing it to an incline of 30% at most, this evolution of the cliff face at Banneg Island contrasting with classic models of cliff retreat.

Finally, this study helped confirm the important role of storm events in the deposition and evolution of boulder accumulations and corroborates the observations made at higher latitudes (Aran Islands in Ireland, Orkney and Shetland Islands in northern Scotland) showing that the origin of these specific deposits are directly attributable to specific extreme marine weather events, such as storms or hurricanes rather than to tsunamis. The spatio-temporal model outlined here also helps to better understand the deposition of parallel ridge series that in many cases raises questions regarding the diachronic or synchronic processes involved. Double or triple series of accumulations can be constructed in successive stages of block quarrying, transport and deposition and result from the relationship between the hydrodynamics of breaking waves and nature of the cliff morphology that they impact.

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