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Coastal evolution on volcanic oceanic islands: a complex interplay between volcanism, erosion, sedimentation, sea-level change and biogenic production

Ricardo S. Ramalho
University of Bristol

Rui Quartau
Instituto Portugues do Mar e da Atmosfera

Alan S. Trenhaile
University of Windsor

Neil C. Mitchell
University of Manchester

Colin D. Woodroffe
University of Wollongong, colin@uow.edu.au

See next page for additional authors

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Keywords

complex, interplay, between, volcanism, erosion, sedimentation, sea, level, islands, change, coastal, biogenic, production, oceanic, evolution, volcanic, GeoQuest

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Coastal evolution on volcanic oceanic islands: a complex interplay between volcanism, erosion, sedimentation, sea-level change and biogenic production

Ricardo S. Ramalho[a,b,1,*], Rui Quartau[c], Alan S. Trenhaile[d], Neil C. Mitchell[e], Colin D. Woodroffe[f], Sérgio P. Ávila[g,h,i]

[a]School of Earth Sciences, University of Bristol, Wills Memorial Building, Queen's Road, Bristol, BS8 1RJ, UK
[b]Lamont-Doherty Earth Observatory at Columbia University, Comer Geochemistry Building, 61 Route 9W / PO Box 1000, Palisades, NY-10964-8000, USA
[c]Unidade de Geologia Marinha, Laboratório Nacional de Energia e Geologia, Estrada da Portela, Zambujal, Alfragide, Apartado 7586, 2611-901 Amadora, Portugal
[d]Department of Earth and Environmental Sciences, University of Windsor, N9B 3P4 Windsor, On, Canada
[e]School of Earth, Atmospheric and Environmental Sciences, the University of Manchester, Williamson Building, Oxford Road, Manchester M13 9PL, UK
[f]School of Earth and Environmental Sciences, University of Wollongong, Wollongong, NSW 2522, Australia
[g]Faculdade de Ciências da Universidade do Porto, Rua do Campo Alegre s/n, 4169-007, Porto, Portugal
[h]CIBIO, Centro de Investigação em Biodiversidade e Recursos Genéticos, InBIO Laboratório Associado, Pólo dos Açores & Departamento de Biologia, Universidade dos Açores, Rua Mãe de Deus, 9501-855 Ponta Delgada, Azores, Portugal
[i]MPB Marine PalaeoBiogeography Working Group of the University of the Azores, Departamento de Biologia, Universidade dos Açores, Rua Mãe de Deus, 9501-801 Ponta Delgada, Azores, Portugal

Abstract

The growth and decay of oceanic hotspot volcanoes are intrinsically related to a competition between volcanic construction and erosive destruction, and coastlines are at the forefront of such confrontation. In this paper, we review the several mechanisms that interact and contribute to the development of coastlines on oceanic island volcanoes, and how these processes evolve throughout the islands' lifetime. Volcanic constructional processes dominate during the emergent island and subaerial shield-building stages. During the emergent island stage, surtseyan activity prevails and hydroclastic and pyroclastic structures form; these structures are generally ephemeral because they can be rapidly obliterated by marine erosion. With the onset of the subaerial shield-building stage, coastal evolution is essentially characterized by rapid but intermittent lateral growth through the formation of lava deltas, largely expanding the coastlines until they, typically, reach their maximum extension. With the post-shield quiescence in volcanic activity, destructive processes gradually take over and coastlines retreat, adopting a more prominent profile; mass wasting and marine and fluvial erosion reshape the landscape and, if conditions are favorable, biogenic processes assume a prominent role. Post-erosional volcanic activity may temporarily reverse the balance by renewing coastline expansion, but islands inexorably enter in a long battle for survival above sea level. Reef growth and/or uplift may also prolong the island's lifetime above the waves. The ultimate fate of most islands, however, is to be drowned through subsidence and/or truncation by marine erosion.

Keywords: coastal evolution, oceanic island volcanoes, volcanism, erosion, sedimentation, sea-level change, biogenic production

1. Introduction

Coastlines are the ever-changing boundary between the land and the sea. They are complex threshold-driven, non-linear dynamical systems (Naylor and Stephenson, 2010) whose evolution is the product of mutual adjustments in topography and fluid dynamics in response to changes in external conditions (Wright and Thom, 1977; Trenhaile, 1997; Woodroffe, 2002). Few places on our planet experience more dramatic and rapid changes in topography and external conditions than the coasts of hotspot islands, which experience the effects of volcanism, flank collapses, and exposure to open ocean. Additionally, coastlines on oceanic island volcanoes have a clear beginning (through volcanic emergence) and a predictable end (through island subsidence/erosion), and evolve as the edifices themselves evolve. Oceanic island volcanoes are, thus, prime natural laboratories to study the different processes that interplay in a complex manner to shape coastlines.

Oceanic hotspot islands are prominent, dynamic geological features that rise from the deep seafloor by a combination of volcanic, intrusive and tectonic processes. Coastlines are established as soon as oceanic island edifices breach sea level, and they become the forefront of a raging battle that, in the long term, is lost. This confrontation is, essentially, a competition between volcanic (and biogenic) construction on one side, and erosive destruction on the other. This balance, or more appropriately this imbalance, of powers varies in space and time as island edifices evolve. In this paper we offer an overview on the main mechanisms that shape oceanic island coastlines, and how these mechanisms act differentially throughout the successive stages of island evolution. An analysis of the various factors that control coastal evolution on oceanic island volcanoes

*Corresponding author. Tel.: +44 (0)117 331 5141; fax: +44 (0)117 925 3385.
Email address: ric.ramalho@bristol.ac.uk (Ricardo S. Ramalho)
[1]Formerly at Institut für Geophysik, Westfälische Wilhelms-Universität, Münster, Germany

is also presented here and discussed, albeit in a qualitative manner.

2. Oceanic hotspot island evolution and development of coastlines

Oceanic island volcanoes frequently follow an evolutionary trend that exhibits some basic similarities across different hotspot systems (Schmincke, 2004; Ramalho, 2011). This evolutionary trend is essentially caused by variations in the rate of magma-supply through geological time. This, in turn, is a function of plate motion relative to the hotspot melting source, plate age/thickness, proximity to a plate boundary, and melt source characteristics (Ramalho, 2011). Internal factors such as those described above directly influence the distribution, style and intensity of magmatism over space and time, and the relative position of the edifices with respect to sea level. External factors also contribute to edifice evolution: environmental conditions not only control the nature and intensity of the erosive agents that act upon the edifices but also control the relative importance of biological processes in the system. The evolution of oceanic island systems is, thus, somewhat different according to the different geodynamic/geographic settings where the hotspots are located (Menard and Ladd, 1963; Menard, 1986; Schmincke, 2004; Ramalho, 2011). On fast-moving plates, the evolution of oceanic island systems is generally characterized by short edifice magmatic lifetimes and a long-term subsidence trend - creating linear age-progressive chains (Morgan et al., 1995; Lipman and Calvert, 2013); conversely, on stationary or slow-moving plates, oceanic islands tend to occur in a cluster, typically have longer magmatic lives, and barely subside or experience uplift (Carracedo, 1999; Schmincke, 2004; Menendez et al., 2008; Ramalho et al., 2010a,b,c; Madeira et al., 2010; Ramalho, 2011). Nevertheless, despite the many differences found across hotspot systems, most oceanic island volcanoes can be described through a set of main evolutionary stages. These are (see Macdonald et al., 1983; Clague and Dalrymple, 1987; Carracedo, 1999; Schmidt and Schmincke, 2000; Schmincke, 2004; Ramalho, 2011): a) seamount stage; b) emergent island stage; c) subaerial shield-building stage; d) capping or post-shield stage (alternatively considered as part of the shield-building stage); e) erosional stage; f) post-erosional or rejuvenated stage; g) atoll/razed island stage; and h) guyot or drowned island stage. In general terms, the stages reflect an intermittent magmatic life that is normally defined by a vigorous, voluminous first eruptive cycle, followed by additional eruptive cycles involving significantly lower volumes (sometimes of more differentiated magmatic products) and intercalated by long periods of volcanic quiescence (see Fig.1). In the long term, as volcanism wanes, erosion and sedimentation gradually assume a dominant role in the evolution of island landscapes. The topographical evolution of oceanic islands inevitably follows a similar trend, with islands reaching their maximum height after the first volcanic cycle and then slowly decaying (see Fig.1). Dominant clastic processes experience a transition from submarine hydroclastic and pyroclastic to subaerial hydroclastic and pyroclastic, and finally to epiclastic (see Fig.1), as the offshore marine record at the volcanic apron also shows (Schmincke et al., 1995; Schmincke, 2004).

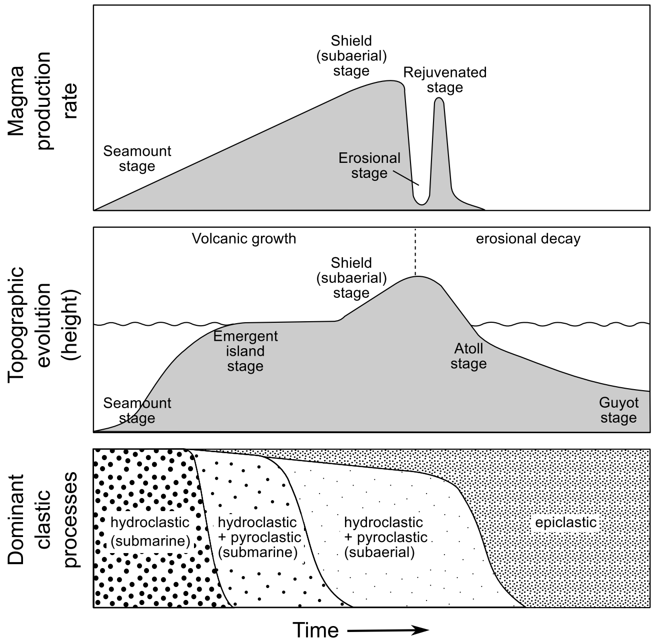

Figure 1: Evolution of magma production rate, topography and dominant clastic process in oceanic islands according to the ageing of the edifices. Modified from Schmincke (2004)

Coastal evolution on oceanic island volcanoes is intrinsically linked to the evolution of the edifices themselves. Coastlines are first established when seamounts grow above sea level. This transition between seamount and oceanic island occurs during the emergent island stage and marks the beginning of coastal evolution. Conversely, coastlines cease to exist when the island edifice is finally drowned by subsidence and/or truncation by marine erosion. The moment of drowning marks the transition between the atoll/razed island and guyot or drowned island stages, and has been termed the Darwin Point in the context of coral atolls (Grigg, 1982). The two moments - emersion and drowning - define the island's lifetime and consequently the lifespan of coastlines. During this lifespan, the processes that shape coastlines change dramatically as the edifice evolves: while volcanic construction dominates coasts during the initial stages of island building, erosional and biogenic processes gradually dominate as the edifice ages. In the following sections, we first introduce the main factors that control coastal evolution on oceanic island volcanoes and then provide a detailed overview on the main coastal mechanisms that operate on these edifices from emergence to drowning.

3. Factors controlling coastal evolution

Several factors interplay in a complex manner to shape coastlines at oceanic island volcanoes. As one or more of the factors change, feedback mechanisms immediately cause adjustments in the other factors, driving the system back towards equilibrium (though it never reaches this state). These factors are, chiefly: volcanism, tectonics, mechanical properties of shoreline lithologies, wave energy parameters, the amplitude of eustatic change, mass wasting, subaerial erosion, sediment production and availability, reef growth and biogenic production, and uplift *versus* subsidence (see Fig. 2).

3.1. Volcanism

Volcanism is the foremost agent of insular growth, and one that changes coastal outlines very rapidly and dramatically. The nature, intensity and distribution of volcanism over space and time will condition the way volcanism shapes coasts, as described in section 4. Effusive volcanism generally adds large tracts of land to island edifices, actively contributing to expanding shorelines (Peterson, 1976); paradoxically, hydrovolcanism can simultaneously build and destroy coastal landscapes, and contribute significantly to sediment production.

3.2. Tectonics

Active and inactive tectonic structures, such as fault scarps and dyke swarms, may control coastal geometry and influence erosion rates, either by triggering mass-wasting events (during catastrophic fault slips) or by their contribution to differential erosion.

3.3. Mechanical properties of shoreline lithologies

The structure and lithology of the materials composing shorelines directly influences the style and intensity of the different erosional mechanisms acting upon these landscapes. Consequently, different erosion rates may occur over several time and space scales as a function of the mechanical properties of the eroding lithologies (Trenhaile, 2011). Likewise, these properties may also influence the quantity and type of available mobile sediment in the system. The morphology of coastlines is thus indirectly influenced by the mechanical properties of the geological materials being eroded.

3.4. Wave energy parameters

Marine erosion is, together with subaerial erosion, the foremost agent of destruction on most oceanic islands, actively contributing to coastline retreat. Wave energy parameters play a dominant role in controlling the intensity of marine erosion and the rate of shoreline regression (Trenhaile, 1987, 2000, 2001). The wave energy reaching the coast is largely determined by near and far-field wind and climate patterns and bathymetrical morphology, which vary in space and time. The intensity and recurrence period of extreme-wave events like storms and tsunamis, will also have a very high impact on long-term evolution of often exposed insular coastlines. Storms and tsunamis constitute high-energy and low-frequency events that generate extensive erosion, sediment transport and deposition in a few minutes or hours and over large areas (Noormets et al., 2002; Paris et al., 2009).

3.5. Amplitude of eustatic change

The amplitude of eustatic change has a profound effect on long term coastal evolution because of its effect on marine erosion and reef growth. Insular shelves result from the combined work of wave erosion and sea-level oscillations and so their profile - in particular their width and maximum depth - experiences adjustments that are determined by the degree of variability of the oscillations (Trenhaile, 1989, 2001). Large eustatic variations allow the development of wider and deeper insular shelves and increase subaerial erosion by lowering fluvial base levels during lowstands, with indirect effects on sediment removal and coastline retreat. Conversely, small eustatic variations limit the role of marine and fluvial erosion. Eustatic changes also have a strong impact on reef accretion and development (Neumann, 1985; Paulay and McEdward, 1990; Woodroffe et al., 1999; Woodroffe, 2008).

3.6. Mass wasting

Gravitational mass wasting at different scales contributes significantly to coastal regression (Holcomb and Searle, 1991; Mitchell, 2003). Large volume, catastrophic lateral flank collapses may remove huge volumes from island edifices, dramatically changing coastal outlines and reseting marine erosion (Mitchell, 2003). Small, either gradual or catastrophic, mass movements are simultaneously one of the most effective mechanisms of cliff erosion and an important source of coarse sediment (Griggs and Trenhaile, 1994; Bird, 2008; Trenhaile, 2011). Small inland landslides may both dam river valleys and temporarily interrupt riverine sediment discharge or contribute to a catastrophic delivery of sediments to coasts. The stochastic occurrence of mass wasting further contributes to the threshold-driven behavior of rocky coastal evolution.

3.7. Subaerial erosion

Subaerial erosion directly affects coastal evolution because it contributes to cliff erosion and plays a decisive role in the cycle of sediment production, transport and delivery to coastal regions (mainly through stream solid discharge) (Wentworth, 1927; Draut et al., 2009; Ferrier et al., 2013). Coastal morphology (rocky vs sandy) is thus influenced by subaerial sediment-producing mechanisms, and subsequent erosion, redistribution and re-sedimentation, essentially by marine mechanisms and more rarely by eolian processes. Fluvial erosion stands as one of the major contributors to the topographical decay of oceanic islands (especially in high islands where orographic rain occurs) and to the delivery of sediments to coasts (Draut et al., 2009; Ferrier et al., 2013); the regime of fluvial erosion - perennial vs ephemeral and torrential - influences the distribution of sediment delivery in space and time, and consequently has an impact on coastal processes. In systems in which substantial amounts of fine sediments are available at coastlines, eolian processes can likewise play a significant role in coastal evolution, forming coastal sand dunes.

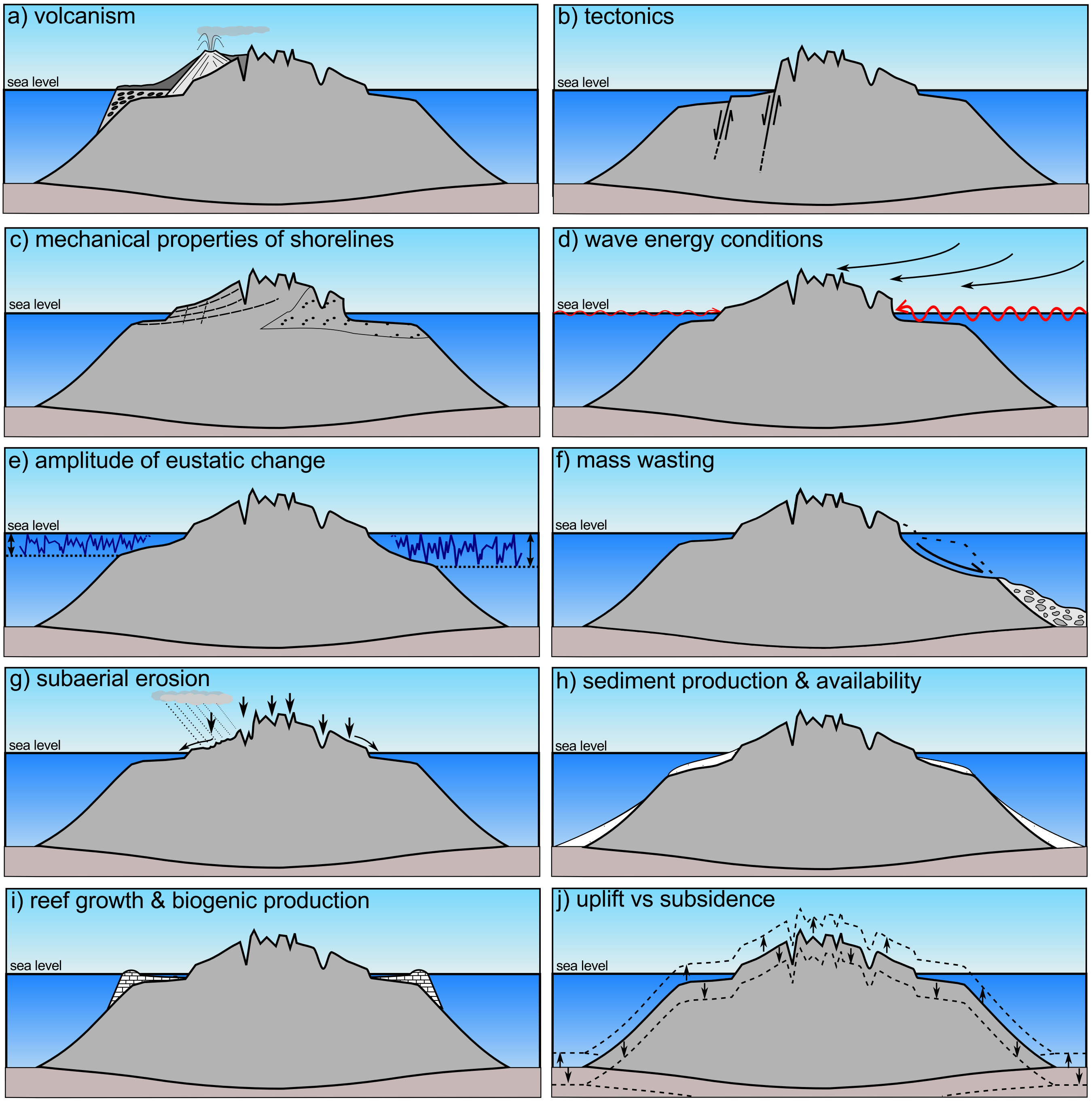

Figure 2: Factors controlling coastline evolution on oceanic islands.

3.8. Sediment production and availability

Subaerial and marine erosive processes, together with volcaniclastic and biogenic processes, produce mobile sediment that is transported to and along coastlines. Mobile sediments simultaneously act as abrasives, further enhancing cliff retreat on rocky coasts, or act as barriers to erosion when available in enough quantities for, and when dynamic conditions allow, beach formation. Sediment production and availability will thus influence coastal evolution. Sediment production greatly depends on existing sources such as explosive volcanism, high riverine discharge or the existence of large biogenic structures (e.g. coral reefs). Volcaniclastic sediment production is typically maximized during periods of high hydrovolcanic input, chiefly during the emergent island stage or during post-erosional stages. In contrast, biogenic sediment production generally increases with edifice age; as edifices and their coasts evolve, biogenic structures increase in size and complexity, progressively contributing larger amounts of skeletal remains. Riverine sediment discharge also tends to increase with edifice age until a threshold is reached, whereby subaerial erosion - and consequently terrigenous discharge - necessarily diminish with topographical decay. This threshold is normally achieved when an island topography has been reduced to below the critical height for orographic rainfall and/or when river profiles reach a gradient that inhibits sediment transport to the sea (except during storms) (Menard, 1986). Overall, as island edifices age, sediment production tends to increase, resulting in a gradual transformation from rocky to sandy coasts.

3.9. Reef growth and biogenic production

Coral growth is restricted to regions with favorable oceanographic conditions, typically in tropical and subtropical waters away from large freshwater, mud and dust inputs (Stoddart and Steers, 1977; Chappell, 1980; Spalding and Grenfell, 1997; Dullo, 2005). Coral reef growth, however, has a profound influence on coastal evolution: coral reefs protect island edifices from marine abrasion (incuding during storm surges and tsunamis), provide a barrier that allows significant sediment accumulation within island shelves, and frequently are the most important sediment supplier on the edifices in which they grow (Kennedy and Woodroffe, 2000; Kunkel et al., 2006; Gelfenbaum et al., 2011). Organic growth in general and coral reef growth in particular gain prominence as edifices age (Scott and Rotondo, 1983). As volcanism wanes and marine erosion creates wide shallow-water shelves, coral growth is promoted and fringing and barrier reefs develop. Gradually, as volcanic edifices are eroded and their topography is lowered, terrigenous discharge diminishes and so reduces the constraints it imposes on coral growth, allowing the development of larger reef structures. Larger reefs also provide more carbonate sediment for organisms and increase reef biodiversity, boosting biogenic production to the point that, in very old edifices in tropical waters - such as atolls - coral growth is the single most important agent of insular growth, effectively prolonging the island's life above sea level. In fact, organic growth is the source of all the materials that compose an atoll (Ladd et al., 1950). In islands where oceanographic conditions are unfavorable for reef growth, biogenic processes nevertheless play an important role in sediment production albeit at a much slower pace. On these edifices, the proportion of biogenic elements in mobile sediments increases as a function of edifice age.

3.10. *Uplift* vs *subsidence*

Vertical movements may have a profound impact on coastal evolution. Vertical movement will have a direct effect on coastline extension, increasing or decreasing the length of the coast as uplift or subsidence respectively occur. Second, vertical movements also have a direct effect on the erosional, sedimentary and biological processes acting upon shores due to changes in relative sea level. The growth and morphology of coral reefs, for example, is strongly influenced by vertical movements affecting the island edifices (Chappell and Veeh, 1978; Scott and Rotondo, 1983; Toomey et al., 2013). In fact, very large reefs such as those in ring atolls are only possible if an optimal subsidence rate is met (Chappell and Veeh, 1978). Subsidence typically reduces island size and consequently reduces the lifetime of the edifice above sea level. With subsidence, coastal retreat is facilitated not only because of the sub-conical shape of the edifice but also because marine erosion typically increases in intensity as insular shelves get deeper and wave energy dissipation diminishes. The acceleration of marine erosion consequently results in higher coasts, intensifying cliff erosion by means of small mass wasting, further contributing to coastline retreat. This retreat, in turn, may even help to maintain the intensity of subaerial erosion by keeping river profiles steep despite the general tendency for subsidence to raise base levels. Uplift, on the other hand, almost invariably results in the expansion of coastlines and consequent extension of the islands' lifetime above sea level. Uplift exposes island shelves and their sediment cover and so may induce a long term change in coastal morphology from high rocky shores to low sandy ones. On coral islands, however, uplift may trigger a transition to rocky shores with karstified limestone cliffs. Uplift also creates a coastal terrace morphology that allows for sediment recycling (from former raised beaches), further contributing to coastal sedimentation and reduced marine erosion. Uplift may be responsible for accelerating subaerial erosion by lowering riverine base levels, gradually contributing to topographic decay. Sediment accumulation over shallower, flatter coastal areas and increased wave attenuation over shallower insular shelves, however, will provide further protection to the coast from erosion.

Most oceanic island volcanoes are subjected to long term subsidence, initially driven by surface flexural loading (Walcott, 1970; Watts and ten Brink, 1989) and then mostly by plate cooling as a function of age (Stein and Stein, 1992), and hotspot swell decay (Detrick and Crough, 1978; Morgan et al., 1995; Ramalho, 2011). Subsidence associated with the initial flexural loading imposed by fast-growing volcanic edifices typically amounts to several kilometers and happens at very fast rates (e.g. 2.6 mm/yr for Hawai'i) for the duration of volcanic growth (Brotchie and Silvester, 1969; Walcott, 1970; Watts and ten Brink, 1989; Ludwig et al., 1991; Lipman, 1995; Watts, 2001). Subsidence associated with flexural loading is greatly dependent on several factors, chiefly the rheological parameters of the lithosphere (such as the elastic thickness), the viscosity of underlying mantle, and parameters that control the proportion of the magma that rises to the surface *versus* the amount of magma that gets stalled at different levels within the lithosphere (see Brotchie and Silvester, 1969; Watts and ten Brink, 1989; Watts et al., 1997; Watts, 2001; Minshull and Charvis, 2001; Ali et al., 2003). Rheological parameters greatly depend on plate age/thickness and the amount of thermal rejuvenation exerted by hotspots, with older and thicker, colder plates stiffer than young ones (Watts, 2001). The role of intrusions in modifying plate flexure is well known (e.g. Watts and ten Brink, 1989; Watts et al., 1997; Ali et al., 2003) but extremely dependent on where these intrusions occur, i.e. if they act as subsurface loads (when they occur at the base of the rigid plate), if they are neutral (if they occur within the rigid plate), or act as surface loads (if they occur within the island edifice) (Ramalho et al., 2010a; Ramalho, 2011). Moreover, little is known on about the factors that control the relative importance of intrusive *versus* extrusive processes on hotspot systems, although it is suspected that plate velocity with respect to the melting source may be important (Ramalho, 2011). Subsidence by plate cooling and hotspot swell decay operate at much longer time scales (Morgan et al., 1995). Plate thermal subsidence typically follows a known depth/age relationship with rapid-to-moderate rates for very young lithosphere that asymptotically decays towards vertical equilibrium which is attained when lithosphere reaches 100-120 Myrs (Stein and Stein, 1992). The component of subsidence associated with hotspot swell decay is, however, very poorly constrained by observational data and numerical models vary in their estimative (e.g. Detrick and Crough, 1978; Sleep, 1990; Morgan et al., 1995) although it seems to be similar in scale to plate thermal subsidence.

There are, however, a number of circumstances under which islands can be uplifted (although our knowledge concerning these processes is still in its infancy). For example, islands located on plates that are stationary with respect to their melting source, such as the Cape Verdes, are prone to long term uplift movements (up to 400-450 m at 0.4 mm/yr) possibly driven by basal intrusions and by hotspot swell growth (Madeira et al., 2010; Ramalho et al., 2010a,b,c; Ramalho, 2011). Cumulative far field effects of surface loading by other islands in the vicinity may also cause uplift, as is inferred for O'ahu (0.02-0.07 mm/yr), Moloka'i (0.04-0.19 mm/yr), and Lana'i (0.15-0.29 mm/yr) in Hawai'i, and for other Pacific uplifted atolls in the Cook-Society island region (McNutt and Menard, 1978; Grigg and Jones, 1997; Rubin et al., 2000; McMurtry et al., 2010). Likewise, islands located near active plate margins may experience uplift as a result of dynamic topography (near divergent plate margins) or plate flexure/buckling associated with outer trench rise (near convergent plate margins) (Karig et al., 1976; Melosh, 1978). The role of flexural rebound associated with mass wasting and erosional unloading may also add a modest contribution to uplift, as has been suggested to some of the Hawaiian and the Canary Islands (Smith and Wessel, 2000; Menendez et al., 2008). A detailed appraisal of oceanic island uplift mechanisms, magnitudes and rates is outside the scope of this paper; for that we recommend the reader to works such as Smith and Wessel (2000); Zhong and Watts (2002); Ali et al. (2003); Klügel et al. (2005a); Ramalho et al. (2010a,b); Madeira et al. (2010); McMurtry et al. (2010); Ramalho (2011).

4. Volcanic growth and synvolcanic erosion and sedimentation

Oceanic hotspot islands are essentially very large polygenetic volcanic edifices and so volcanism is, inevitably, the foremost agent of insular growth and shoreline expansion. The style of volcanism, however, dictates the type of morphology created by volcanic activity. Eruptive style, in turn, is mainly controlled by magma composition and the mixing mass ratio of water/magma in the system (Wohletz and Sheridan, 1983); since hotspot magmas are overwhelmingly basaltic in composition, it is mostly the water/magma mass ratio in the system that controls eruptive style on oceanic island volcanoes. Hence, the position of active eruptive vents with respect to sea level plays a crucial role in controlling eruptive style and the type of structure and morphology created along oceanic island coasts. As edifices grow above the ocean surface, the increasing distance between the main eruptive vents and the coast reduces the water/magma ratio in the system (unless there is interaction with ground or inland water bodies), changing the dominant eruptive style from highly explosive hydromagmatic to a more effusive magmatic. As a consequence, processes of volcanic growth will also be different according to the stage of evolution of an edifice.

4.1. Emergent island stage

The emergent island stage corresponds to the transition between seamount and emersed island, typically a shield volcano (Schmidt and Schmincke, 2000, 2002; Schmincke, 2004). This stage is characterized by the onset of surtseyan activity, as the edifice shoals (Thorarinsson, 1967; Kokelaar, 1983, 1986; Moore, 1985; Schmidt and Schmincke, 2000; Ramalho, 2011). At decreasing depths, volatile pressure starts to overcome the hydrostatic pressure allowing the nucleation of bubbles, an increase in vesicularity and the onset of magma fragmentation resulting in increasingly higher explosivity (Thorarinsson, 1967; Kokelaar, 1983; Wohletz and McQueen, 1984; Moore, 1985). This phase - informally termed the "jacuzzi" phase due to the characteristic bubbling and upwelling - marks the transition from submarine effusive eruptions to increasingly more explosive hydromagmatic eruptions. Gradually, steam-driven cupressoid tephra-finger explosions (characterized by wet surges and mass flows) start to disrupt the sea surface with increasing intensity, producing large quantities of hydroclastic and/or pyroclastic material that accumulate around the active vent and slowly bring it to shallower levels (Kokelaar, 1983; Moore, 1985; Kokelaar, 1986). The resulting structure is a tuff cone that will eventually breach the sea surface and form a precursory small pyroclastic island, as it happened at Capelinhos (Azores) and Surtsey (Iceland) (Thorarinsson, 1967; Kokelaar, 1986; Schmidt and Schmincke, 2000; Cole et al., 2001). As the cone builds up, the explosion crater rim rises above the sea surface further limiting the influx of water able to flow over or percolate through the cone to the explosion centre where it is converted to steam (Moore, 1985). Gradually, influx of sea-water to the active vent diminishes, affecting the mixing mass ratio of water/magma. As the latter decreases, explosivity is intensified, and a continuous set of uprush explosions ensue (Moore, 1985), adding large quantities of fallout tephra to the existing cone (Moore, 1985; Sohn, 1995). Eventually, if the mixing mass ratio of water/magma reaches an optimum for powerful hydroexplosions (0.3-0.8 mass ratio), the system may exhibit taalian-type hydrovolcanic activity, albeit very transient (Wohletz and McQueen, 1984; Sohn, 1995). With complete isolation of the vent from the sea, water in the system is completely consumed and the eruptive style changes very rapidly to strombolian/hawaiian (Kokelaar, 1983; Moore, 1985; Kokelaar, 1986). This change in eruptive style - if sustained and followed by high extrusion rates - allows the transition to the subaerial shield-building stage (Kokelaar, 1986; Schmidt and Schmincke, 2000). However, if extrusion rates are low, or if extrusion rates cannot compensate ongoing marine erosion, the system may change back to surtseyan. Thus, active emergent volcanoes are very unstable systems that rapidly change in eruptive style, morphology and size, according to fluctuations in extrusion and erosion rates. Unless high extrusion rates are sustained for a period long enough to allow the transition to a shield volcano, the emergence of a volcano above sea level may be punctuated by numerous setbacks, requiring a long time to produce a more permanent island (Schmidt and Schmincke, 2002); in reality, few volcanoes get to the island stage (Schmincke, 2004).

4.1.1. *Precursory stages of island growth by surtseyan activity*

Surtseyan eruptions build precursory islands that may or may not develop into larger shield volcanoes. These proto-island edifices generally correspond to low-aspect ratio morphologies - cones or rings - composed of poorly sorted, poorly stratified and poorly consolidated hydromagmatic fine glassy tephra (see Fig. 3a). Consequently, wave erosion is very efficient in destroying these poorly consolidated structures, creating shallow marine abrasion surfaces (Schmidt and Schmincke, 2002); currents and wave-induced shear stresses on the bottom may also contribute to the destruction of emergent island edifices down to considerable depths (at least down to storm base). Thus, because emergent surtseyan cones are usually very prone to marine erosion, their existence may be very ephemeral (Schmidt and Schmincke, 2002), varying from days to a few tens of thousands of years. Reports of such ephemeral existences are common, e.g.: an eruption in D. João de Castro Bank (Azores), during December 1720, produced an ephemeral island that attained a length of 1.5 km and an elevation of about 250 m above sea level (asl) before it was eroded beneath the sea two years later, the summit presently being at a depth of 14 m (Agostinho, 1931; Weston, 1964); in 1811, an eruption offshore Ponta da Ferraria (São Miguel, Azores) created a 90 m high island - named Sabrina after Captain Tillard from *HMS Sabrina* "claimed" the new land for the British Crown - but was eroded in a matter of days (Madeira and Brum da Silveira, 2003); the top of Surtla (Iceland) has been reduced by erosion from near sea level to a depth of 45 m in just 17.5 years from its formation (Kokelaar and Durant, 1983; Moore, 1985).

The newly established coastlines on precursory islands are thus a complex product of a number of processes that change very rapidly in relative importance and intensity, over space and time. During the emergent island stage - more than on any other stage of island evolution - volcanic processes are simultaneous agents of construction and destruction, imprinting a vigorous dynamic in the system. Due to the interaction with water, eruptive mechanisms are essentially volcaniclastic and vary from hydroclastic to submarine and subaerial pyroclastic (Schmidt and Schmincke, 2000). Simultaneously, epiclastic processes also contribute significantly to the production of mobile sediment, owing to efficient erosion during this stage. The dominance of clastic processes over other processes is thus characteristic of shores on emergent islands, resulting in the "instant" production of large quantities of incoherent, mobile sediment. As a consequence, large accumulations of subaerially settled and/or water-settled tephra, together with reworked and resedimented hydroclastic debris, are deposited either as proximal "tuff cone aprons" or distally, in the deep submarine flanks of the edifice and surrounding flexural moat, forming an insular sedimentary apron (Menard, 1956; Schmincke et al., 1997; Schmidt and Schmincke, 2000; Schmincke, 2004). During periods of volcanic quiescence or in areas away from the active vents, epiclastic processes gain importance and are responsible for the generation of conglomeratic deposits generally composed of well-rounded sideromelane granules and small pebbles (Schmidt and Schmincke, 2002). Owing to the abundance of clastic material, coastlines generally consist of sand and pebble beaches, entirely composed of young volcanic mineral and lithic clasts. Eolian processes may also help to shape coasts - due to the abundance of fine mobile sediments - but are of modest importance.

4.1.2. *Breaching sea level by means other than summit volcanism*

Volcanic oceanic island edifices generally breach sea level through vertical growth by summit eruptions, implying a phase of surtseyan volcanic activity. However, the geological record of many eroded islands (e.g. Madeira Island in the Madeira Archipelago, La Palma in the Canaries, and Santiago in Cape Verde) does not exhibit evidence for an emergent surtseyan phase (Schmidt and Schmincke, 2000, 2002), but rather an erosive unconformity between the seamount series and the subaerial products (see Serralheiro, 1976; Klügel et al., 2005a; Brum da Silveira et al., 2010; Ramalho, 2011). This probably means that the transition between seamount and emersed island may take place through mechanisms other than summit volcanism. The presence of an erosive unconformity suggests a relative fall in sea level through either uplift, eustatic changes or a combination of both, between eruptive periods. When long term eruption rates are low, volcanism is very sporadic; if the recurrence time between eruptive periods is larger than 10^4 years, relative sea level may vary enough to allow the emergence (or maintenance) of the edifice above the sea surface be-

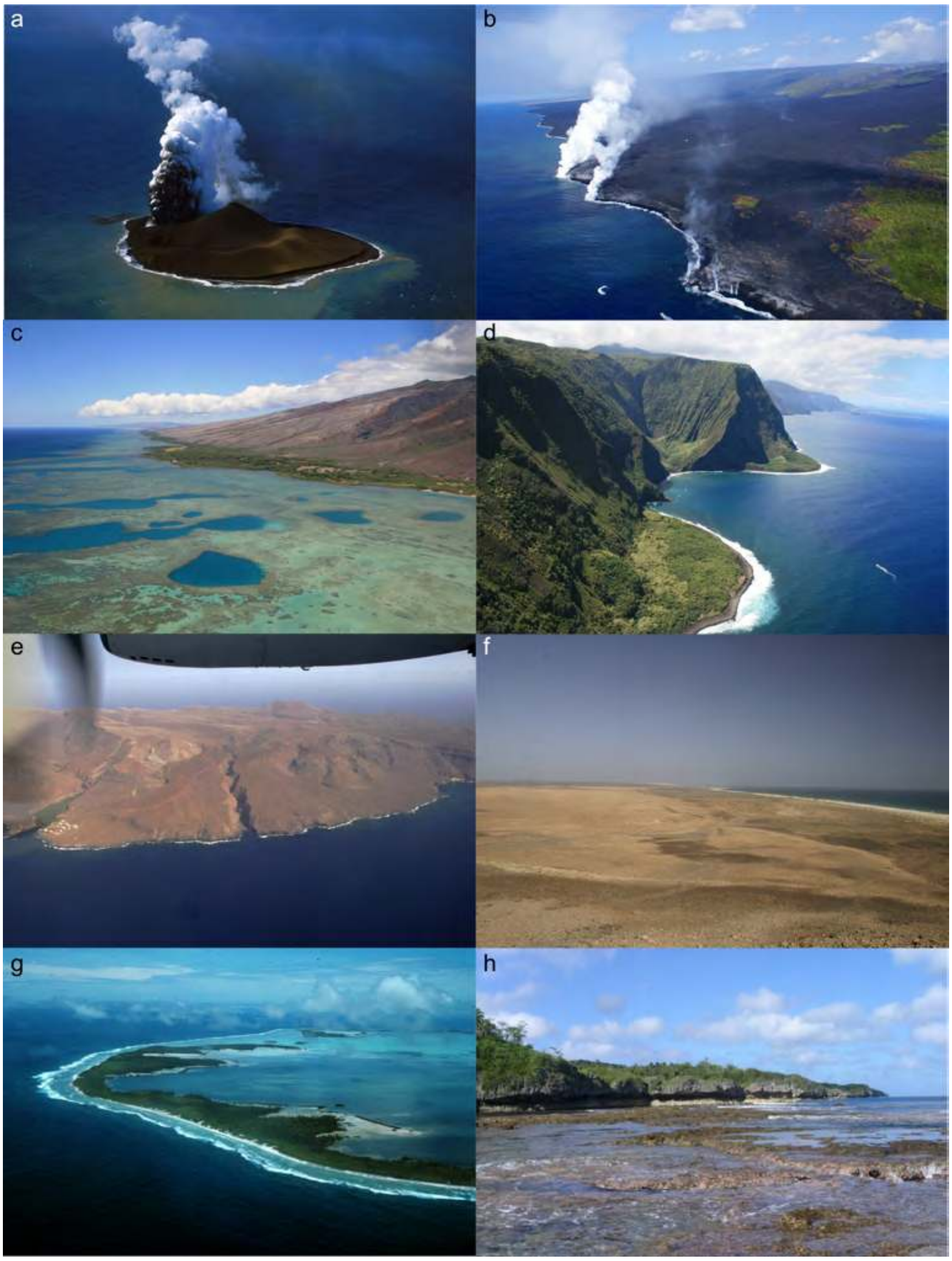

Figure 3: Examples of coasts at different stages of evolution. (a) Surtsey Island, Iceland, a precursory island built in 1963 through emergent eruptive activity (photo by Arctic-Images.com). (b) Shoreline expansion during the shield-building stage and through the progradation of lava-fed deltas, along the southeast coast of Kīlauea Volcano (Hawai'i) during April 2008 (USGS photo by J. Kauahikaua, courtesy of Hawaiian Volcano Observatory, USGS). (c) Southern coast of Moloka'i (Hawai'i) example of a leeward, poorly dissected and low shoreline of a mature island, protected by a well developed fringing reef. (d) Northern coast of Moloka'i (Hawai'i) example of windward high shoreline of a mature island, deeply dissected and created by a massive flank collapse (structurally controlled by a rift zone) and stronger marine erosion. (e) Example of a topography-filling, post-erosional coastal lava delta on the southeastern shore of São Nicolau (Cape Verde). (f) Example of an uplifted, terraced shoreline in the southern side of Boa Vista (Cape Verde). (g) Annular atoll, Cocos (Keeling) Islands. (h) Example of a coast of an uplifted atoll, exhibiting a large extant sediment-free shore platform and a limestone cliff with a well-developed wavecut notch, Niue Island (photo courtesy of D. Kennedy).

fore the next eruptive episode takes place. If this is the case, the subaerial shield-building lavas may cover the seamount products, unconformably; the products of a surtseyan phase were either removed through erosion or never existed in the first place and the edifice rose above the waves simply by a relative fall of sea level. Coastal processes acting during the emergence of such edifices are, consequently, different from those acting upon surtseyan edifices; the former are exclusively erosive and the coasts are rocky.

4.2. Shield-building subaerial stage

The onset of the shield-building subaerial stage is characterized by a dramatic change in the nature of volcanism from surtseyan/taalian to hawaiian/strombolian due to the complete isolation of the eruptive vent from sea-water influx (Thorarinsson, 1967; Kokelaar, 1983; Moore, 1985; Kokelaar, 1986). This change in eruptive style from highly-explosive hydromagmatic to largely effusive magmatic, allows the emission of lava flows that will mantle the emergent cone, forming a lava-flow cap, as it happened during the late stages of the eruption at Surtsey during 1963-67 (Thorarinsson, 1967; Kokelaar, 1983; Moore, 1985; Kokelaar, 1986; Schmidt and Schmincke, 2000; Ramalho, 2011). It is precisely the sustained growth of an erosion-resistant capping lava shield that allows an edifice to survive above sea level in the long term, and become a stable island. The successive extrusion of lava flows from summit vents - typically along rift zones - down the flanks of the volcanic edifice and towards the coast, gradually builds up to form the large shield morphology so typical of most young oceanic island volcanoes. The shield-building stage is generally characterized by high magma-supply rates - and consequently by high accumulation rates - and so edifices grow rapidly in height and size. During this stage, edifice lateral growth is characterized by rapid coastal progradation, sustained by the successive generation of coastal lava deltas as flows enter the sea along the fringes of the subaerial edifice (Moore and Schilling, 1973; Peterson, 1976; Skilling, 2002; Umino et al., 2006). Flank growth and the advancement of coastlines is further enhanced by the development of lava tubes, which allow lava to be carried long distances with little cooling (Peterson, 1976; Pinkerton and Wilson, 1994; Orton, 1996; Umino et al., 2006). Volcanic effusive processes dominate coastal processes during the vigorous subaerial shield-building stage; ongoing erosion and sedimentation occur alongside volcanic growth, but these processes play a minor role on coastal evolution during this stage.

Rates of coastal advancement by the formation of lava deltas during shield-building stages may be quite rapid over the short term. For instance, along the southern shore of Kīlauea during a single growth period of 3 months in 1990, lava-fed deltas extended the coastline seawards by 300 m, burying almost 407,000 m^2 of the 5-m deep Kaimū bay with nearly 3.95×10^6 m^3 of lavas at an averaged rate of 2,240-22,640 m^2/day (Umino et al., 2006). Along the same coastline, another delta was built between 1992 and 1994 which attained a length of 2.9 km and a width of 500 m (Mattox and Mangan, 1997). Likewise, during 1971 flows fed by lava tubes from Mauna Ulu created a 1.5 km-wide lava delta on an adjacent stretch of coast that extended the shoreline by 450 m (Moore and Schilling, 1973). Erosion, however, can also be quite fast in reducing newly-formed deltas. For example, the same 1971 lava delta mentioned above was reduced by 80 m and 100 m in length and width, respectively, by marine erosion in about one month (Moore and Schilling, 1973). Thus, at longer terms, coastal advancement rates are more modest, as subsidence, sea-level eustatic rises and marine erosion all contribute to slower volcanic growth. Effectively, at time scales of $10^3 - 10^6$ years, magma-supply rates must result in accumulation rates at coastlines that exceed subsidence and erosion rates in order to sustain coastal advancement (Lipman and Moore, 1996). For example, Mauna Loa must have experienced magma-supply rates in excess of 100×10^6 m^3/yr in order to sustain the coastal accumulation rates that were necessary to outpace a subsidence rate of 2.6 mm/yr (Lipman, 1995; Lipman and Moore, 1996). In a similar fashion, Mitchell et al. (2008) suggested that for significant prograding of the submarine flank of a volcanic island to occur, volcanic output must be high enough to infill the erosional shelf and overcome the shelf break onto the slope of the edifice, as happened on the southern flanks of Pico (Azores) and on Kīlauea.

4.2.1. Lateral growth through the progradation of coastal lava deltas

Shoreline progradation by successive generations of coastal lava deltas is the foremost process of lateral growth in erupting oceanic island shield volcanoes such as Hawai'i, La Reunión, and Isabela and Fernandina in the Galapágos (Peterson, 1976; Skilling, 2002; Umino et al., 2006) (see Figures 3b and 6a). The rate of formation of lava-fed deltas depends largely on effusion rates/volume flux of the lava flows entering the sea, and on the submarine bathymetry (Moore and Schilling, 1973; Hon et al., 1993; Mattox and Mangan, 1997). Likewise, the facies architecture of the resulting lava delta sequences reflects the influence of several, often interdependent factors (see Figs. 4 and 5) such as: coastal topography; offshore gradient, width and topography; water depth (and tidal variations); lava viscosity and effusion rates; location, number and spacing of feeding lava streams; and coastal orientation with respect to the dominant waves (Furnes and Sturt, 1976; Walker, 1992; Gregg and Fink, 1995; Schmincke et al., 1997; Gregg and Fornari, 1998; Kennish and Lutz, 1998; Gregg and Fink, 2000; Schmidt and Schmincke, 2002; Skilling, 2002; Gregg and Smith, 2003). As extrusion rates increase, the generation of pillows and hyaloclastites is gradually substituted by the generation of lobate lavas and megapillows, coalescent megapillows and leveed and unleveed submarine sheet flows (Griffiths and Fink, 1992; Gregg and Fink, 1995; Kennish and Lutz, 1998; Gregg and Fornari, 1998; Gregg and Fink, 2000; Gregg and Smith, 2003; Ramalho, 2011). An increase on the bottom slope may produce a similar effect as an increase in the effusion rate, resulting in more massive morphologies, but only to a threshold around $10^\circ - 15^\circ$; on steeper slopes, gravitational forces pull the lava lobes to a point at which their flow fronts disrupt and break apart into smaller 'fingers' of lava, thus favoring the generation of pillowed structures; on even steeper slopes ($> 25^\circ$) disruption at the flow front is such that angular, blocky rubble and hyaloclastites become the dominant products (Gregg and Fink, 2000; Gregg and Smith, 2003). Offshore slope or shelf width exert some influence on the formation of lava deltas in the sense that it controls accommodation space: gentler and wider shelves allow a more rapid lava delta progradation than steeper and deeper shelves because smaller volumes of erupted material are necessary to replace the existing water column (Lipman and Moore, 1996). Over longer time scales, other factors such as tectonics, uplift and subsidence, and glacio-eustatic variations also influence the architecture of lava deltas (Skilling, 2002, see Fig. 5).

The seaward progradation of lavas under moderate effusion rates (generally under 5-10 m^3/s; Walker et al., 1973; Rowland and Walker, 1990; Griffiths and Fink, 1992; Gregg and Fink, 2000) and over steep offshore slopes generates lava-fed delta structures similar to Gilbert-type river deltas (Jones and Nelson, 1970; Cas and Wright, 1987; Porebski and Gradzinski, 1990; Skilling, 2002; Schmidt and Schmincke, 2002) (see Figs. 4a and 5). Under such conditions, the entrance of subaerial lava flows (typically pāhoehoe) into the water body generates quenching and fragmentation that, due to gravity-driven processes, results in the accumulation of large quantities of coarse-grained, poorly sorted, volcaniclastic wedges on the progradational front of the advancing flows. The resulting structure is typically composed of three units (Porebski and Gradzinski, 1990): a) a basal unit composed of a mix of marine sediments and pebble breccias enveloped in a sandy hyaloclastite matrix; b) a foreset unit of flow-foot breccias, comprising foresets of seaward-thickening wedge-shaped bodies of hyaloclastites and pillow-breccias, often intercalated with thin tube-like pillow lavas; and c) a topset unit that caps the foresets and is composed of flat-lying subaerial (or more rarely submarine) flows. These sequences fill existing space (from the seabed to the water surface) by progradation, rather than aggradation. The seawards growth of such lava deltas is frequently intermittent and accompanied by lateral advances. As effusion rates wane, enhanced quenching and shattering of lavas upon entering the ocean may result in barriers of solidified lava and debris which divert later flows laterally along the coast, promoting the alongshore emplacement of primary flow lobes of the delta (Umino et al., 2006). On coastlines subjected to large accumulation rates and rapid sea-level rise, the successive superposition of Gilbert-type lava deltas leads to the generation of coastal effusive plains because deltas tend to enlarge landwards independently of seaward progradation (see Fig. 5); this happens because younger lavas, as they flow over the flat-lying top surfaces of previous deltas (partially flooded by sea-level rise) tend to grow upwards by intrusive inflation before fragmenting takes place at the edge of the new delta (Lipman and Moore, 1996). This is probably the mechanism under which the fringing effusive coastal plains so characteristic of very active shield volcanoes - like Kīlauea and the ones on the Galápagos - were formed, i.e. by a combination of high accumulation rates and rapid post-glacial sea-level rise. Lava deltas generated under these conditions tend to be more stable than deltas generated under stable sea level (Lipman and Moore, 1996). However, Gilbert-type lava-fed deltas are frequently very unstable and are prone to slumps that quickly change the shore outline and generate large quantities of volcaniclastic debris (Kauahikaua et al., 1993)

The entrance of lava flows (typically ‘a‘ā or pāhoehoe sheet flows) into water under high effusion rates (generally in excess of 5-10 m^3/s; Walker et al., 1973; Rowland and Walker, 1990; Griffiths and Fink, 1992; Gregg and Fink, 2000) and over offshore gently-dipping or flat-lying bottoms, in contrast, results in the accumulation/aggradation of submarine sheet flows that may branch out to form dendritic patterns but generally remain coherent subaqueously (Moore and Schilling, 1973; Umino et al., 2006; Mitchell et al., 2008; Ramalho, 2011; Stevenson et al., 2012). Thus, sequences formed under such conditions typically fill existing spaces more by aggradation rather than simple progradation, and are normally composed of an alternation of thick subhorizontal or gently-dipping submarine sheet flows and marine sediments, exhibiting little syn-eruptive hyaloclastitic material (see Fig. 4b). Such sequences are, generally, more erosion-resistant and help consolidate the island edifice; their formation, however, is enhanced by the existence of a previous ledge or shore platform where the pile may accumulate.

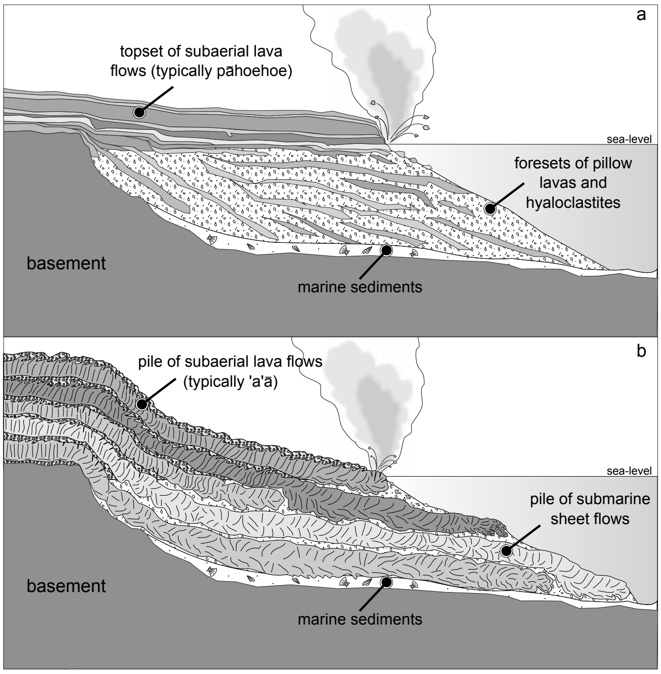

Figure 4: Formation of lava-fed deltas under different effusion rates. (a) Under low-to-moderate effusion rates (generally under 5-10 m^3/s), the entrance of subaerial lava flows (typically pāhoehoe) into the water body generates much quenching and shattering that, mostly by gravity-driven processes, result in structures similar to Gilbert-type deltas with prograding foresets of pillow lavas and hyaloclastites (proportion between pillow lavas and hyaloclastites typically increases with increasing flow rates) capped by a topset of flat-lying subaerial lava flows; available space is typically filled through progradation rather than aggradation. (b) Under moderate-to-high flow rates (generally in excess of 5-10 m^3/s), subaerial flows (typically ‘a‘ā) enter the water without much quenching and fragmentation and turn into submarine sheet flows by maintaining their coherence subaqueously; available space is typically filled through aggradation rather than progradation.

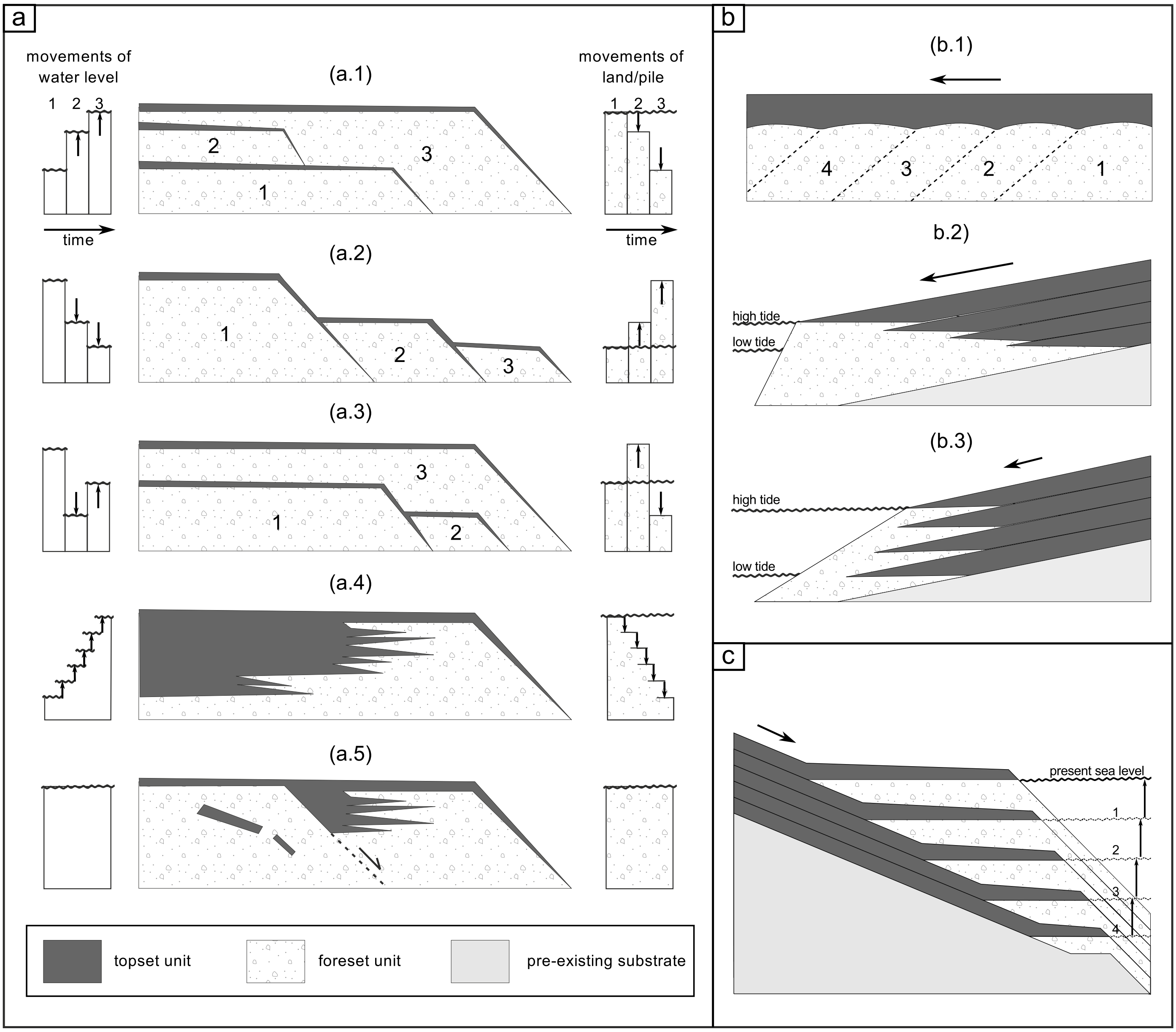

Figure 5: Structural relationships and passage zone morphologies of Gilbert-type lava-fed deltas during periods of vertical movement of water level or of volcanic pile, at different time scales. (a) General structural relationships of deltas that experienced vertical movements of water level or of volcanic pile (modified from Jones and Nelson, 1970). (a.1) Sequence of 3 overlapped lava deltas extruded during a period of rising water level or subsiding land. (a.2) Sequence of 3 onlapping lava deltas extruded during a period of falling water level or uplifting land. (a.3) Sequence of 3 lava deltas (exhibiting onlapping and overlapping) generated during a period of fluctuating water level or variable land movements. (a.4) A single lava delta formed during a period of extremely rapid water level rise or land subsidence. (a.5) Single lava delta affected by a syn-eruptive slump. (b) Morphology of passage zone for lava deltas subjected to tidal change (black arrows represent flow directions and effusion rates) (modified from Furnes and Fridleifsson, 1974; Furnes and Sturt, 1976). (b.1) Undulating passage zone on a flat-lying lava delta that was extruded continuously during several tidal cycles (numbered sequences each correspond to one tidal cycle). The amplitude of undulations mostly depends on tidal range whereas the wavelength mostly depends on effusion rates. (b.2) Geometry of the passage zone for a lava delta built by higher effusion rates and one cycle of rising tide; effusion rate is enough to cause progradation despite rising water level. (b.3) Geometry of the passage zone for a lava delta built by lower effusion rates and one cycle of rising tide; effusion rate is not enough to cause progradation and rising water level causes overlapping of the passage zone associated with each individual flow. (c) Overlapping of lava deltas under rapid relative sea-level rise (at centennial to millennial time scales) (modified from Lipman and Moore, 1996); successive lava deltas enlarge landward, whether or not they prograde seaward, forming coastal effusive plains.

4.2.2. The formation of littoral cones

Littoral cones are structures formed by violent steam explosions resulting from some subaerially erupted lava flows entering the sea (Moore and Ault, 1965). Observations by Mattox and Mangan (1997) on Hawai'i suggest that explosive interactions between molten lava and water to form littoral cones require high entrance fluxes of lava flows into water and are frequently (but not exclusively) initiated by collapse of a developing coastal lava delta. When pāhoehoe lava flows reach the ocean for the first time, the reaction may be relatively quiescent with lava dripping over older seacliffs or mantling established beaches. However, as lava deltas gradually develop and flow-foot breccias are formed by quenching and shattering of lavas on the progradation front of the deltas, the tube systems at the seaward edges of deltas are often established at or below sea level, increasing the chances for sudden water/magma mixing. If subsidence or collapse of the unstable front of the delta takes place, then favorable conditions for littoral hydrovolcanic explosions may be established. Explosive mixing of seawater and lava may occur under confined or unconfined (open) conditions according to the geometry of the mixing region (Mattox and Mangan, 1997). Unconfined mixing characteristically leads to the generation of tephra jets and steam rock blasts (see Figs. 6b and 6c), particularly, but not exclusively, when a complete delta collapse severs an active lava tube and waves interact with a streaming cascade of lava; resulting structures are typically small unconsolidated semi-circular littoral cones and fields of blocks built on the landward side of the delta (Mattox and Mangan, 1997) (e.g. on the southern shore of Hawai'i, see Figs. 6b and 6c). In contrast, confined mixing normally occurs when a partial collapse, subsidence or receding flow rates allow water to enter active lava tubes, causing steam expansion that generates lava bubble bursts and littoral lava fountains (continuous uprush explosions); resulting structures typically are low-profile spatter-dominated circular littoral cones and mounds, located at some distance inland from the delta front (Mattox and Mangan, 1997). Littoral lava fountains and bubble bursts may also occur when 'a'ā lavas reach the ocean (Moore and Ault, 1965). In this case the fragmentation is induced by the trapping and confinement of water/steam by the cooler (and more brittle) mantling 'a'ā flow (Moore and Ault, 1965).

The explosive interaction of seawater and molten lava along coastlines is also responsible for the production of small quantities of mobile volcaniclastic sediment. Furthermore, because of their location on the coast, littoral cones are very transient features and are rapidly eroded by waves, further contributing to sediment supply (Moore and Ault, 1965). Thus, the "instant production" of small pocket sandy beaches on the margins of advancing coastal lava deltas occurs alongside the effusive processes that so prominently create the low rocky coasts typical of young island shield volcanoes (e.g. Moore and Schilling, 1973). Notwithstanding their small expression, these sandy beaches may constitute the substratum and pathway for the biological colonization of the edifice.

4.2.3. Rift-edge explosive hydromagmatic volcanism

During the shield-building stage, vents are generally concentrated along well-defined to diffuse rift zones that typically develop in large oceanic island volcanoes. Thus, fissure-fed effusive subaerial eruptions constitute the prevailing extrusion style during this stage. This means that volcanic coastal morphologies are dominantly effusive structures (lava deltas) that originate when remotely-fed lava flows enter the sea, and that large hydromagmatic eruptions seldom occur. This is not always the case, however, in coastal areas where rift systems meet the sea (Németh and Cronin, 2009a). Here, due to the interaction between active rift zones and seawater, vigorous hydromagmatic volcanism may occur and form explosive morphologies and structures. The morphologies will vary from maars to tuff rings and to tuff cones, depending on the water/magma mixing mass ratio, i.e. depending on the relative position of the vent with respect to sea level; in onshore areas maars and tuff rings are more common while in offshore shallow-water areas tuff cones prevail. The development of syn-eruptive shorelines near rift systems may thus be characterized by explosive hydroclastic and associated epiclastic processes, acting alongside effusive processes, as suggested by the examples of Upolu on Western Samoa and Ambae Island on Vanuatu (the latter in an island-arc setting) (Németh and Cronin, 2009a,b).

4.3. Late stage coastal volcanism

Late stage or post-shield volcanism is generally characterized by a drastic reduction in eruption rates and by a change in magma composition (Macdonald et al., 1983; Clague and Dalrymple, 1987; Schmincke, 2004). In Hawaiian volcanoes (and other Pacific hotspots) this change is typically from very fluid tholeiitic basaltic magmas to more viscous alkali basaltic magmas, consequently changing the dominant eruption style from hawaiian to strombolian. This change in eruptive style is also frequently accompanied by a change in the distribution of volcanic centers, from fissure-fed rift zones to a more diffuse field of monogenic edifices, dotting the surface of the large shield volcano with numerous small cinder cones as on Mauna Kea (Hawai'i) and Isla Santa Cruz (Galápagos) (Macdonald et al., 1983; Clague and Dalrymple, 1989; White et al., 1993; Clague et al., 2000). If these eruptive vents are located near or at coastlines, surtseyan or taalian eruptions may take place, producing explosive hydromagmatic structures and products (e.g. on the western shore of Darwin Volcano, Isabela Island, Galápagos). In other settings, such as in many Atlantic hotspots (e.g. Cape Verde, the Canaries), the change in magma composition is often from alkali basaltic magmas to more evolved magmas such as phonolites and trachytes (Schmincke, 2004). This implies an even more dramatic change in the dominant eruption style from strombolian to plinian (or other highly explosive eruptive styles), enabling the generation of pyroclastic density currents that may reach coastal areas. On coastlines exposed to such eruptions, deposits of such density currents instantly cover existing deposits (beach, fringing reefs etc) and morphologies, frequently preserving their original architecture.

Figure 6: Examples of coastline expansion through volcanic activity. (a) Formation of an effusive lava-fed delta along the southeast coast of Kīlauea (Hawai'i) during March 2005 (USGS photo by R. Hoblitt). (b) Unconfined explosive interaction between an active severed lava tube (skylight visible on the foreground) and the ocean, forming a small crescent shape littoral cone (on the background) on the landward side, southeast coast of Kīlauea (Hawai'i) during July 2008 (USGS photo by Tim Orr). (c) Unconfined explosive interaction between active lava flows and the ocean, causing a tephra jet, southeast coast of Kīlauea (Hawai'i) during September 2006 (USGS photo by A. Doherty). (d) constructional volcanic processes (lava delta formation) occurring alongside destructive processes (quarrying of joint blocks by wave action) along the southeast coast of Kīlauea (Hawai'i) during August 2006 (USGS photo by C. Heliker).

4.4. Post-erosional stages

On most oceanic island volcanoes, an episode (or episodes) of rejuvenated volcanism may take place after a long quiescence period that follows the shield-building stage, and after the island edifices have been significantly eroded (Macdonald et al., 1983; Clague and Dalrymple, 1987; Carracedo, 1999; Schmincke, 2004; Ramalho, 2011). Rejuvenated volcanism typically involves small volumes of alkalic magmas erupted from a set of small monogenetic edifices (Schmincke, 2004). Because insular edifices at this stage are already deeply incised by marine and fluvial erosion and by large flank collapses, the geometry of the newly erupted structures and sequences is somewhat conditioned by the pre-existing topography. Thus, rejuvenated volcanism typically generates complex volcano-sedimentary, topography-filling sequences. Structures such as effusive valley-filling sequences, steep coastal lava fans/deltas, and cliff-edge effusive morphologies are common and contacts between rejuvenated products and the preceding units are typically irregular erosive unconformities or conformable transitions between overlying lavas and underlying sediments and biogenic structures such as coral reefs. As rejuvenated volcanic products are superimposed over mature and well developed coasts - with established shore platforms, beaches, cliffs and biogenic structures - resulting sequences frequently incorporate sharp transitions between subaerial and submarine volcanic products and marine/coastal sediments, reflecting rapid changes in shore morphology. The generation of complex coastal volcano-sedimentary sequences and morphologies is thus characteristic of edifices in a post-erosional stage.

4.4.1. Post-erosional coastal volcanism

Because rejuvenated volcanism normally erupts through a diffuse set of vents, explosive hydromagmatic volcanism may occur when vents are located along coastlines. Resulting structures will mostly vary from tuff cones to tuff rings and maars according to the water/magma mixing mass ratio. Volcanic processes acting upon coasts subjected to explosive hydrovolcanism will thus vary from proximal to distal hydroclastic, pyroclastic and effusive, working simultaneously with sedimentary processes. As vents may erupt through considerable sedimentary bodies or biogenic structures (e.g Diamond Head on O'ahu, Hawai'i), it is not uncommon to find sedimentary lithic fragments and even individual fossils mixed with fine juvenile tephra, as these were transported and deposited by explosive volcanic processes (e.g Hickman and Lipps, 1985).

4.4.2. The generation of post-erosional lava deltas

Inland rejuvenated volcanism may generate varying amounts of lava flows that typically move down the existing drainage network; lava deltas form when flows reach the sea. Post-erosional lava deltas are, typically, steep effusive fan-like morphologies that either form at the mouth of the valleys that served as channels for lava flows, or at the base of seacliffs when volcanism takes place at or close to cliff edges. Good examples of such lava deltas include Keāna Point on Maui (Hawai'i), Seixal on Madeira Island, Ponta da Ferraria on São Miguel (Azores), and those on the eastern and southeastern shore of São Nicolau (Cape Verde, see Fig. 3e). The architecture and morphology of such structures, however, may vary depending on extruded volumes, flow rates, channel steepness and width, and the height of the river mouth or cliff edge relative to sea level. The formation of littoral cones at the edges of post-erosional lava deltas is, however, rare, because the steepness of such structures generally prevents sea-water invading lava tubes to produce the necessary water/magma mixing ratio for explosive volcanism. A rare example of a littoral cone at a margin of a steep post-erosional lava delta can be seen at Ponta da Ferraria on São Miguel (Azores).

During post-erosional stages, shoreline morphology typically increases in complexity as focused formation of lava deltas create protruding headlands and contributes to more ragged coastal outlines. This, in turn, may create lesser energetic conditions for sediment deposition, leading to the generation of small gravel and sand beaches in sheltered bays and increasing diversity of shoreline facies. Sheltered bays may also constitute preferential areas for the accumulation of skeletal remains of marine life.

5. Coastal erosion

Oceanic island volcanoes are exposed to the destructive forces of the ocean from the moment they breach the sea surface until they (eventually) are drowned. The relative importance of marine erosion processes in shaping oceanic island coasts greatly depends upon the role and character of the other agents (e.g. volcanism, coral reef growth, etc). Marine erosion is the dominant agent (or at least an important one) of coastal evolution during: the emergent island stage of any island, especially when volcanic activity is dormant (see section 4.1); during all the post-shield stages on reefless islands; and eventually during the erosional and rejuvenated stages of islands with coral reefs. On the latter edifices, however, marine erosion is transferred seaward to where the protecting barrier reef is located, and it is only felt during storms.

5.1. Marine erosion and the development of shore platforms

The evolution and morphology of volcanically inactive oceanic island coastlines vary according to the structure, lithology, and mineralogy of the rock. The physical and chemical characteristics of the materials partly determine: the intensity and efficacy of the erosional processes; the amount, type, and mobility of loose material at the cliff foot, and consequently whether it plays a significant abrasive or protective role; surface irregularity in the intertidal and subtidal zones, which influences rates of wave attenuation; and because of the effect of rock resistance on erosional efficacy, the degree to which a coast retains vestiges of former sea levels and climates.

Frost, wetting and drying, salt and chemical weathering, and bioerosion can be dominant erosional mechanisms in some regions where there are suitable climates and weak wave activity but may be inhibited in places by the presence of coral reefs or sea ice (Guilcher et al., 1962; Guilcher and Bodere, 1975). Weathering also prepares rocks for eventual dislodgment and removal by waves in the more vigorous environments of the middle latitudes (Porter et al., 2010). Nevertheless, mechanical wave erosion usually dominates on volcanic islands with narrow shelves in exposed, oceanic regions (Quartau et al., 2010), and despite rapid attenuation, waves still play an important role in removing the products of weathering and mass movements where there are very wide shelves.

Mechanical wave erosion is accomplished by a number of processes, including the quarrying or dislodgment of joint blocks and other rock fragments by water impact (wave hammer), high shock pressures generated by breaking waves, and, probably most importantly, by air compression in joints and other rock crevices. As these processes depend upon the alternate presence of air and water they are most effective in a narrow zone extending from the wave crest to just below the still water level. The vast majority of coastlines on oceanic islands correspond to effusive sequences, to piles of shallow-dipping subaerial lava flows. These flows usually exhibit columnar and slab jointing that, together with the contacts between the flows, promote wave quarrying and the dislodgment of joint blocks (see Figs. 6d and 7a). Likewise, clinker and pyroclastic layers between flows are also easily eroded, facilitating quarrying processes. This explains how marine erosion can carve metric to decametric cliffs on young effusive structures in a matter of months or years, e.g. on the southeast coast of Kīlauea (Hawai'i) where lava deltas from 2010-2011 eruptions already exhibit a 3-4 m cliff. As a consequence of effective quarrying processes - that produce substantial amounts of very large blocks (Fig. 7a) - boulder accumulations at the base of cliffs are common (either at or below the intertidal zone), further contributing to marine abrasion. Abrasion occurs where wave-generated currents sweep rock fragments and sand back and forth or swirl them around within potholes. Although abrasion is not as closely associated with the water level as other wave erosional processes, its efficacy rapidly decreases with increasing depth.

Effective erosion of sea cliffs by waves and weathering (with removal of the weathered debris) produces shore platforms

Figure 7: Examples of marine and biological erosion features on oceanic islands. (a) Wave quarrying of recent lava flows (erupted during 1990) along the southeast coast of Kīlauea (Hawai'i) and resultant boulder beach (USGS photo by T. Orr, courtesy of Hawaiian Volcano Observatory, USGS). (b) Shore platform carved on pillow lavas, north coast of Fogo Island (Cape Verde). (c) Shore platform carved on subaerial tuffs, western coast of Sal Island (Cape Verde; photo courtesy of C. M. da Silva). (d) Cave, littoral arch and wavecut notch in plunging cliff on the southern coast of Santa Maria Island (Azores). (e) Modern and MIS5e (ca. 130 ka) wavecut notches on plunging cliff on the northern coast of Santa Maria Island (Azores). (f) Bioerosion (sea-urchin holes) on a shore platform (carved on tuffs) on the western coast of Sal Island (Cape Verde).

(previously known as wave-cut or abrasion platforms), horizontal to gently sloping rock surfaces extending between the high and low tidal levels (Trenhaile, 1987, 2011). The gradient of intertidal shore platforms - and consequently erosional rocky shelves - is determined by the tidal range, rock resistance, and the wave regime (Trenhaile, 1987, 2000, 2011). The absolute and relative efficacy of these process suites varies spatially and temporally, so that wave erosion or weathering can be dominant at a particular time or in a particular place on a platform surface. Shore platforms on oceanic islands are typically irregular when carved on subaerial flows, bulbous when carved in submarine volcanic units rich in pillow lavas (due to the rheological contrast between the pillows and the hyaloclastite matrix; Fig 7b) and smoother when carved in tuffs (Fig. 7c), hyaloclastitic deposits poor in pillows and pillow debris, fine terrigenous sediments, and carbonates.

There is usually a shore platform, boulder deposit, or beach at the foot of a cliff, but some cliffs plunge directly into deep water (Cotton, 1974). Plunging cliffs develop because of slow wave erosion, the small size of the islands, which prevents large amounts of material from being transported alongshore and deposited at the cliff foot, and relative sea level that rose much faster than sediment could accumulate. Plunging cliffs are particularly common around basaltic volcanic islands in the Southern Hemisphere, where these factors are often met. Despite approximately constant sea level, erosion is inhibited today by resistant rocks, the lack of abrasives at the water level, and the reflection of incoming, non-breaking waves. Nevertheless, hydraulic quarrying by the rise and fall of standing waves and the compression of air in rock clefts produce caves, notches (Figs. 7d and 7e), and narrow platforms which will eventually allow more erosive, breaking waves to attack the cliffs and destroy the plunging condition (Cotton, 1974; Trenhaile, 1987).

The depth to which waves are able to planate submarine surfaces has important implications for the development of shore platforms and insular shelves. It was once thought that erosion could occur at considerable depths, and this was reflected in the first conceptual models for shore platform and erosional shelf development, which assumed that very wide shelves could develop under constant sea-level conditions (Davis, 1896; Johnson, 1919; Challinor, 1949). Estimates of the maximum depth of erosion have diminished through time, however, from 183 m by Johnson (1919), 90 m by Barrell (1920), 46 to 92 m by Rode (1930), to 9 to 10 m by Dietz and Menard (1951). Menard and Ladd (1963) proposed that wave erosional surfaces are produced at sea level, although they can extend a little below that level under stable sea-level conditions. The corollary to modern acceptance of shallow water erosion is that wide erosional rocky shelves develop as intertidal zones migrate landwards with rising relative sea level, thereby maintaining water depths that allow sufficient wave energy to be expended at the cliff foot to undermine the slope and remove the resulting debris (Trenhaile, 1989, 2001). Thus, shore platforms become submarine terraces and island shelves after a relative sea-level rise, and subaerial terraces following a relative sea-level lowering. Erosional terraces, which must be distinguished from morphologically similar structures consisting of former beaches, coral reefs, and effusive coastal structures, can develop during Quaternary sea-level stillstands when the waves operate at essentially constant levels, but they are subsequently truncated or eliminated by erosion at lower elevations.

5.2. The development of insular shelves

Insular shelves develop from shore platforms that are produced within intertidal zones that migrate landwards and seawards with changing sea level (see Fig. 8). Therefore, an insular shelf corresponds to the low-lying submarine zones around island edifices extending from the coastline to the depth at which there is a marked increase in gradient to the submarine slopes of the volcanic edifice. In the absence of major vertical movements and on islands that have experienced at least one glacial/interglacial eustatic cycle, shelf depths should range from 0 to 130 m (Quartau et al., 2010). Earlier researchers (e.g. Menard, 1983) have suggested that the growth of insular shelves reflects the long-term competition between processes infilling the shelf (e.g. progradation of lava deltas) and those enlarging it (e.g. shoreline erosion). During early growth stages, large-scale landsliding can remove sections of islands and their associated shelves. Following the phase of active volcanism, shelves widen progressively, so that shelf widths increase with edifice age; the width increase, however, does not follow a linear relationship as the rate of wave erosion decreases with increasing shelf width and decreasing shelf gradient (Menard, 1983, 1986; Ablay and Hürlimann, 2000; Mitchell et al., 2003; Llanes et al., 2009; Quartau et al., 2010). Also, complicating this simple picture are the effects of sea-level oscillations, varied substrate resistance, sediment deposition, tectonics and subsidence or uplift of the island (Quartau et al., 2010).

Sea-level oscillations mostly affect the shelf profile. Theory and modeling suggest that changes in the amplitude of sea-level oscillations have little effect on shelf gradient - this depends mostly on tidal range, rock resistance, and wave regime - unless there are concomitant changes in the geological or morphogenic conditions (Trenhaile, 1989, 2001). However, shelf width and maximum depth increase with the amplitude of the eustatic oscillations. Thus, an increase in the amplitude of eustatic change results in adjustments to the shelf profile and lead to larger amounts of coastal retreat. Modeling also suggests that shelves become progressively wider and more gently sloping when subjected to eustatic sea-level cycles of approximately constant amplitude, resulting in increasing wave attenuation and slower shore erosion. These shelves trend towards a state of static equilibrium, which is attained when, because of low bottom gradients, wave stresses at or close to the water surface, at each point along their profiles, are less than the threshold or minimum (critical) stresses to erode the rock (Sunamura, 1978; Trenhaile, 1989, 2001). Erosion continues today at the shore not only because there has not been enough time for static equilibrium to be attained in hard rocks, variations in rock resistance, and changes in wave regime and other factors over time, but also because differences between sea-level oscillations require adjustments to the offshore profile. Erosion may also continue on coasts where weak waves remove fine-grained, weathered material in suspension, so that the effect of progressively gentler bottom slopes on wave attenuation rates is of little importance.

Wave attenuation rates, the distance that waves break from the shoreline, and the width of the turbulent surf zone increase as shelves become wider and more gently sloping. Assuming that tidal range is 3 m and that the minimum slope for an intertidal shore platform cut by waves is from 0.5 to 1°, the corresponding maximum width would be from about 170 to 340 m. Therefore, only very narrow islands could be reduced to shallow banks or guyots under constant sea-level conditions. The formation of shelves ranging up to 10 km or more in width, and in some cases the abrasion of entire islands is the result of Quaternary or earlier changes in sea level and subsidence of the island edifice. Nevertheless, even with these changes in relative sea level, the ability of waves to level entire volcanic islands depends upon such factors as island width and material strength, wave climate and tidal range, and the inherited morphology of the shelf.

5.3. The windward/leeward asymmetry

Oceanic islands are frequently exposed to dominant winds such as low latitude trade winds, the mid-latitude westerlies, or the polar easterlies. This exposure to a dominant wind typically imprints a hydrological, erosional and biological asymmetry to island edifices and their coastal regions. This is particularly dramatic at lower latitudes, where the relative exposure to the trade winds has a profound effect on precipitation.

Precipitation on oceanic islands generally follows a circular pattern with rainfall closely related to elevation, size and morphology of the edifice; rainfall is greater on larger islands with higher topography than on islands with similar areas but lower orography (Nullet and Mcgranaghan, 1988; Yang and Chen, 2008; Sobel et al., 2011). Small and low islands - especially those located in tropical and subtropical regions - do not generate enough diurnal cycling of elevated surface heating and mechanically forced upslope flow (from topographic barrier effects to trade-winds) to significantly enhance rainfall (e.g. Sobel et al. (2011) suggested the threshold of 315 km^2 for tropical islands to be able to generate significant enhanced rainfall). However, on tropical and subtropical high islands (e.g. the Hawaiian Islands), topography intersects the trade-winds enhancing total rainfall by a factor of up to 3-4 times when compared to oceanic rainfall values (Nullet and Mcgranaghan, 1988). The elevation threshold for orographic cloud bands and enhanced rainfall on oceanic islands is generally controlled by the altitude of the lifting condensation level and whether or not this level is below the trade wind inversion (Cao et al., 2007; Smith et al., 2009). This elevation threshold depends on many climatic factors and is regionally, locally and seasonably variable. Notwithstanding these variations, it is typically located around 600-800 m in elevation (Garrett, 1980; Giambelluca and Nullet, 1991; Barcelo and Coudray, 1996; Prada and da Silva, 2001; García-Santos et al., 2004; Cao et al., 2007; Smith et al., 2009). For islands with peaks above the trade wind inversion (frequently at 2000-2500 m in tropical and subtropical regions and usually lower at higher latitudes), maximum rainfall occurs on the windward slopes; conversely, for islands with mountaintops below the trade wind inversion (but above the lifting condensation level), maximum rainfall occurs on the mountaintops and rainfall is less asymmetrically distributed (Barcelo and Coudray, 1996; Yang and Chen, 2008). Consequently, the windward sides of high islands are more dissected, eroding 30 to 40 times faster than the leeward sides (Wentworth, 1927; Menard, 1986). Erosion by running water progressively dissect the volcanic terrain, giving rise to deep valleys, box-head ravines, and the sharply dissected topography that is characteristic of the Hāmākua coast in Hawaiʻi or the Nā Pali coast of Kauaʻi. The dissection leads to a process that Cotton (1969) has called skeletonisation and has a direct impact on the retreat of windward coastal regions of prominent islands.

Waves also reinforce erosion on the windward sides since stronger and/or more frequent winds increase wave height and frequency, resulting in higher erosion rates, as on Prince Edward island in the Indian Ocean, Madeira Island in the Atlantic, and Hawaiʻi in the Pacific (Menard, 1986; Mitchell et al., 2003; Brum da Silveira et al., 2010). Thus, large and high islands at latitudes subjected to trade-winds are frequently asymmetrical, with higher, more dissected coasts and wider shelves on the windward side (see Fig.3d). This is generally not the case, however, for very young and active volcanoes because either coastal topography is constantly renewed by volcanic activity (extending coastlines and infilling incipient valleys) or because in younger, more porous edifices rainwater tends to become groundwater (Menard, 1986).

In contrast with islands with enhanced windward fluvial erosion, on small and low islands waves are more effective erosional agents than streams because precipitation is low and rivers do not have space to develop. On these islands, shelf width and cliff height are generally related to the amount of time that coastal sectors have been exposed to wave action, and to the frequency and significant height of the waves, which may be different according to the windward/leeward exposure (Quartau et al., 2010).

5.4. Coastal erosion during extreme-wave events

Extreme-wave events such as large storms and tsunamis have the potential to cause great damage to coastal regions of oceanic islands, constituting major geomorphic crises (Paris et al., 2009). Storm surges and tsunamis have immense erosive power and are capable of quarrying and transporting enormous blocks from the shoreface (which adds to their abrasive capability) to the inshore. These megaclasts and boulders are typically derived from the surface and edges of adjacent reefs or rocky platforms (when these are present), or from the upper seaward edge of the adjacent seacliffs; the quarrying of the these megaclasts from both the shore platform edge and surface is facilitated by partial detachment along joints and cracks in the limestone or basaltic bedrock by marine erosion processes (Noormets et al., 2002, 2004; Richmond et al., 2011; Paris et al., 2011; Etienne et al., 2011). Storm- and tsunami-induced erosion, coupled with mass wasting, is, in fact, the main agent of change along volcanically-quiescent rocky shorelines of islands that are frequently impacted by extreme-wave events (e.g

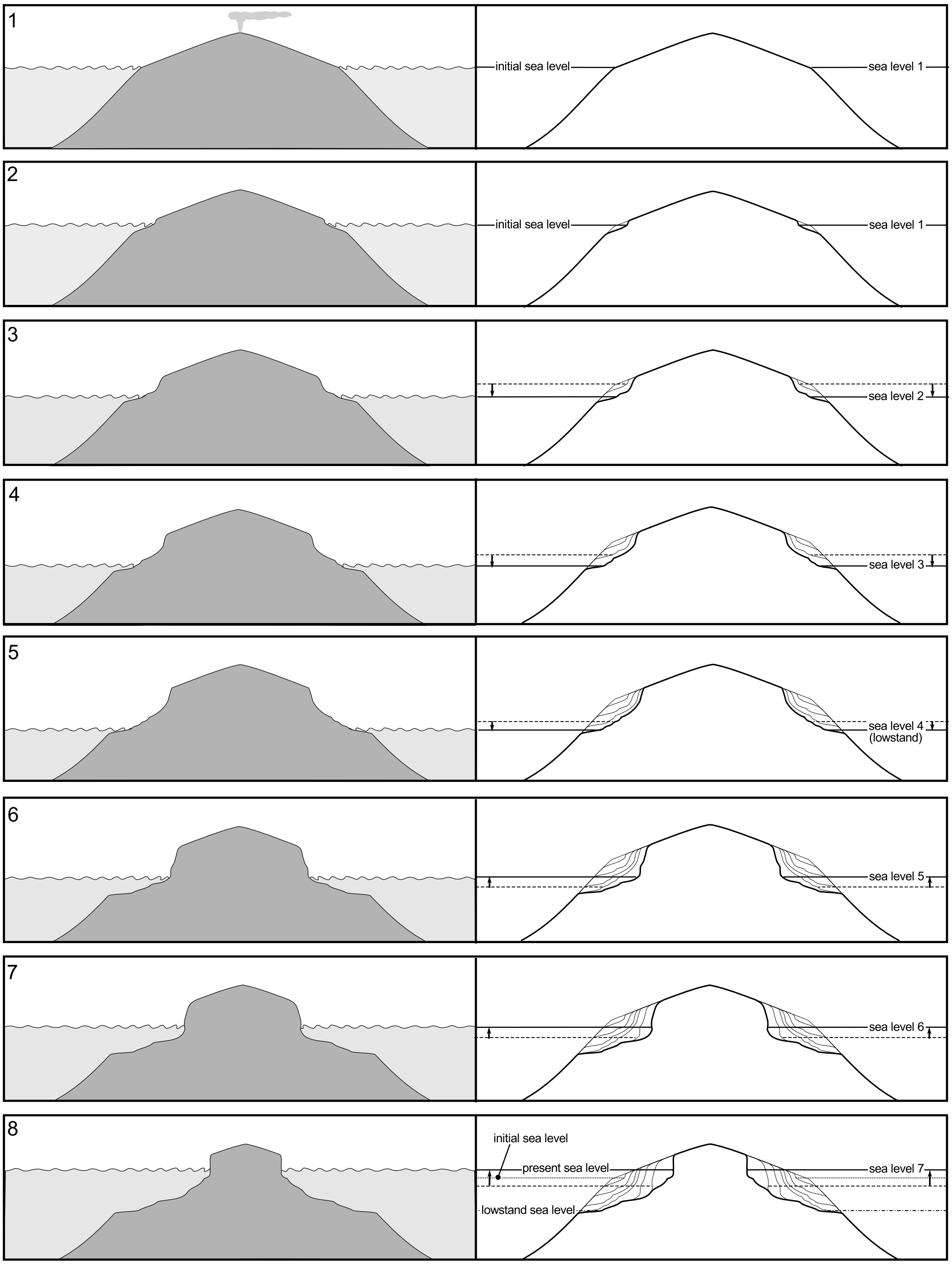

Figure 8: Schematic cross-section of a volcanic island showing the generation of a marine erosional shelf as a function of time during a single glacio-eustatic oscillation, and in the absence of uplift and subsidence (comparative profiles on the right-hand side). (1) Formation of a juvenile coast by volcanic processes. (2) Rapid incision of shore platforms, essentially within the intertidal zone and at the same mean sea level. To improve visibility at this scale, erosion appears exaggerated below sea level, although it occurs essentially within the intertidal zone. (3-5) Gradual sea-level fall, causing the seaward migration of shore platforms and onsetting the generation of an insular shelf; decelerating regression of seacliffs by backwearing and/or downwearing and essentially through processes of mass wasting and subaerial erosion. (5) Sea-level lowstand is reached during glacial maximum and seaward migration of the shore platforms ceases. (6-8) Gradual sea-level rise to present or interglacial level (a highstand), causing the landward migration of the shore platforms and increasing coastal recession; initial, previous, and lowstand sea levels also shown on (8) for comparison. Figures not to scale and topographical decay by subaerial processes is not represented.

the northern shore of Oʻahu). Shoreline evolution in such settings is, thus, mostly an event-driven process (Noormets et al., 2002). The erosive impact of storms and tsunamis is also very dramatic on clastic coasts, as mobile sediments are very easily transported by the high-energy waves and currents generated during these events. Along these coasts, extreme-wave events breach beach barriers, shift sand and infill coastal lagoons, and can even erode away entire beaches (since many of them are small and perched). This is particularly evident on shorelines of archipelagos with high storminess and exposure to local and far-field tsunamis like the Azores or Hawaiʻi. In the Azores, for example, the return period of storms capable of generating significant waves (Hs) of 12 m is just 5 years (Carvalho, 2003) and the islands have been impacted by tsunamis at least 23 times within the last 500 years (Andrade et al., 2006).

6. The role of biogenic processes

Living organisms established on coastlines may act as erosive agents (e.g. lithophagous organisms), constructors of large lithic structures that further enlarge and protect coasts from erosion (e.g. coral reefs, coralline algae), or sediment suppliers (e.g. organisms with mineral shells, such as mollusks and crustaceans); they may also simultaneously assume a combination of any of these functions (e.g. echinoderms) (Taylor and Wilson, 2003; Wilson, 2007; Wisshak et al., 2010; Davidson, 2011). Thus, it is not surprising that biogenic processes frequently play a significant role in island coastal evolution (e.g. in atolls). This is especially true for islands subjected to geographic or oceanographic conditions favorable to biogenic production, such as in tropical waters away from upwelling or freshwater and dust inputs. The relative importance of biological and biogenic processes in coastal evolution of remote oceanic island systems is, however, not only related to the prevailing environmental conditions, but also to the processes and patterns of dispersal and successful colonization, including arrivals by chance (Carlquist, 1966; Jokiel, 1990; Parker and Tunnicliffe, 1994; Scheltema et al., 1996; Ávila et al., 2009b), and to the assembly rules for ecological communities (Diamond, 1975; Connor and Simberloff, 1979).

6.1. Biological colonization

Fresh lava flows are easily and rapidly colonized even on such isolated oceanic islands as the Azores, Ascension Island or Tristan da Cunha in the Atlantic Ocean, and Easter Island in the Pacific Ocean (Ávila, 2006). The geographical location of the island is an important factor that will constrain the future biodiversity of the island. Usually, high-latitude or isolated islands will have fewer species over the long term when compared to low-latitude islands or islands located near source populations (Fridriksson and Magnússon, 1992). Also, the existence of favorable winds and ocean currents, as well as animal (e.g. birds) migration routes, are key factors that strongly influence the chances of dispersal. Birds play a key role on both seed dispersal and fertilization of soils by their excretions (Fridriksson, 1987). On the marine realm, algae propagules and planctotrophic larvae of several invertebrate phyla (e.g., gastropod mollusks, Balanidae and Chthamalidae crustaceans, sponges, echinoderms) are usually amongst the first settlers of these empty habitats (Carlquist, 1966), whereas on the terrestrial realm the pioneer colonizers are usually vascular plants on volcanic ash, and mosses and lichens on hard substrates (Brock, 1973; Fridriksson and Magnússon, 1992). On thermal habitats, located in the vicinities of fumaroles, blue-green algae are dominant (Brock, 1973). As these pioneering species settle, successfully reproduce, and spread along island coasts and inland, conditions may slowly arise for a viable colonization by other species that arrive later, and are typically dependent on the shelter and/or nutrients provided by the first settlers. One of the biggest problems these first colonizers have to cope with is the frequent shortage of water, as rainwater is rapidly lost by percolation through tuff, sand and juvenile lava onto deeper levels (Fridriksson and Magnússon, 1992). Gradually, with time, cumulative successful arrivals and colonization lead to more evolved, complex and diverse ecosystems (Diamond, 1975; Thornton, 1997; Fattorini, 2011). Thus, on remote islands, time plays a crucial role in the assembly of insular communities (Drake, 1990). Usually, in such isolated ecosystems, the probability of success of chance events of long-distance dispersal is quite low and the number of colonizers is usually small - the well-known "founder effect" (Mayr, 1954); as a consequence, a high variability is expected in species composition from archipelago to archipelago, and sometimes even between islands belonging to the same archipelago. Additionally, the profound isolation to which some of these populations are subjected may lead to genetic drift, speciation processes and the rise of endemisms by adaptive and non-adaptive radiation (Grant, 1981; Gittenberger, 1991; Vasconcelos et al., 2010), further contributing to this variability. The extreme isolation of many oceanic islands also subjects many populations to intense pressure, often on the verge of collapse, so when environmental conditions change rapidly many of these populations experience bottleneck effects, local disappearances, or even extinction (Carson, 1992; Ávila et al., 2008a,b). Even on less remote islands such as Surtsey, many pioneers did not survive, and 25 years after the emergence of the island, only 14% of the recorded species were successful colonizers (Fridriksson, 1989). Consequently, when compared to continents, most oceanic islands are usually characterized by a depauperate and disharmonic fauna and flora, in the sense that common species on continents are frequently absent in insular habitats (Whittaker et al., 1997; Ávila, 2006).

Biological colonization of the coasts of young oceanic island volcanoes is, thus, strongly influenced by the geometry and age of the archipelago (i.e. the history of emergence above sea level over space and time) (Fernández-Palacios et al., 2011), the patterns of ocean currents and winds that affect the newly-built edifices (Ávila et al., 2009a), the geographical location, and the distance to other well-established biological communities that may constitute the source of potential colonizers (MacArthur and Wilson, 1967; Fridriksson and Magnússon, 1992; Whittaker et al., 1997, 2008; Whittaker and Triantis, 2012). In linear, age-progressive island chains with small inter-island spacing -

like Hawai'i, Galápagos or the Society Islands - the process of biological colonization of younger volcanoes is typically facilitated by the existence of more solidly established communities in nearby older edifices - geologically "downstream" from the hotspot but biologically "upstream" - that serve as potential sources of colonizers. In these archipelagos, the linear array of island edifices serves as "stepping stones" for organisms to expand their geographic range, thus promoting the process of internal colonization (i.e. within the same archipelago) - a process that explains how some species are apparently older than the geological age estimates of the islands they inhabit (Rassmann, 1997; Sequeira et al., 2000). On smaller, more remote archipelagos in which islands are arranged in clusters instead of linear chains and where the history of edifice emergence above sea level is typically non-linear, the process of biological colonization of new edifices is generally more complex and somewhat more erratic. On single islands or small clusters of islands in extreme isolation - like the Azores or Easter Island and adjacent islets - the process of colonization is totally dependent on new arrivals from outside the archipelago and by resistant long-distance travelers (Scheltema et al., 1996), resulting in less diverse communities. After successful colonization of these islands, their extreme isolation usually promotes the genetic and morphological differentiation that ultimately leads to speciation (Rosenzweig, 1985; Carson, 1992).

6.2. Bioerosion

Bioerosion also begins almost immediately after the settling by algae that attack the rocky substrate by chemical dissolution (biocorrosion), by patellid gastropods that graze on biofilms, by sea-urchins (Fig. 7f) that scrape and burrow the basalts by mechanical means (bioabrasion), and by boring sponges, probably the most effective of the endolithic bioeroders (Neumann, 1966). Sponge bioerosion is a twofold mechanism: it is mediated by enzimes (carbonic anhydrase and acid phosphatase) that are responsible for mineral dissolution and digestion of organic components (Schönberg, 2008); and the chemical attack is aided by mechanical displacement of fragments of the substratum and their later transport out of the sponge galleries (Hatch, 1980). Sponge bioerosion rates and bioerosion sponge abundances are dependent on a large variety of environmental parameters, the most important are water flow, nutrient or sewage concentration and substrate density; salinity and temperature of the water, light conditions, water depth, and age and size of the sponge play a minor role (Schönberg, 2008). The boring of hard substrates such as basalts, by bivalve mollusks is also documented in the scientific literature for both the fossil record and the recent. For example, in the Atlantic islands, bivalve basalt borers are known from the Miocene of Porto Santo Island, Madeira archipelago (Santos et al., 2011a,b), from the Late Miocene of Santa Maria Island, Azores (Ávila, unpublished data) and from the Plio-Pleistocene of Santiago Island, Cape Verde (Santos et al., 2011b). Such species are also known from the geological record of volcanic islands on island-arc settings (e.g. Japan) as well as from their modern shorelines (Masuda, 1968; Masuda and Matsushima, 1969; McHuron, 1976; Fang and Shen, 1988; Haga et al., 2010), suggesting that bivalves capable of boring hard basalt substrates are quite common. Wisshak et al. (2010) showed that as bioerosion by photoautotrophic endoliths and grazers (that feed upon them) is a function of light, bioerosion rates are stronger in the photic zone and rapidly decrease towards deeper waters. In tropical latitudes, some littoral species of echinoids have impressive rates of bioerosion, as suggested by examples from analogous settings (e.g. the Caribbean). For example, *Echinometra lucunter* - a species that is common in the Bahamas and Bermuda but also on many Atlantic hotspot islands such as the Cape Verdes (see Fig. 7f) - is capable of eroding 6670 to 7000 $g/m^2/yr$ (Hunt, 1969; Hoskin et al., 1986). Another good example is *Paracentrotus lividus*, which is the most conspicuous bioeroding echinoderm in the Azores, Madeira, the Canaries and the Cape Verdes, being especially abundant at 1-2 m depth (Madeira et al., 2011). On these basaltic rocky shores, this echinoid actively uses its Aristotles lantern to bore cup-shaped to deep-pocket depressions, with a narrow entrance opening (Asgaard and Bromley, 2008).

6.3. Reef development and evolution

If a volcanic island occurs in seas that are suitable for coral growth then it is likely to rapidly develop a reef around it. Coral reefs cover more than 250,000 km^2 of the Earth's surface largely within the tropics where sea surface temperatures exceed 18 °C (Spalding and Grenfell, 1997). The greatest diversity of corals occurs in the Indo-Pacific with a second region centered on the western Atlantic; reefs are absent in the Mediterranean and reef development is limited in the eastern Atlantic and eastern Pacific Oceans. For example, corals are frequently stressed by El Niño-Southern Oscillation (ENSO) extremes in the Galápagos Islands where reefs are poorly developed, and reef development is also limited in the Marquesas Islands.

Mid-ocean reefs are typically classified into fringing reefs, barrier reefs and atolls. This distinction was first adopted by Charles Darwin (1842) who postulated that these represented successive stages in the development of reefs. Darwin realized that coral reefs could grow upwards fast enough to remain in shallow waters as the underlying basement subsides. His insight was to realize that these three types of reef represent stages in an evolutionary sequence, driven by gradual subsidence of the volcanic island around which the reef had initially formed. The Darwinian view involves prolonged subsidence of mid-ocean volcanoes and can now be integrated within the plate-tectonic framework to explain the relative location of different island types (Scott and Rotondo, 1983), particularly in the context of linear island chains in the Pacific Ocean. In such age-progressive island chains, the long-term subsidence trend that facilitates reef development is essentially driven by plate cooling and hotspot swell decay as the plate rapidly moves away from the hotspot (Morgan et al., 1995; Scott and Rotondo, 1983; Ramalho et al., 2010b). The Darwinian model is still generally accepted as a valid framework for reef evolution, and the classification of reef morphologies is still useful. However, at a finer scale, subsidence alone cannot explain the entire variety of modern and preserved reef morphologies present in the oceans (Toomey et al., 2013). These morphologies range from actively accreting fringing and barrier reefs to terraces preserved by drowning or subaerial exposure, and other effects need to be considered in order to fully explain their differences (Scott and Rotondo, 1983; Toomey et al., 2013). For instance, different subsidence rates may lead to different reef morphologies. Additionally, the effects of eustasy need to be integrated, because sea-level changes have had an important effect on the establishment, growth and renewal of reefs (Daly, 1934; Paulay and McEdward, 1990; Woodroffe et al., 1999; Woodroffe, 2008), and a sequence of interglacial reef limestones have been recognized with depth beneath atolls (McLean and Woodroffe, 1994). In a similar fashion, accretion rates may differ according to environmental and oceanographic conditions and will have a profound effect on reef morphology. Local uplift - whose origins are still poorly understood or are the focus of hot debate (see Scott and Rotondo, 1983; McNutt and Menard, 1978; McMurtry et al., 2010; Ramalho et al., 2010b) - may further contribute to the variability in reef morphology. It is now perceived that the diversity of modern reef morphology essentially arises from the combined effects of island subsidence (or sometimes uplift), coral accretion rates, and Pleistocene glacio-eustatic cycles (Woodroffe et al., 1999; Toomey et al., 2013). This modern perspective somewhat reconciles the competing models proposed by Daly and Darwin, in that reef profiles and the formation of barrier reefs are controlled by subsidence and vertical coral accretion (Darwin's hypothesis), yet the morphology of modern reefs bears the strong imprint of Pleistocene eustatic variations (Daly's hypothesis) (Toomey et al., 2013). At an even finer scale, wave energy may also contribute to shape reef morphology (Storlazzi et al., 2003).

The development of coral reefs around volcanic islands and over the timescales of glacio-eustatic cycles is essentially controlled by the vertical rate of reef accretion, which is the integrated result of coral growth, local production of detritus, and lithification (Dullo, 2005; Montaggioni, 2005; Toomey et al., 2013). Reef accretion rates typically decrease with increasing water depths, mostly because reduced light intensity limits coral growth (Bosscher and Schlager, 1992); as light intensity reduces with depth, it limits photosynthesis by the zooxanthellae algae that live in symbiosis with corals and on which they depend. Coral growth is thus dependent on water depth, with maximum growth values close to sea level, at the reef crest, and no growth below a critical depth. However, other environmental factors such as temperature, salinity, oxygenation levels etc, and, equally importantly, the availability of accommodation space, also control coral growth. Accommodation space (and consequently accretion rate) is, in its turn, typically maximized on particular conditions of relative sea-level rise, i.e. at particular rates of reef submergence. This means that, on longer geological timescales, accommodation space is normally controlled by tectonic subsidence of the substrate on which coral reefs grow. At smaller timescales, however, rapid glacio-eustatic oscillations - that typically happen at a much faster pace than tectonic vertical motions - have a strong impact on reef accretion (Paulay and McEdward, 1990). It is precisely the integration of these several factors controlling coral growth and reef accretion, at different timescales, that will shape reef development (see Fig. 9). At smaller timescales, and particularly during a postglacial transgressive period (as today), reef behavior is typically classified as "keep-up", "catch-up" and "give-up" (Neumann, 1985). Under slow to moderate rates of relative sea-level rise, and when accretion rates are able to match submergence, reefs can keep up with sea level, maintaining their crests close to this interface. However, during glacial terminations, sea-level rise happens at much faster rates (> 25 mm/yr) and can significantly outpace maximum reef accretion rates (typically < 10 mm/yr), submerging the reef (Neumann, 1985; Toomey et al., 2013). Then, when sea-level rise slows towards the peak of the interglacial highstand, two scenarios may happen: either the reef remained shallower than the drowning depth and eventually catches up with sea level, or the reef is submerged to depths beyond which accretion is not possible and is forced to "give up" (Neumann, 1985). Present-day reef morphologies are, of course, the result of reef development over longer geological timescales, i.e. they result from the integrated effects of multiple glacio-eustatic cycles, and long-term vertical motion. The impact of fast glacio-eustatic oscillations on reef development is, in general terms, to limit fringing reef growth, to make drowning more likely, and to generate structures that are typical of either sea-level lowstands or highstands (Toomey et al., 2013). With these insights in mind, it is now possible to refine the Darwinian model of atoll formation, with a higher level of detail than previously.

The first stage in atoll formation is the development of a fringing reef. Under favorable conditions, reefs are able to grow rapidly around the perimeter of young volcanic islands, until they completely or partially surround the volcanic edifice (see Figs. 3c and 3g). In some cases, reefs form on volcanic substrates that are still active; for example, coral establishment has been observed on recent lava flows around the Island of Hawai'i (Grigg et al., 1981). The evolution of a small fringing reef into the next stages of reef development, over timescales greater than a single glacio-eustatic cycle, will depend greatly on vertical motion rates (see Fig. 9 and Toomey et al., 2013). With low subsidence or slight uplift, accommodation space is very limited (more so because of sediment infill) thus constraining reef accretion; under these circumstances, reefs typically retain their fringing morphology (albeit some slow lateral growth or migration) or, if slight uplift occurs, become emergent single reefs or terraces. On the other hand, under slow-to-moderate subsidence rates (0.05 to 0.4 mm/yr), reef accretion is generally able to keep up with relative sea level; slowly, as island edifices subside, upward coral accretion maintains the reef near the sea surface and, gradually, fringing reefs convert into lagoon-bounding barrier reefs. This transition, however, bears the imprint of recent rapid sea-level rise through the (typically temporary) drowning of the reef and by stimulating accretion at the reef crest, either by a "keep-up" or "catch-up" process; thus, lagoon-bounding barrier reefs are typical highstand features (Woodroffe, 2008; Toomey et al., 2013). In the long term, however, as subsidence continues and reefs keep up with sea level, any trace of the volcano disappears below the sea sur-

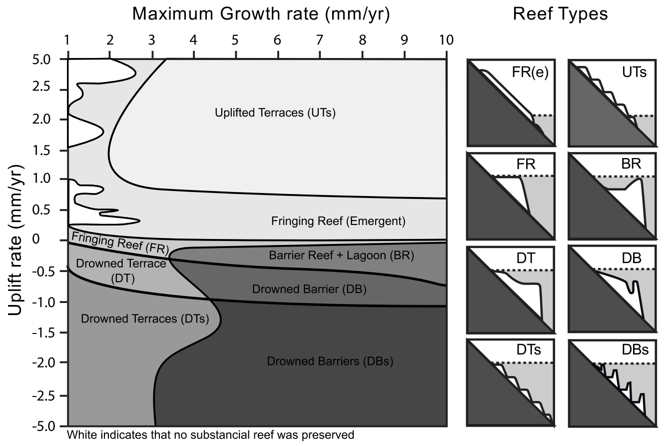

Figure 9: Model for reef development as a function of uplift rates and maximum reef growth, using a proxy eustatic curve. Modified from Toomey et al. (2013)

face, leaving the reef as an annular atoll (Darwin, 1842). Under more moderate subsidence rates (0.4 to 1 mm/yr), the reef may behave differently. Under these circumstances, reefs struggle to keep up with sea level, and a single drowned lagoon or terrace is produced during the postglacial sea-level rise; single drowned terraces, where the reef crest is still struggling to keep-up with sea level, are also typical highstand features. In extreme situations where islands are subjected to fast subsidence rates (in excess of 1 mm/yr), reefs do not have time to develop beyond fringing structures, and when they eventually submerge beyond the critical depth for coral accretion, they give up, and turn into drowned terraces; permanent drowning (give-up) generally occurs during lowstands and consequently multiple drowned terraces are typical lowstand features. The transition between atoll and guyot occurs when environmental conditions restrict, or when subsidence is fast enough to restrict, "keep-up" and "catch-up" of the reef crest, causing a definitive "give-up" of the reef system, at the so called Darwin Point (Grigg, 1982). In uplifting systems, by contrast, and under moderate-to-fast uplift rates (> 1 mm/yr), a staircase of wide raised terraces is generated when successive highstand reef structures get stranded subaerially.

The Society Islands demonstrate well the stages of reef development and of erosion of the volcanic core; in this chain it is possible to observe successively smaller subaerial remnants from Moorea along the chain to Huahine, Taha'a and Bora Bora culminating in the atoll of Tupai. The Hawaiian Island chain, including Hawai'i Island on which Kīlauea is active, exhibits the successive stages of dissection, particularly by fluvial processes, of volcanic islands of increasing age, as the chain moves to the northwest with migration of the Pacific plate; atolls are found to the northwest, and the chain continues with the submerged Emperor seamounts. The northern limit to coral growth sets a threshold beyond which this chain extends, which has been called the Darwin Point (Grigg, 1982). At the southern limit to reef development in the Pacific, the Lord Howe Island chain represents the opposite extreme, with Balls Pyramid, a spectacular erosional remnant 562 m high, beyond those seas in which reefs can form, but with an incipient fringing reef around Lord Howe Island itself, and atoll-like reefs to its north (Woodroffe et al., 2006).

The formation of a coral reef around a volcanic island has a profound effect on the relative rates of erosion of the island's shoreline (Menard, 1986). Cliff erosion, which is described elsewhere in this paper, is very effective at eroding volcanic islands, except where they are protected by a reef. The reef attenuates wave energy; swell breaks on the reef crest, and smaller secondary waves prevail in backreef environments (see Fig. 3c). Between 70 and 90% of wave energy is expended on the reef crest, ensuring that the shallow lagoon or reef flat behind it is comparatively sheltered (Hopley et al., 2007).

6.4. *The role of biogenic sediment production*

Living communities may contribute various amounts of biogenic sediment (typically of carbonate composition), depending mainly on local biological productivity levels. Biogenic sediments are mostly composed of the skeletons of calcareous organisms (such as corals and foraminifera), the broken-down products of abraded calcareous organisms (such as coral boulder and shingle, or shell hash) or the erosional products of bioeroders (such as the fine sediment excreted by parrotfish).

Reef growth is rapid on islands in reefal seas, particularly at the reef crest. The nature of the benthic habitats associated with the coral reef that surrounds such islands is a function of several environmental factors (Chappell, 1980). Behind the reef crest, which is often an intertidal rim veneered with coralline algae, backreef environments comprise either a shallow lagoon, which acts as a trap for sediments as at Lord Howe Island, or a near-horizontal reef flat that can be exposed by the lowest tides, the upper surface of which is often veneered with coralline algae or a thin sediment cover. Carbonate production is rapid; rates of vertical reef accretion, as distinct from coral growth, have been determined from coring and dating, and are often up to 8 mm/yr (Hopley et al., 2007). Higher maximum accretion rates, however, have been recorded at analogous settings, including 12 mm/yr on the Alacran Reef off the northern Yucatan Peninsula, 15 mm/yr on St. Croix, and 14.3 and 20.8 mm/yr on a fringing reef on the Pacific coast of Panama (Adey, 1975; Macintyre et al., 1977; Glynn, 1977). The smaller secondary waves and solitons formed behind the reef crest are directed landwards and deliver biogenic skeletal sediments, dominated by coral, foraminifera, coralline algae and mollusks, to the beaches that surround the island. On basaltic oceanic islands encircled by reefs, carbonate production typically exceeds the supply of sediment from the island itself and beaches are overwhelmingly carbonate in composition. For example, Lord Howe Island, at the southern limit of reef development, has a fringing reef along only 6 km of its perimeter, the remainder primarily comprising steep cliffs cut in basalt. However, corals grow at rates comparable to those in more tropical locations (Harriott, 1999), and the lagoon consists of reef-derived carbonate sediments several meters thick, with the beaches on its landward margin also > 95% carbonate (Kennedy and Woodroffe, 2000). Reef environments do undergo a range of erosional processes, through the action of borers and grazers; however, the net outcome is generally the accumulation of further carbonate, and through the aggregation of these processes, the reef gets larger and is cemented into the limestone structures that have been preserved around the tropics as reef limestone. Nevertheless, about 20 – 25% of the produced carbonate ends up being exported (e.g. transported offshore by storms) or lost to natural dissolution (Harney and Fletcher, 2003).

On reefless islands, where biogenic production is limited, the principal living organisms that contribute significantly to sediment production in terms of biogenic skeletal remains are mollusks and calcareous algae. These organisms are mostly sand-dwelling, and so their presence depends largely on the availability of suitable substrate. Truncation of volcanic edifices because of their exposure to wave abrasion in the absence of reefs, results in the formation of broad shelves around many such reefless islands. For example, Balls Pyramid lies near the centre of a planated shelf that is > 10 km in diameter. Lord Howe Island, which must have been reefless for much of the 6 million years since eruption until plate migration brought it into reefal seas, is surrounded by a shelf at least 8 km wide on all sides around the island, but dominated by sediments derived from algal material (Woodroffe et al., 2006). As reefless island coasts evolve and mature, insular shelves progressively widen and become veneered by increasing amounts of volcaniclastic sediments transported from shallower levels during storms. However, the continuous productivity of the calcifying organisms results in increasing contributions of skeletal remains over time. As a consequence, the biogenic content of sandy deposits in reefless islands increases as a function of time, as the edifices and coasts themselves evolve. This is especially true in uplifted old islands, where the possibility for sediment recycling is enhanced (e.g. in Santa Maria Island, Azores, or in Sal, Boa Vista and Maio Islands, Cape Verde, see Fig. 10h). The relative proportion of skeletal remains in sedimentary bodies of reefless islands may thus represent a qualitative way to determine the age of the edifice and its stage of development.

7. Sedimentation along coastlines and insular shelves

Oceanic islands are, by nature, exposed to wave action and thus sediment dynamics along their coasts and insular shelves are typically wave- to storm-dominated. The presence of reefs, however, as we previously argued, changes sediment dynamics considerably. When coral reefs are present, carbonate sedimentation increases significantly, either from erosion of the reef framework or from direct deposition as skeletal components on islands with fringing (Calhoun et al., 2002; Harney and Fletcher, 2003) or barrier reefs (Gischler, 2011), and on atolls (McLean and Woodroffe, 1994). Sandy beaches can thus rapidly form on embayed coasts protected by reefs, especially in places less exposed to wave energy (Bochicchio et al., 2009; Conger et al., 2009) (Fig. 10a). Hence, it becomes necessary to treat islands differently, according to the stage of development of their reefs.

7.1. *Reefless volcanic islands*

On reefless volcanic islands or on sections of an island that are not shielded by fringing reefs, sediment generation, transport and deposition along the coast typically vary according to the island's size and orientation relative to the trade-winds. On small, intermediate or low-lying reefless islands wave erosion is the dominant process contributing to shelf sedimentation with decreasing contributions from subaerial erosion, explosive volcanism, lava quenching when entering the sea and biological productivity (Quartau et al., 2012). This is the case on the Selvagens Islands, Corvo, Graciosa and Santa Maria in the Azores, and Prince Edward Island in the Indian Ocean. On these islands, the majority of coastal sediments derive from cliff erosion and/or from small, incipient reefs if present. Sediments are normally produced in small amounts, and on more energetic stretches of coast they are transported offshore by downwelling currents that develop during storms, stripping these coasts of sandy sediment. Conversely, large and prominent islands like the Hawaiian Islands, Madeira, Tenerife and Gran Canaria in the Canaries, Santo Antão and Santiago in the Cape Verdes, and La Reunión, tend to be eroded by fluvial processes and mass wasting rather than by waves (Wentworth, 1927; Menard, 1983, 1986; Salvany et al., 2012). On these islands, particularly on their windward sides, enhanced rainfall permits the development of large rivers that carry substantial amounts of sediments that are delivered to, and redistributed along, coastal areas (Fig. 10b) (Wentworth, 1927; Draut et al., 2009; Ferrier et al., 2013). In fact these rivers, which have very steep profiles, are extremely potent eroding agents, exhibiting some of the highest erosion rates on the planet despite their very limited catchment areas, lengths and seasonality (Louvat and Allègre,, 1997; Louvat and Allègre, 1998; Terry et al., 2006; Ferrier et al., 2013). For example, the Hanalei river basin on Kaua'i experienced a million-year averaged erosion rate of 545 ± 128 t/km^2/yr and discharged up to 690 t/km^2/yr (with a mean of 369 ± 114 t/km^2/yr for the period between 2003-2009) of suspended sediments to the adjacent coastline (Ferrier et al., 2013). Likewise, streams on São Miguel Island in the Azores erode about 184-525 t/km^2/yr and may discharge (individually) up to 495 ± 153 t/km^2/yr of suspended sediment load (Louvat and Allègre, 1998). Mechanical erosion rates at La Reunión are amongst the highest in the world, corresponding to 1200-9100 t/km^2/yr with peak erosion and discharge occurring during or immediately after cyclones, and amounting to a total denudation rate of 470-3430 mm/kyr (Louvat and Allègre,, 1997). It is not surprising, thus, that along the coasts of such prominent islands sediment contribution from subaerial erosion significantly increases relatively to that from other sources, a fact that further inhibits the development of coral reefs (Draut et al., 2009). On such coasts, sediments tend to accumulate nearshore in embayed gravel and sandy beach systems adjacent to river mouths (Fig. 10c). Along cliffed coasts, and away from riverine discharge, sediments can accumulate as narrow and steep boulder/gravel beaches on top of shore platforms mainly sourced from cliff erosion (Felton, 2002)(see Fig. 10d); on these shores, cliff mass wasting is a main contributor of very coarse sediments, and responsible for creating large boulder accumulations

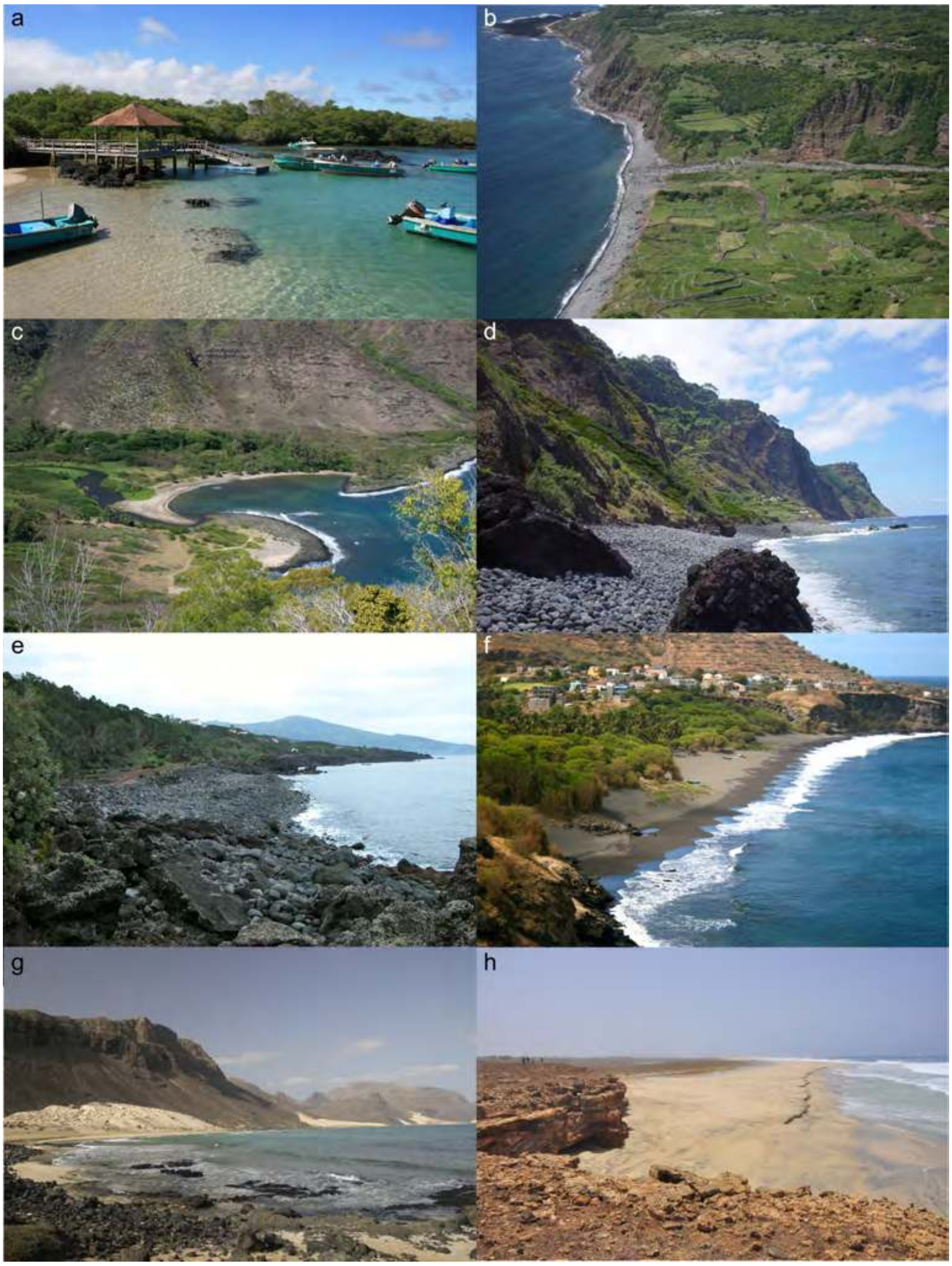

Figure 10: Examples of sediment accumulation along oceanic island coastlines. (a) Perched coral sand beach in an area of high biogenic productivity and low wave energy - a coastal lagoon created by young lava flows - along the virtually uneroded coast at Puerto Villamil on Isabella Island, Galápagos. (b) Large boulder beach in the vicinity of a river mouth on Flores Island (Azores). High (torrential) riverine discharge supplies large amounts of coarse sediment that is redistributed and reworked by marine erosion along the adjacent coastline. (c) Sand and gravel beach at the mouth of Hālawa valley, NE Moloka'i. (d) Linear boulder beach on the windward side of Madeira Island, derived from cliff erosion and failure, and wave action (photo courtesy of P. E. Fonseca). (e) Pocket perched boulder beach bounded by rocky promontories, on the leeward side of Pico Island (Azores). (f) Pocket sand beach at Ribeira da Prata, a protected bay on the leeward side of Santiago Island (Cape Verde); riverine solid discharge from valleys in the vicinity contribute fine sediment to this stretch of coast. (g) Bioclastic coastal shoulder sand dunes on the windward side of São Vicente Island (Cape Verde); local high biogenic productivity coupled with windy conditions promote the creation of bioclastic sandy beaches that are deflated, producing littoral sand dunes. (h) Large bioclastic sandy beach on the leeward side of Maio Island (Cape Verde) created by local high biogenic productivity and recycling of biogenic sediments from uplifted Pleistocene carbonate marine terraces (in the foreground); the high availability of fine bioclastic sediments, the leeward conditions, and a wide and shallow insular shelf allow the development of very large sandy beaches and supratidal salt pans (in the background) (photo courtesy of C. M. da Silva).

that temporarily extend coastlines seawards (see Fig. 11d-g). Additionally, on protected bays or other sites with locally high biogenic productivity, small bioclastic-rich beaches may form and even sustain coastal sand dune systems (e.g. Johnson et al., 2013) (see Fig. 10g).

Pocket beaches composed of boulders, gravel or sand may form between headlands or lava flows that confine and protect the deposits from currents and waves, allowing their stabilization (Figs. 10d and 10e). These are frequently perched, i.e. they are underlain or fronted seaward by more rigid structures such as reefs, lava flows or dykes exposed by differential erosion (Gallop et al., 2013). Sediment dynamics on exposed beaches is generally variable - with cells of erosion and accretion spaced within scales of hundreds of meters (Romine and Fletcher, 2013) - but is generally characterized by significant cross-shore movements of sand that reshape the beachface during large swell events such as storms; subsequent beach recovery begins rapidly following such events when weak to moderate fair-weather wave conditions favor onshore sediment transport (Dail et al., 2000). Along-shore currents are common along insular coasts and are mainly generated by the interaction between trade-winds and coastal morphology; tradewinds are capable of resuspending bottom sediment through the combined effect of wind-driven currents and wave-orbital velocities (Ogston et al., 2004; Calhoun et al., 2002). Alongshore currents further add to sediment transport, shifting fine sediment from bay to bay, particularly during more energetic seasons when much of the fine sediment may be transported in suspension (Norcross et al., 2002). The resulting sediment dynamics, with strong cross- and along-shore currents, means that sediments constantly shift along and across the shelf. In fact, in all these settings, sediments (especially fine ones like small gravel, sand or silt) tend to have a short residence time nearshore since high storm waves effectively remove coastal sediments and transport them offshore (Tsutsui et al., 1987; Quartau et al., 2012; Romine and Fletcher, 2013; Meireles et al., 2013). These sediments are typically deposited further offshore to form clinoform bodies on the deeper parts of the shelf (Chiocci and Romagnoli, 2004; Quartau et al., 2012). The accumulation of sediments within the island shelf is, however, transient, since most sediments end up being transported over the shelf break to the surrounding seafloor; sediment accumulations on the steep and narrow insular shelves are very prone to submarine mass wasting, producing turbidity currents that contribute to the formation of deep-sea volcaniclastic aprons (Menard, 1956, 1983; Schmincke et al., 1995; Schneider et al., 1998; Ávila et al., 2008a; Carey et al., 2011). The transport of sediments over the shelf break is generally episodic (seasonal or occasional), occurring mostly during major storms (Tsutsui et al., 1987; Saint-Ange et al., 2011) or as result of earthquakes, and are more frequent during sea level lowstands when island shelves are reduced in size (Ávila et al., 2008a; Quartau et al., 2012). Transport is greatly enhanced when submarine canyons cut through the shelf break (Saint-Ange et al., 2011; Sisavath et al., 2011; Romine and Fletcher, 2013). As result of offshore transport, volcaniclastic aprons accumulate in the periphery of island edifices and gradually infill the surrounding flexural moat induced by volcanic loading; they are composed of pyroclastic, hydroclastic and epiclastic sediments (mixed or interbedded with biogenic sediments) that often reflect the evolutionary stages of island building (Schmincke et al., 1995; Schneider et al., 1998; Carey et al., 2011). The transport of shelf sediments offshore, however, may be significantly limited on very wide shelves (Trenhaile, 2000, 2001), along small stretches of coast protected by reefs (Woodroffe, 2002; Harney and Fletcher, 2003; Storlazzi et al., 2004; Romine and Fletcher, 2013) or where specific coastal morphologies dissipate wave energy. In a similar fashion, on tropical, subtropical or even on temperate environments, sediment dynamics may be affected by the formation of beachrock. The cementation of beach sediments onto beachrock reduces the volume of available littoral mobile sediment, may make coasts more resilient to erosion, and change coastal morphology (Meyers, 1987; Cooper, 1991; Calvet et al., 2003). In the medium to long term, beachrock formation on stretches of coast that lack significant sediment input from riverine sources may induce a change from a sandy to a rocky shore, during multiple Pleistocene transgressions and regressions (Cooper, 1991)

7.2. Islands with fringing and barrier reefs

When a fringing reef forms around volcanic islands, it protects the shoreline, substantially reducing the rate of cliff recession. In addition to its protective role, the surrounding reefs also serve as sediment traps that retain the eroded volcanic material in the lagoon behind the reef, but the rate of terrestrial erosion and sediment delivery is almost always small in comparison to the prolific sediment production by calcareous organisms. Thus, large sandy beaches, dominantly composed of carbonate-rich sediments, are ubiquitous along shores protected by coral reef. As subsidence continues, the fringing reef becomes a larger barrier reef farther from the shore with a bigger and deeper lagoon inside. The lagoon fills in with eroded material from both the reef and the island, i.e., with both calcareous and lesser terrigenous sediment. Trade-wind-driven and tide-driven processes are the dominant control for circulation and sediment dispersal on the shallow, broad reef flats of fringing reefs; along-shore transport is mainly induced by the influence of trade-winds whereas across-shore transport is essentially controlled by tides and large swell (Ogston et al., 2004; Storlazzi et al., 2004; Presto et al., 2006). Sediment dynamics may alter significantly during storms. Sediments can be transported offshore to the top of the wave-eroded fore reef (Fletcher et al., 2008; Grossman et al., 2006) with an increase in the terrigenous component of the sediments due to high riverine discharge (Bothner et al., 2006).

7.3. Annular atolls

Ultimately, the volcanic edifice sinks below sea level but the barrier reef displaying an outer rim consisting of corals keeps growing and accreting producing a typical coral atoll (e.g. the Tuamotus, the Maldives, etc) . The outer reef encloses an open lagoon partially filled by fine-grained carbonate sediments that bury the volcanic basement. The production of carbonate on, or adjacent to, the reef crest is typically supplemented by localized accretion on patch reefs, and from other epibiotic organisms in the lagoon; this has been likened to the infill of a leaky bucket, which progressively fills (Purdy and Gischler, 2005). Finally, when subsidence outpaces coral accretion, the island becomes a guyot (Grigg, 1982; Thomas et al., 2011).

Atolls and almost-atolls (islands with only small volcanic outcrops remaining, like Aitutaki in the Southern Cooks or Bora Bora in the Society Islands) constitute circular to elliptical isolated carbonate platforms, reef-rimmed and aggraded to sea level (Guilcher, 1988; McLean and Woodroffe, 1994; Woodroffe et al., 1999). Margins are generally steep, with an outer escarpment, and exhibit facies belts, typically with a windward/leeward asymmetry (Mullins and Neumann, 1979; Guilcher, 1988). Inner lagoons are generally 30-90 m deep and are floored by carbonate muds and sands that are interrupted by high submerged reef knolls and patch reefs (McLean and Woodroffe, 1994; Woodroffe et al., 1999). Towards the rim, sediments typically consist of back-reef sand and cemented reef rubble, dotted by patch reefs and isolated coral heads, and sometimes forming small sandy islands that are made more resilient due to beachrock consolidation. The rim is constituted by reefal boundstones that extend a few tens of meters downslope on the seaward side until coral and algal sands can be found again, fining into skeletal sands and silts with loose reef blocks (McLean and Woodroffe, 1994; Woodroffe et al., 1999). Windward margins differ from leeward ones in the amount of loose sediment available; windward margins, due to higher wave energy, typically have less sediment (Guilcher, 1988; Woodroffe et al., 1999). On atolls, windward rim islands are generally sandy with shingle ridges, and are referred to as 'motu', a Polynesian term for island (Newell, 1961; Woodroffe et al., 1999; Woodroffe, 2008); conversely, reef islands composed entirely of sand are more common on the leeward rim and are known as 'cays' (Stoddart and Steers, 1977; Woodroffe et al., 1999; Woodroffe, 2008). Sediment dynamics are also essentially driven by trade-wind related processes, tides and storms.

7.4. Uplifted atolls and razed islands

Edifices such as uplifting coral atolls or uplifting razed islands typically experience a change in coastal morphology. Uplifting atolls tend to change from a sand and reef barrier facies to one more typical of rocky coasts, with near vertical limestone cliffs and wide, sediment-free shore platforms, as on Makatea, Rurutu and Niue in the Pacific, or Aldabra in the western Indian Ocean (see Fig. 3h) (Montaggioni et al., 1985; Menard, 1986; Stoddart and Spencer, 1987; Stoddart et al., 1990). Erosion of those limestones often occurs through mechanical and biological processes which lead to the development of a distinct intertidal notch (Trudgill, 1976), or a sequence of notches at different heights which implies little overall retreat of the limestone between successive sea-level stands. In contrast, uplifting razed islands such as Sal, Boa Vista and Maio in the Cape Verdes, may experience a more subtle change that typically involves an increase in sediment accumulation along the coast, either from recycling of previous marine terraces or simply from the exposure of flat shore platforms that favor sedimentation (Fig. 10h); these platforms, in their turn, facilitate the generation of large coastal alluvial fan-deltas and of aeolian coastal dune systems by the deflation of sandy beaches (see Fig. 10g).

7.5. Sediment fluxes during extreme-wave events

On oceanic islands, extreme-wave events are the main agents of rapid across-shore and across-shelf mass sediment transport (both offshore- and inshore-directed) and provide the only means for significant inshore transport of coarse sediments to supratidal zones and inland areas. On rocky coasts, overwash from tsunami and storm surges typically result in supratidal deposits that include solitary coarse clasts, pockets and clusters of coarse clasts, thin sand sheets in topographic depressions, and incipient development of low ridges (Noormets et al., 2002, 2004; Richmond et al., 2011). Deposits from large storm and/or swell waves are generally more confined to areas closer to the shoreline and typically correspond to prominent and multi-generational shore-parallel ridges of sediment and other positive relief features that mask the underlying topography and are regularly modified (Richmond et al., 2011). In contrast, tsunamis typically generate boulder-strewn gravel fields with megaclasts and thin sediment accumulations with blankets of sand and gravel in topographic lows, and frequently extend several hundreds of meters inland and even upslope. They result in a mixing of offshore, coastal and subaerial sediments, in a range of different sizes. Storms and tsunamis are also responsible for inland inundation of river valleys and overwash of littoral regions, dumping marine sediments on fluvial systems, coastal lagoons and coastal flats (e.g. Chagué-Goff et al., 2012). Well-studied examples of insular rocky coastlines impacted by storms and tsunamis include the Kohala and Kīlauea shores in Hawai'i, the northern shore of O'ahu, the Agaëte valley in Gran Canaria, and at Tarrafal of Santiago in the Cape Verdes (Noormets et al., 2002; McMurtry et al., 2004; Felton et al., 2006; Pérez-Torrado et al., 2006; Richmond et al., 2011; Paris et al., 2011; Chagué-Goff et al., 2012).

Low islands such as atolls and almost-atolls are particularly vulnerable to storms and tsunamis. On these islands, extreme-wave events can be both constructional and erosional agents that can affect considerably shoreline morphodynamics (Bayliss-Smith, 1988; Woodroffe, 2008). According to Bayliss-Smith (1988), extreme-wave events are the only marine processes capable of emplacing large-volume ridges of coral rubble, while more frequent but less intense fair-weather waves are important for reworking storm deposits into stable features. Storms and tsunamis both affect sediment transport but storms also affect diagenesis as a result of the enormous volume of freshwater carried and discharged along their paths; the processes of lithification-solution (karstification) acting upon the carbonate rim are directly influenced by storms and sea-level fluctuations (Bourrouilh-Le, 1998). Effectively, hurricanes, typhoons and tropical cyclones frequently play a very important role on the modulation of the surface morphology of the atoll rim and the supply of sediment to the lagoon (Bourrouilh-Le, 1984; Bourrouilh-Le et al., 1985; Woodroffe, 2008). Hurricane-stricken atolls may experience several changes such as: the build-up of seaward ridges and banks (made of coral rubble) that may add to the accretion of shorelines; a general drop of beach tops due to sand removal; the appearance of new 'hoa' (storm-induced passage of marine waters over the rim) resulting in washover fans behind breached barriers; and the migration of sandy islands (Bourrouilh-Le, 1998; Woodroffe, 2008). On atolls exposed to hurricanes, 'motus' may even experience storm-induced lagoonward movements of up to 30 m in a single season, or up to 100 m in about 25 cyclonic periods, as in the case of Mataïva atoll (Bourrouilh-Le, 1998). Atolls exposed to storms, like the Tuamotus, consequently exhibit more abundant conglomeratic platforms and seaward ridges; conversely, those atolls that are relatively free of storms, such as the Maldives or the Gilbert chain in Kiribati, comprise reef islands largely built from sand (Woodroffe, 2008). Tsunamis have similar effects on coral atolls, with both erosional scarps and overwash deposits concentrated on the tsunami-exposed side of the islands and accretionary structures such as spit and cuspate foreland extensions on the tsunami-leeward side (Kench et al., 2008).

8. Edifice slope failure (landsliding)

A distinction is made here between massive failure and lava delta collapse, both of which typically occur during the shield-building stage, and seacliff failures occurring wherever surf erosion generates unstable cliffs.

8.1. Massive failure

Systematic surveying using side-scan and multibeam sonars around the submarine slopes of volcanic oceanic islands has allowed the identification of widespread deposits of giant landslides and confirmed that many deep embayments in the topography of such islands are actually landslide headwalls (Holcomb and Searle, 1991; Moore et al., 1989) (see Fig.3d). Before this discovery, embayments could still be argued to be the result of caldera-collapse, other volcano-tectonics or perhaps even extreme coastal erosion, depending on the physiography. The Nuuanu landslide complex north of Oahu is one of the largest deposits with a total volume of around 2000 km^3 and includes the 1.8-km high Tuscaloosa Seamount, which is a giant slide block (Moore, 1964; Moore et al., 1989; Lipman et al., 1988). The discovery of these features has spawned debate over the origins of oceanic volcano flank collapse, which has been well summarized by Keating and McGuire (2000). The large variety of proposed causes of failure summarized in their review arises because we have very poor knowledge of the conditions at the time of failure, hence there is considerable uncertainty. Dominating the factors leading to instability is the volcanic constructional process itself, which produces steep slopes prone to failure. Those steep slopes are then acted on by a variety of processes that can push the slope to failure, including ground accelerations and shaking associated with earthquakes (either plate-tectonic if near a plate boundary or volcano-tectonic), elevated pore-fluid pressures associated with rainfall and groundwater heating by intrusions, hydrothermal alteration leading to weakening of rock bodies and stresses caused by dense intrusions. An example of a magnitude 7.1 earthquake possibly involved in movement of the Hilina slump system of Hawai'i occurred in 1975 (Crosson and Endo, 1982). An aseismic movement of the Hilinia Slump system has been recorded with continuous GPS measurements and linked with a major rainfall event (Cervelli et al., 2002) so pore pressures are clearly important. Also suggesting links with pore pressure, Quidelleur et al. (2008) put forward evidence for incidences of landslides occurring during glacial-induced eustatic sea-level change and Mitchell (2001, 2003) outlined a transition in the morphology of edifices at > 2000 m height that may have a pore-pressure origin amongst other possibilites. Concerning links to intrusions, Clague and Denlinger (1994) suggested that the high density and weak rheology of hot olivine cumulates may contribute to the movement of the Hilina slump. In the Canary Islands, Hurlimann et al. (1999, 2000) suggested that failure can be promoted by weak palaeosoil horizons buried beneath later lavas and loss of support because of strong coastal erosion at the base of terrestrial slopes and deeply entrenched canyons that subsequently form the sidewalls of landslide valleys. Failure of slopes could also be linked to caldera collapse (Martí et al., 1997).

The immediate effect of massive failure is commonly to produce a major indentation of the coastline associated with deep landslide headwalls and chutes. Mass-movements almost certainly generate tsunamis affecting adjacent coasts of an archipelago, as reviewed by McMurtry et al. (2004) for the Hawaiian islands. Although predicting the exact impact of tsunamis with modeling depends on accurate reconstruction of the mobile volume and its displacement history (Satake et al., 2002), there are clear examples of marine deposits now on land interpretable as tsunami deposits, showing that, as described earlier in this work, tsunamis are capable of moving significant quantities of sedimentary material from the shelf and coast of islands. Tsunamis are shallow-water type waves so they are refracted by bathymetry and may therefore affect other coasts in ways that may not seem immediately obvious without modeling of wave propagation.

The effect of the entrenchment of the edifice is well illustrated, for example, in the morphology of guyots in the Pacific Ocean (Vogt and Smoot, 1984). Modern examples include those on Madeira Island (Brum da Silveira et al., 2010), El Hierro in the Canaries (Masson, 1996), Guadalupe off Mexico, and Isabela in the Galápagos (Mitchell, 2003). However, failure may not always generate embayments; the westerly flank of La Palma shows more of a convex-seawards coastline and a debris cone on the submarine slope of the island (Urgeles et al., 1999) and the slump system of Kīlauea (Hilina) has been continually overlain by new lavas (e.g. Smith et al., 1999). The amount of material moved during these events can appear to be quite large (exceeding 10^3 km^3 in some instances (Lipman et al., 1988), but they are usually minor compared with the volume of the edifice. Their isostatic effects are therefore also usually minor; Smith and Wessel (2000) estimated by modeling that even the effect of unloading the island of Oahu by the 1200-5000 km^3 of the Nuuanu slide probably caused uplift of 10-109 m (the extreme value perhaps sufficient to expose much of the island shelf).

Post-failure lavas commonly fill the embayment, leaving

Figure 11: Examples of coastal mass wasting on oceanic islands. (a) Multi-stage rotational (slump) massive flank collapse of Fajãzinha - Fajã Grande, on Flores Island (Azores)(yellow line represents approximate maximum headwall). (b) and (c) Lava delta before and after the slump of 28 November 2005, Lae'apuki (Hawai'i) (USGS photos by R. Hoblitt and T. Orr, courtesy of Hawaiian Volcano Observatory, USGS). (d) Landslide (debris avalanche) of 30th October 2012 on the NW coast of Corvo Island (Azores); this debris avalanche occurred during a storm and created an islet some distance from the coastline, made of boulders, gravel and sand (photo was taken a few days after the collapse, photo courtesy of F. Cardigos, SIARAM). (e) Detail of the previous landslide deposit (photo courtesy of F. Cardigos, SIARAM); note the quantity of fine sediment in suspension in the water; about 20 days after the event, current and wave dynamics remobilized sand and gravel and created a sand bar connecting the islet to land. (f) Fajã dos Cúberes (on the foreground) and Fajã de Santo Cristo (on the background) on São Jorge Island (Azores), two large debris avalanche deposits that created coastal lagoons; their origin is probably associated with slips in the fault that controls this portion of the island's coastline (photo courtesy of F. Cardigos, SIARAM). (g) Ponta da Fajã, on Flores Island (Azores), a large debris avalanche deposit that has been mostly eroded by wave action.

lower-relief sea cliffs than elsewhere along the coasts. This is observed on parts of Madeira Island (Brum da Silveira et al., 2010), the Canary Islands (Masson et al., 2002), the Kīlauea coast of Hawai'i (Umino et al., 2006) and south-east Pico island of the Azores (Mitchell et al., 2012). Until strongly modified by coastal erosion, construction of lava deltas typically produces a coastline that is convex-seawards in plan-view. The combination of both the failure and subsequent building out by lavas usually leaves a reduced or even missing shelf over those segments.

In contrast to large, deep-seated island flank collapses, smaller (but still large) collapses with shallower detachment surfaces may have a different immediate effect on coastal morphology. In these cases, because detachment surfaces are shallower, the mass-movement can itself extend the coastline by the seaward mass transfer caused by collapsed debris or rotated slump blocks. This type of collapse typically creates low-lying, seaward-convex swaths of land with irregular or hummocky topography, with a steep inland concave headwall - the collapse scar. A good example of such feature is Ponta Delgada on Madeira Island, a peninsula that still extends > 1 km into the sea (Brum da Silveira et al., 2010). These protrusions, however, are generally composed of incoherent collapsed material, which may be easily eroded so that shorelines rapidly recede back to their previous position, as in the case of Fajãzinha - Fajã Grande collapse on Flores Island, Azores (see Fig. 11a). With continued erosion, and when collapsed material has been completely cleared away, shorelines may recede to the base of the collapse headwall leaving large embayments with shallow water depths. These, although similar in morphology (above sea level) to large embayments created by deep-seated massive collapses, differ in the sense that they typically exhibit a large, shallow adjacent marine shelf.

8.2. *Lava delta collapse*

Large lava-fed deltas, with a structure resembling Gilbert-type river deltas (Jones, 1966), involve the building out of lava flows over flow-foot hyaloclastitic and pillow-lava breccias. Loading by the advancing lava flows on these poorly consolidated slopes makes them prone to collapse (see Figs. 11b and 11c) (Kauahikaua et al., 1993; Mattox and Mangan, 1997). The collapses involve small rotational or translational slumps, which generate benches (when a partial collapse/subsidence occurs) and instantaneous cliffs. As they are formed in recently erupted sequences, they erode rapidly in surf and produce further volcaniclastic sediment.

8.3. *Cliff failure*

Wave erosional processes operate on the lower part of sea-cliffs whereas subaerial mechanisms, including weathering and mass movement, dominate higher up, as in rocky cliffs in other environments. Areas of dominantly large significant wave height tend to produce steep or undercut cliffs, whereas strong weathering produces more gentle, convex slopes. Coastal cliff slopes are therefore generally steeper in vigorous, storm wave environments than in other environments where frost and chemical weathering are more important. Cliff undercutting by waves, solution and other weathering, or bioerosion, promotes various types of mass movement which reflect the structure, lithology, and other characteristics of the rock, and the absolute and relative efficacy of subaerial and marine processes (Trenhaile, 1987; Sunamura, 1992; Hampton et al., 2004). Basal erosion and hydrostatic pressures exerted by water entering rock clefts can generate a variety of essentially surficial failures on cliffs in well fractured rocks, including rock and slab falls, sags, and topples. Other falls can develop through pressure release and the formation of tension cracks parallel to the erosion surface following cliff erosion and retreat. Deep-seated movements are larger but less frequent than falls. They include translational slides in seaward-dipping rocks (e.g steeply-dipping piles of lava flows), alternations of permeable and impermeable strata (e.g. alternation between lava flows and weathered tuffs), massive rocks overlying incompetent materials (e.g. massive flows above palaeosoils), and argillaceous and other easily sheared rocks with low bearing strength (e.g. weathered tuffs, palagonitic tuffs, etc). Rotational slides, or slumps, usually occur in thick, fairly homogeneous deposits of tuffs, palagonitic tuffs, hydrothermally altered volcanic units, and other weak and weathered/altered substrates. Thick effusive sequences of prominent and mature shield volcanoes (in the erosional or post-erosional stage), due to their compaction, groundwater alteration or weathering, may be subjected to large slumps that somewhat resemble the massive flank collapses typical of the shield-building stage; these rotational slides may involve large volumes and be responsible for reshaping extensive swaths of coastline, e.g. on the northern coast of Madeira Island (Brum da Silveira et al., 2010).

Recession of cliffed coastlines essentially takes place through episodes/cycles of marine under-cutting, cliff failure, erosion and removal of collapsed material; this is particularly evident on the windward side of many islands or on locally very energetic stretches of coast (compare Figs. 11f and 11g). This process is thus recurrent and is characterized by successive local shoreline advances (when collapse takes place and collapsed material accumulates at the base of cliffs) and retreats (when marine erosion removes collapsed material) resulting, in the long term, in a net retreat of the coastline. Coastal retreat is consequently the net result of a continuous horizontal erosive component (marine erosion) and an episodic vertical erosive component (mass wasting).

Coastal cliff failures (rockfalls and topples and, more rarely, slumps) are responsible for creating (often temporarily) large accumulations of collapsed material (boulders) at the base of cliffs, typically with low profiles and flat or hummocky surfaces and occasionally forming small islets or coastal lagoons (see Figs. 11d-f). Examples of such features are well-known in the Azores, Madeira and Cape Verde archipelagos, where they are locally called coastal (detritic) "fajãs" (Figs. 11d-g). In these archipelagos, such morphologies are very common because of the highly energetic wave regime and the lack of reef protection that characterize their coasts.

9. Structural and tectonic control on coastline evolution

On many oceanic island volcanoes, coastal morphology is partly controlled by active or inherited tectonic features and/or volcano-tectonic structures. These result from either regional tectonic stresses or local volcano-tectonic and/or gravitational stresses affecting the edifices (Fiske and Jackson, 1972; Dieterich, 1988; Walker, 1993). Structural control on coastal morphology is especially evident on island edifices that are located at or near plate boundaries, i.e. island edifices built by fissure volcanism along major tectonic features such as leaky transforms or spreading ridges (e.g. Azores, Iceland). On these edifices, coastlines may directly correspond to active fault scarps, and their evolution is frequently influenced by tectonic activity (e.g. earthquake-triggered mass wasting). Examples of coasts that are structurally controlled by active faults associated with a regional stress field include São Jorge in the Azores (see Fig. 11f) (Walker, 1993; Madeira and Brum da Silveira, 2003; Hildenbrand et al., 2008; Silva et al., 2012); the geometry of this island edifice is tectonically controlled (through fissure volcanism), most of its northern cliff corresponding to a fault scarp(s) and many of the large cliff failure events that occurred in historical times were probably earthquake triggered (Madeira and Brum da Silveira, 2003). In mid-plate island systems, the structural control on coastline morphology is probably less evident, but is also common. On these islands, structures are typically controlled by volcano-tectonic stresses coupled with gravitational stresses (Walker, 1999; Carracedo, 1999; Klügel et al., 2005b; Walter et al., 2006). Many oceanic island volcanoes exhibit rift zones defined by swarms of dykes, and these may influence coastal morphology in a direct or indirect way. Dyke wedging along rift zones may, together with gravitational stresses, trigger giant landslides that reshape coastlines (Carracedo, 1999). Likewise, inactive rift zones also constitute structural discontinuities that influence differential erosion and mass wasting, leading to linear stretches of coast composed of high cliffs, e.g. on Madeira and Desertas (Schwarz et al., 2005; Klügel et al., 2009; Brum da Silveira et al., 2010), and Moloka'i (Hawai'i) (Fig. 3d).

10. Conclusions

Oceanic island volcanoes are very dynamic landscape systems and they constitute prime localities to look at how different processes and agents of change interact in a complex ways to shape coastal evolution. Volcanic island coasts are also one of the few places where constructional processes can be observed over short time scales, frequently occurring side by side with destructive processes. Coastal processes typically operate more rapidly on islands than on their continental counterparts. Additionally, oceanic islands constitute a confined environment where boundary conditions are more easily understood, making them excellent natural laboratories to study ongoing coastal processes. The islands' rapid emergence above sea level, as well as their extreme isolation, make insular shores ideal locations to investigate the mechanisms and patterns of biological dispersion and colonization of virgin environments and the evolution of young, pioneering ecosystems. In a similar fashion, the morphological evolution of islands with increasing age (especially in age-progressive, linear island chains) also makes insular edifices ideal places to look at how coastlines and coastal processes respond to external factors such as uplift and subsidence, eustasy, mass wasting, biological activity, climate change etc. Additionnally, the diversity in types of shore morphology - rocky (basaltic and limestone), clastic (either young volcaniclastic or erosional and bioclastic), or reefal - in a relatively small area (sometimes even within the same island edifice) offers excellent opportunities for academically and logistically easier comparisons than coasts on continents. Finally, it should be emphasized that the present work focuses on oceanic island volcanoes, i.e. hotspot-related volcanic islands; nevertheless, many of the processes here described may operate in a very similar fashion at other types of oceanic islands, like those on subduction-related volcanic island arc settings. This is particularly true for island edifices that are overwhelmingly basaltic in nature, as on Jejudo in the Korea Strait or Ambae Island in Vanuatu Archipelago. Many aspects of the present review are relevant to other oceanic island settings and even continental coastlines of volcanic nature.

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