3

Salt Marsh Formation

ANTONIO B. RODRIGUEZ AND BRENT A. MCKEE

3.1 Introduction

Historical records show a massive decline in salt marsh area (Pendleton et al. 2012), $> 50\%$ in many locations, such as sites in Australia (Saintilan and Williams, 2000; Rogers et al. 2006), the British Isles (Baily and Pearson, 2007), and New England, USA (Bertness et al. 2002). These losses are mainly fueled by an underappreciation of the large contributions of salt marsh to maintaining healthy and productive estuaries. Prior to the middle twentieth century, the value of salt marsh primarily depended on its potential for reclamation. Davis (1910) proclaimed that "...[salt marshes] are conspicuous, being generally unutilized for any purpose except for making a small amount of inferior hay, hence they are practically desert places, except where land values are sufficiently high to make it worth while to raise the surface above high tide level for building purposes, or to dike out the tides." We now view salt marsh as a valuable estuarine habitat because it provides coastal protection from waves (Shepard et al. 2011), erosion control (Neumeier and Ciavola, 2004), water purification (Sousa et al. 2008), fish and bird habitat (Peterson and Turner, 1994; Van Eerden et al. 2005), carbon sequestration (Mcleod et al. 2011), and tourism/recreation (Barbier et al. 2011; Altieri et al. 2012). Salt marsh is also a coastal depositional environment that can accrete vertically over millennial time scales at rates equal to, or greater than, sea-level rise (Gehrels et al. 1996; Ouyang and Lee, 2014). The relatively high accretion rates and resistance of salt marshes to erosion (Mudd et al. 2010) make them valuable sites for preserving records of sea level (van de Plassche et al. 1998; Engelhart et al. 2011; Kemp et al. 2017), storms (Donnelly et al. 2001; Boldt et al. 2010; de Groot et al. 2011), and tsunamis (Morton et al. 2007; Komatsubara et al. 2008) in their sediments. Salt marsh loss and associated services have been pervasive globally, mainly due to the direct (grazing, ditching, pollution, etc.) and indirect (climate change) effects of human activities, resulting in the recent emphasis on restoration, conservation, and management (Lotze et al. 2006; Airoldi and Beck, 2007; Gedan et al. 2009). Although recent focus has been on better understanding of those mechanisms and processes that are related to salt marsh degradation, reviewing salt marsh formation and the different modes of salt marsh expansion will aid efforts aimed at preserving and increasing salt marsh habitat area and extracting climate and tectonic information from their sedimentary records.

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3.2 Conditions Conducive to Salt Marsh Formation and Expansion

The successful colonization of substrate by salt marsh species and subsequent growth depends on interacting abiotic and biotic factors (Redfield, 1965; Engels et al. 2011). Salt marsh formation, persistence, and expansion requires relatively low wave and current energy and the intertidal-flat pioneer zone and salt marsh must maintain an intertidal elevation with sedimentation $\geq$ erosion (Dijkema, 1997). Successful establishment of salt marsh vegetation on tidal flats increases with increasing tidal-flat area, which also dampens wave power (Bouma et al. 2016). The seasonal variability of the tidal-flat-bed elevation due to varying rates of sedimentation and erosion also must be low to allow roots to take hold and limit burial of seedlings (Bouma et al. 2016). On established salt marshes, the submergence frequency and duration, wave- and tidal-current energy, and salinity decreases with increasing elevation and distance from the shoreline. These physiochemical gradients create distinct salt marsh zones from the shoreline to the upland boundary (Odum, 1988; Engels and Jensen, 2009).

Salt marsh biota contain both marine and terrestrial characteristics making them stress tolerant of salinities between 12 and 35 ppt (Odum, 1988; Flowers and Colmer, 2008). Although growth and survival are greatest in freshwater (Adams, 1963; Phleger, 1971), salt marsh species are excluded from tidal freshwater areas due to interspecific competition but thrive in higher salinities where physical stress is too great for freshwater marsh plants to survive (Snow and Vince, 1984, Crain et al. 2004; Engels and Jensen, 2009; Engles et al. 2011). Salt marsh vegetation traps sediments by reducing turbulence, which enhances particle settling and lessens erosion (Stumpf, 1983; Neumeier and Ciavola; 2004; Neumeier and Amos, 2006; Mudd et al. 2010). This reduced physical disturbance causes the vegetation to grow better (Bruno, 2000; Van der Wal et al. 2008) and increases the efficiency of the salt marsh sediment trap. Salt marsh formation is thus controlled by both geomorphological and ecological feedbacks (Fagherazzi et al. 2012; Kirwan and Megonigal, 2013).

Broadly, salt marsh formation and evolution strongly depend on relative changes in sediment accommodation and sediment accumulation. Accommodation is defined by Jervey (1988) as the amount of space that is available for sediments to accumulate in. Accommodation is always below base level (the level of erosion) and at the coast is closely tied to sea level. Near the shoreline, in the shallow water parts of estuaries, base level is also influenced by tidal currents and locally generated wind-wave power. At yearly and shorter time scales, storms are the principle driver of changes in salt marsh sediment accommodation because they erode shorelines and shallow substrates (Leonardi et al. 2015). Nearshore bathymetry also controls wave power near estuarine shorelines, with conditions being more conducive to deposition (low wave power) as the fetch and depth of adjacent tidal flats decreases (Mariotti and Fagerazzi, 2013).

Relative sea level (RSL), defined as changes in the volume of water in the ocean plus local tectonic changes (e.g., uplift and subsidence), is the principle driver of salt marsh sediment accommodation at decadal and longer time scales. Rising RSL creates sediment accommodation along the upland boundary of the salt marsh promoting landward salt marsh expansion. In addition, rising RSL increases sediment accommodation across the extant salt marsh, and depending on the rate of rise, tidal range, and suspended sediment concentration, could either enhance vertical accretion or, if the rate of RSL rise is too high, it could result in salt marsh loss through excessive soil waterlogging, adversely impacting soil chemistry (increase in salinity, decrease in oxygen), plant growth, and maintenance of an intertidal elevation (Reed, 1995; Morris et al. 2002; Kirwan et al. 2010). Conversion of salt marsh to subtidal flat is less likely in areas with a high tidal range and high suspended sediment supply (Kirwan et al. 2010). Accumulating marsh sediment is mainly composed of organic material from plant growth and mineral matter that settles onto the marsh from the water column and through ice rafting in middle to high latitudes. Sediment availability in the estuary is important for marsh formation, and increasing suspended sediment concentrations promotes subtidal flat accretion and its transformation to a marsh platform (Marani et al. 2010; Gunnell et al. 2013; Bouma et al. 2016)

3.3 Salt Marsh Classification

Salt marshes are associated with a variety of different estuarine landscapes and are commonly placed into groups based on geomorphology (Kelley et al. 1988; Allen, 2000), continental geography (Adam, 1990), and/or vegetation assemblage (Chapman, 1960). While these groupings are useful for characterizing the spatial distribution and associated genera of salt marsh, they are limited in their application to understanding shifts in salt marsh area and location under those constantly varying conditions associated with global change, such as warming, sea-level rise, and coastal development. Here, we simplify salt marsh classification, basing different types on their mode of formation and morphogenetic relationships between salt marsh and adjacent environments. We classify salt marsh into three general end-member groups, including: (1) Fringing, (2) Patch and (3) Deltaic (Fig. 3.1). Differences among salt marsh classes are mainly defined by their associated ecotones, ontogeny, and the processes that build and/or convert substrate to intertidal elevations. Fringing salt marsh initially forms on upland substrate and is located between the mainland and the open-water parts of estuaries (Fig. 3.1A). Patch salt marsh initially forms on intertidal flats of flood-tidal deltas or washover deposits in wave-dominated estuaries and on tidal sand ridges in tide-dominated estuaries (Fig. 3.1B). Patch salt marsh can also form on intertidal oyster reefs, but that succession of environments is relatively understudied. Deltaic salt marsh is found at the heads of estuaries and along shoreline protuberances where rivers discharge directly into basins. Deltaic salt marsh is commonly associated with the lower delta plain, including interdistributary bays and distributary-channel mouth bars (Fig. 3.1C). The physiochemical conditions that promote growth of salt marsh vegetation can be widespread, making these classifications not inclusive of all settings where salt marsh vegetation grows, such as ditches, inter-ridge swales, and coastal ponds.

3.4 Fringing Salt Marsh

Fringing salt marsh forms adjacent to the upland, typically along the protected shorelines of drowned river valleys and tidal creeks (Fig. 3.1A). This type of salt marsh initially forms as

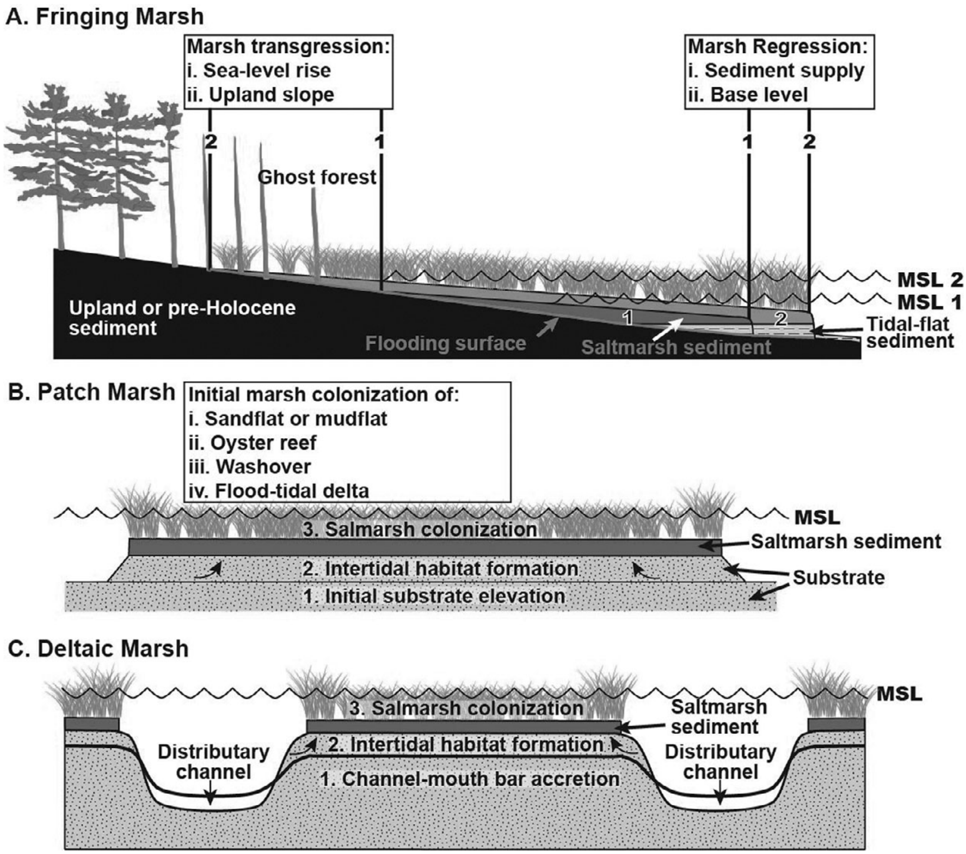

Figure 3.1 Salt marsh is grouped into three general classifications. Fringing marsh (A) forms with salt marsh transgression of the upland, which is modulated by the rate of RSL rise and upland slope. Fringing marsh also forms with salt marsh regression at the shoreline, which is modulated by base level and sediment supply. Patch marsh (B) is disconnected from the upland shoreline of the estuary and forms when salt marsh vegetation colonizes intertidal substrate such as tidal flat, oyster reef, washover fan, or flood-tidal delta bar. Deltaic marsh (C) forms at river mouths where sedimentation at the delta front and delta plain forms intertidal habitat suitable for salt marsh colonization.

RSL rise inundates upland areas. The notion that a salt marsh expands landward with RSL rise is not new (e.g., Davis, 1910 and Redfield, 1965), and in addition to geological data sets that show salt marsh landward expansion over millennia (Gardner and Porter, 2001; Tornqvist et al. 2004; Kemp et al. 2017), salt marsh transgression over the past 100 years is apparent from historic maps, aerial photographs, and satellite imagery (Williams et al. 1999; Feagin et al. 2010; Raabe and Stumpf, 2016). The stratigraphy of the landward portions of all fringing marshes shows a deepening-upward succession of environments with upland soil or old strata (Pleistocene or older) observed directly below salt marsh peat (Davis, 1910; Johnson, 1919). The contact between fringing salt marsh and underlying strata near the upland boundary is always a flooding surface (Fig. 3.1A). As sea-level rises

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the upland inundates, salinizes, and becomes tidal, making conditions conducive to salt marsh colonization. The rate of salt marsh transgression increases as the rate of RSL rise increases and/or as the upland gradient decreases (Belknap and Kraft, 1985; Davis and Clifton, 1987; Kraft et al. 1992; Theuerkauf et al. 2015; Kirwan et al. 2016). In some locations, however, the upland slope is steep (Kraft et al. 1992; Theuerkauf and Rodriguez, 2017) and fronted by a narrow fringing marsh that provides little wave reduction (Möller et al. 2014). In that morphological setting, storm waves could erode the upland and lower its slope. This would make any predictive model of salt marsh formation that relies only on geometrical relationships between RSL rise and the slope of the land being inundated problematic.

Fringing salt marsh can also expand basinward when sedimentation rates exceed the rate sediment accommodation is created (Fig. 3.1A). In most estuaries, wind-driven waves control the depth of erosion by establishing a local wave-base level, which ultimately controls the level beneath which sediments can accumulate (Shideler, 1984; Nichols, 1989; Simms and Rodriguez, 2015). As a result, in estuarine central basins, sediment accumulation generally keeps pace with RSL rise and average water depth remains constant over millennia (Nichols 1989).

The bathymetry adjacent to fringing marsh shorelines is more dynamic than in the central estuarine basin because the preservation of strata is more variable at the shoreline. For example, changing shoreline morphology can produce a local sheltering of wind waves (e.g., spit accretion) and a change in prevailing wind direction can decrease wave-base near a shoreline or transform it from one extreme to another (Gunnell et al. 2013). Under conditions of high sediment input, any decrease in base level can produce a rapid and areally extensive decrease in water depth. This shallowing is different than what is observed near a river mouth, where flow divergence results in sedimentation, formation of clinoforms, and a gradational increase in water depth (Simms at al., 2018). Rather, at the fringing salt marsh shoreline adjacent flats rapidly accrete vertically to an equilibrium elevation where the new intertidal flat is colonized by salt marsh (Fagherazzi et al. 2006). Basinward of this expanded salt marsh there can be a sharp bathymetric transition between the new shoreline position and the adjacent subtidal flat, the depth of which is in equilibrium with wave base (Fagherazzi et al. 2006). Slumping of steep upland terrane along estuarine shorelines can also supply nearshore areas with sediment forming new intertidal flats that rapidly become colonized with salt marsh vegetation (Kelley et al. 1988).

The basinward expansion of fringing marsh occurs episodically, as conditions on the adjacent tidal flat cycle between being conducive for deposition or favorable for resuspension. In the British Isles, cyclic basinward expansion forms a terraced fringing marsh morphology (Allen, 2000; Fig. 3.2). Each terrace represents an episode of marsh colonization of an accreting wave-cut platform, vertical marsh accretion, and finally a transition to an erosional regime with cliff formation and landward migration of the shoreline. As conditions basinward of the cliffed shoreline change to depositional, the cycle begins again as new salt marsh colonizes the accreting platform. This new marsh is at a lower elevation than the landward adjacent older marsh because it is less mature, and even though the new marsh platform is accreting at a greater rate than the landward

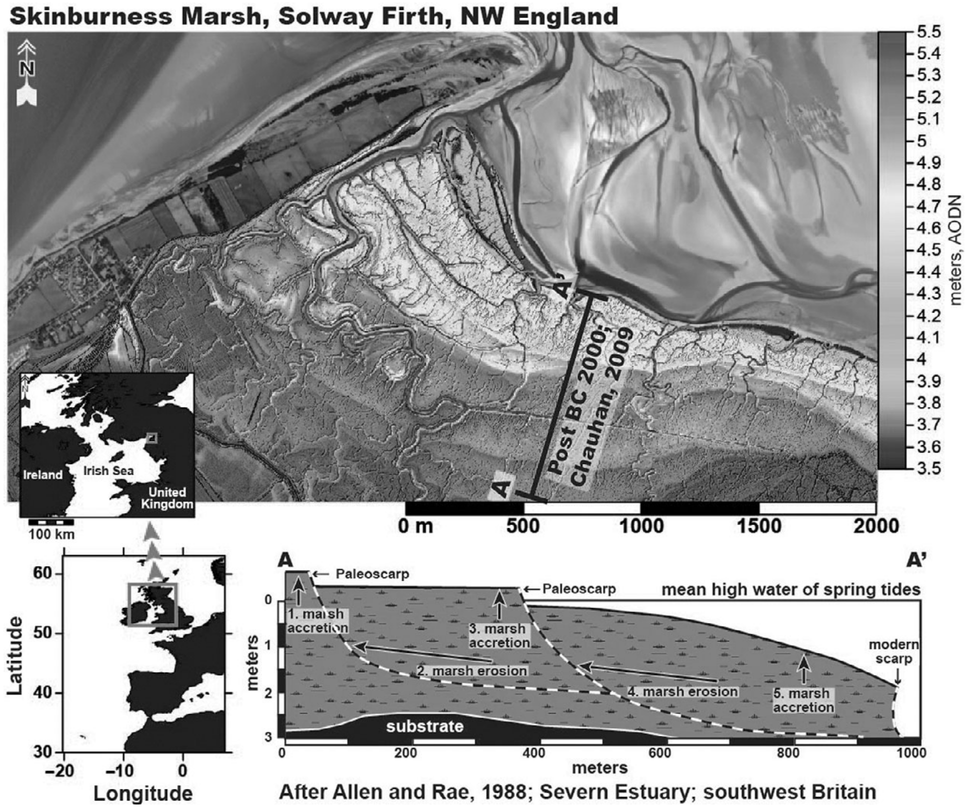

Figure 3.2 Digital elevation model of the Skinburness (Cumbria, UK) fringing salt marsh shows a series of terraces 0.5 m–0.1 m in relief. The delineated area accreted discontinuously over millennial time scales (Singh Chauhan, 2009). Cross section A–A′ is from the margin of the Severn Estuary, another terraced marsh in the UK, and illustrates that the terraces are related to periods of marsh accretion and erosion (after Allen and Rae, 1987). AODN = Above Ordnance Datum Newlyn. Data used to create the digital elevation model was downloaded from data.gov.uk. (A black and white version of this figure will appear in some formats. For the color version, please refer to the plate section.)

higher-elevation marsh terrace, its elevation has not caught up to the older marsh yet (Pethick, 1981; Allen, 2000). In addition to intrinsic factors promoting variable sedimentation rates across the marsh and adjacent tidal flat (Singh Chauhan, 2009; Allen and Haslett; 2014), older marsh terraces are also at a higher elevation due to RSL fall during the Holocene, upon which the marsh accretion–degradation cycles are superimposed (Shennan and Horton, 2002; Allen, 2000; Allen and Haslett; 2014).

In areas of Holocene RSL rise and rapid marsh accretion the same accretion–degradation cycles are recognized during basinward marsh expansion, but are not associated with a terraced marsh surface expression. Schwimmer and Pizzuto (2000) identified two accretion–degradation cycles in Rehobooth Bay, Delaware, USA under conditions of RSL rise (Fig. 3.3). The more recent episode of fringing marsh basinward expansion

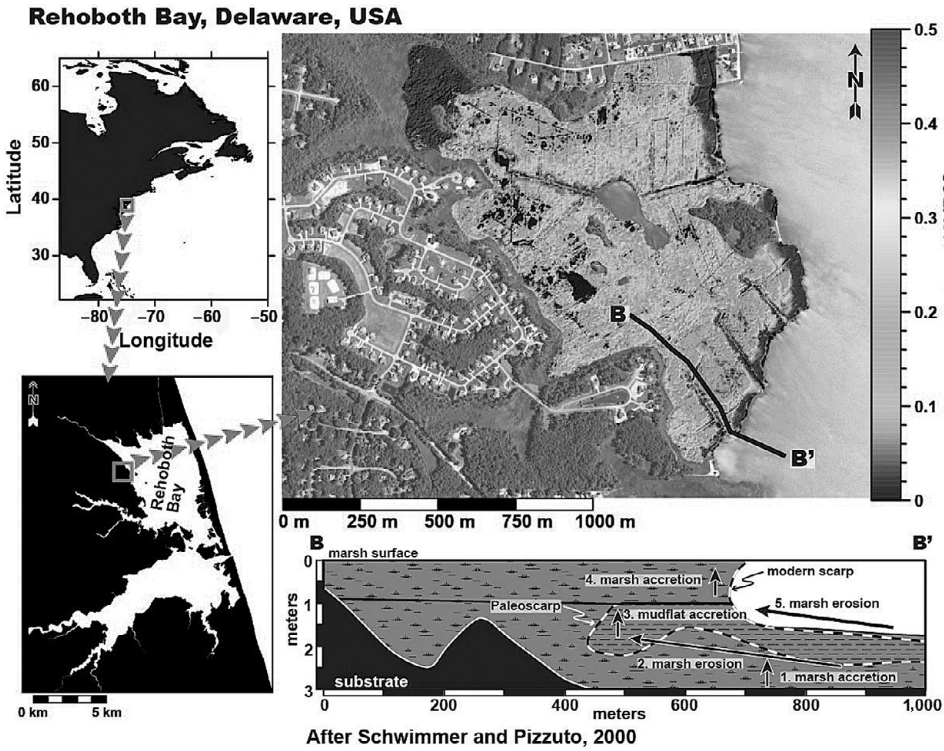

Figure 3.3 Digital elevation model of fringing salt marsh around Rehoboth Bay, DE, shows higher elevation at the shoreline and along the margins of channels. Cross section B–B′ shows discontinuous bayward accretion of the salt marsh shoreline and a buried regional erosional scarp with no surface expression (i.e., a terrace). At around 700 yr BP the marsh shoreline transitioned from erosion to accretion and the paleoscarp formed (Schwimmer and Pizzuto, 2000). NAVD88 = North American Vertical Datum of 1988.

completely buried the paleoscarp that formed during the initial period of erosion (Fig. 3.3). The relief of the preserved paleoscarp is similar to the modern scarp and the paleoscarp is buried in ~1 m of salt marsh sediment. Salt marsh surface topography is relatively flat and shows no evidence of the episodic salt marsh regression preserved in the stratigraphic record, as the current salt marsh elevation achieved equilibrium with sea level.

3.5 Patch Salt Marsh

Patch salt marsh forms in the center of estuaries or in backbarrier environments and patch marshes initially grow on tidal flats not uplands (Fig. 3.1B). The formation of patch marshes is commonly event driven. Storms can cause sandflat migration and vertical accretion to intertidal elevations. Salt marsh will subsequently colonize that intertidal sandflat. Storms can also cause barrier island overwash, washover deposition, and subsequent salt marsh colonization of the new intertidal substrate. Patch marshes are thinner and have existed for a shorter period of time than fringing marshes, which initially formed and expanded landward with the estuary and can reach >5 m depth (Allen and Posamentier, 1993; Belknap et al. 1994; Gehrels et al. 1996). Patch marshes currently in equilibrium with sea level generally show a shallowing-upward succession of depositional environments with mud- or sandflat, oyster reef, flood-tidal delta, or washover fan sediments at the base (Fig. 3.1B).

As the bottom of the estuary accretes above mean sea level (MSL), away from the margin, patch salt marsh formation can occur. In many locations, salt marsh will colonize intertidal flat and oysters will subsequently settle along the margin of the salt marsh and form reefs (e.g., Ridge et al. 2017). Patch salt marsh substrate, however, is not only specific to sand- and mudflats. In warmer latitudes, vertically building oyster reefs can also be colonized directly by salt marsh vegetation (Fig. 3.4). Patch oyster reefs can accrete rapidly (10 cm/yr; Rodriguez et al. 2014) to their maximum intertidal elevation, which is close to MSL (Ridge et al. 2017) and the minimum elevation required for salt marsh accretion (Morris et al. 2002). When the elevation of an oyster reef is above MSL, salt marsh can out compete the oysters. As the salt marsh vegetation begins to colonize, an increase in plant-stem density will reduce flow, increase sediment trapping, and provide positive feedback conducive to salt marsh accretion and detrimental to the health of the oysters that are already stressed from limited inundation associated with their position on top of the reef. In the Shallotte River Estuary, North Carolina, USA, we sampled 48 cm of salt marsh sediment above an intertidal oyster reef in core SRE-MARSH-1, taken from the center of a patch salt marsh with a fringe of oyster reef surrounding its margin (Fig. 3.4). No articulated oysters were preserved at or directly below the contact (45–55-cm depth), indicating productivity on top of the reef was low when marsh colonization initiated. The flanks of the oyster reef remained productive after the top of the reef transformed into salt marsh (Fig. 3.4).

Patch salt marsh also commonly forms in backbarrier environments, where formation of intertidal substrate is directly linked to those changes in barrier-island environments that take place during landward migration (Johnson, 1919; Kraft, 1971; Godfrey and Godfrey, 1974; Theuerkauf and Rodriguez, 2017). RSL rise principally forces barrier islands to migrate landward and during migration, storm overwash creates new intertidal sandflats through washover, and/or flood-tidal delta deposition. After the storm, the barrier island will regain elevation through dune accretion, which will prevent subsequent overwash, and tidal inlets will close or migrate down drift. These post-storm morphological changes to the barrier reduce the tidal energy and stabilizes the backbarrier intertidal sandflats associated with the washover or the flood-tidal delta interdistributary-channel bars, creating conditions conducive to the growth of salt marsh vegetation.

Cheeseman Inlet, located on Bogue Banks, North Carolina, USA, is an example of how changing barrier-island morphology forms new patch salt marsh (Fig. 3.5). Cheeseman Inlet formed after a storm breached the barrier island and connected Bogue Sound with the Atlantic Ocean for 10–15 years (Fisher, 1962) before it closed in CE 1806 (Fig. 3.5A). After Cheeseman Inlet closed, tidal energy in Bogue Sound decreased, and salt marsh colonized the relic flood-tidal delta (Fig. 3.5B). Cores CI-17-2 and CI-17-1, from two individual patch salt marshes, sampled 33 and 16 cm of salt marsh sediment above

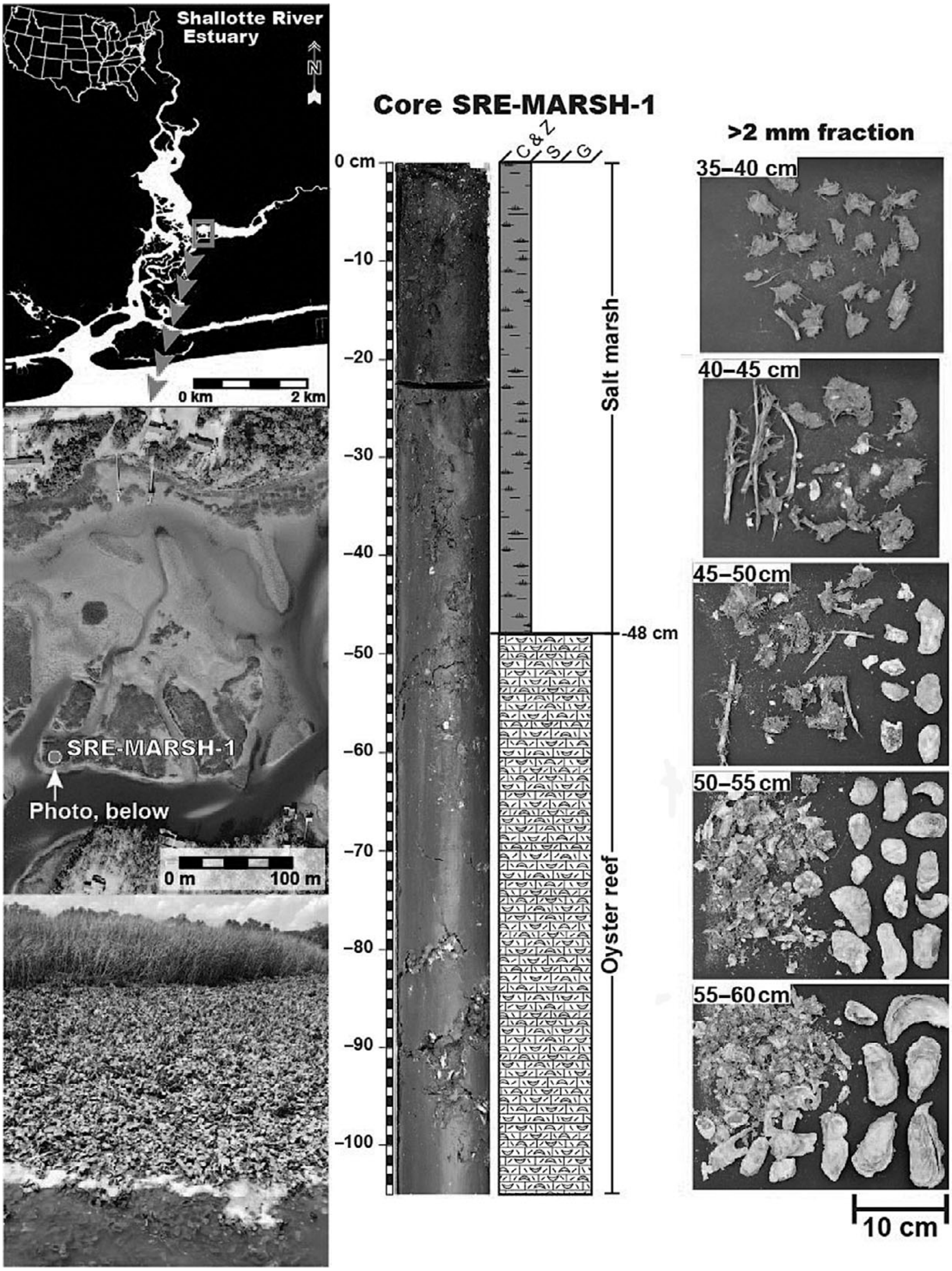

Figure 3.4 Patch salt marsh formed on oyster-reef substrate in the Shallotte River Estuary, NC. Core SRE-MARSH-1, collected from the center of the patch salt marsh, was sampled continuously into 5-cm long sections (10-cm diameter). Those samples were washed through a 2-mm sieve and photographed. The core sampled 48-cm of salt marsh above a >57-cm thick oyster reef. From the top of the core, interval 55–60 cm was the first to sample large articulated oysters (>10-cm long) and no salt marsh vegetation, interpreted to represent the end of a productive oyster reef.

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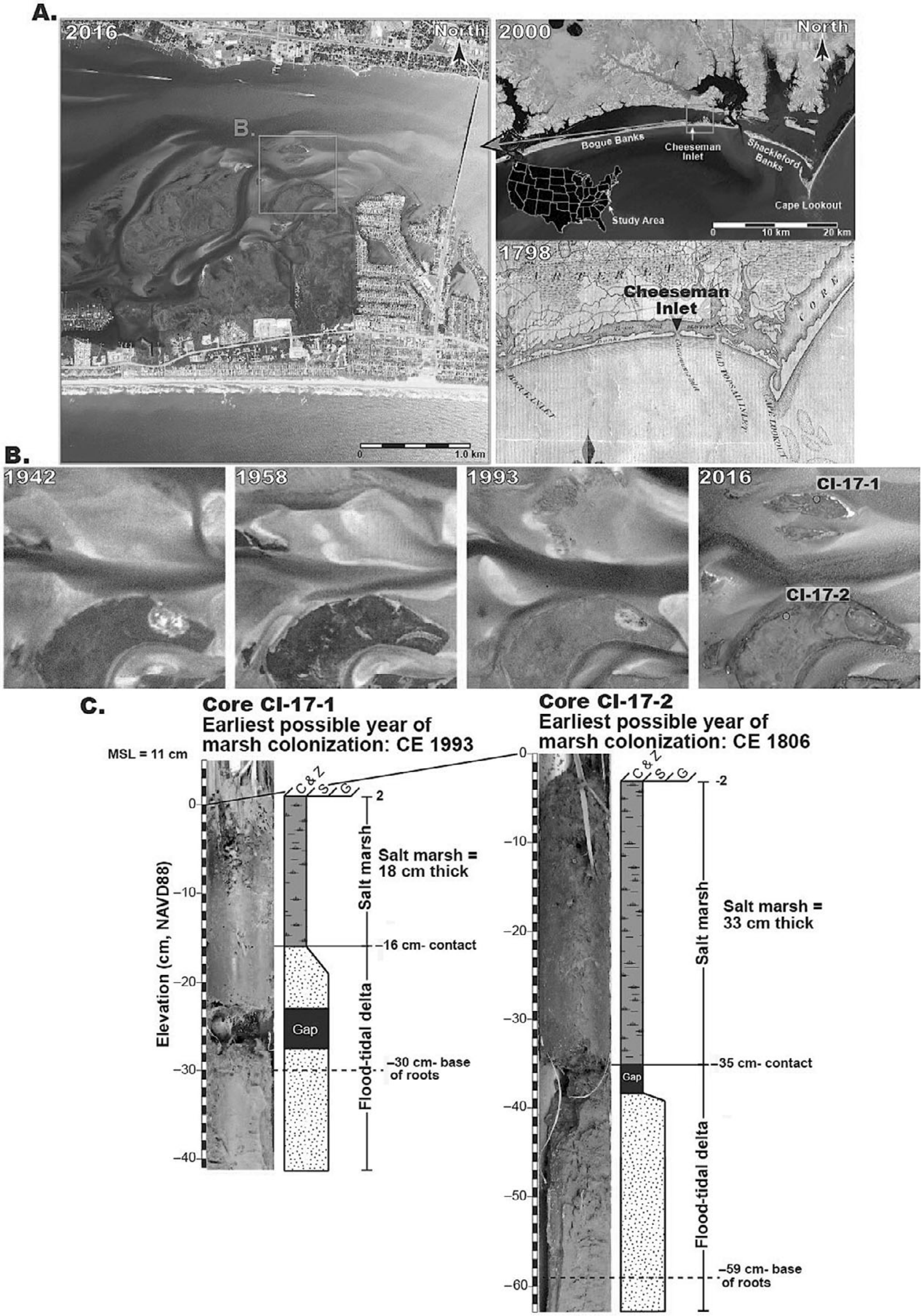

Figure 3.5 (A) Cheeseman Inlet formed around CE 1790 at Bogue Banks, NC. (B) Patch salt marsh colonized the flood-tidal delta soon after the inlet closed in CE1806 and distal parts of the flood tidal delta continue to form new patch salt marsh as physiochemical conditions become conducive for the growth of salt marsh vegetation. (C) Cores CI-17-1 and CI-17-2 were collected from younger and older patch salt marshes, respectively, and accreted vertically at the rates similar to RSL rise. CE = Common Era.

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flood-tidal delta sand, respectively (Fig. 3.5C). Salt marsh likely formed immediately after conditions were conducive for vegetative growth. Core CI-17-2 was collected from a part of the flood-tidal delta colonized by salt marsh shortly after Cheeseman Inlet closed, and the thickness of the salt marsh sampled is within the 43 ±12-cm range of RSL rise from CE 1806 to 2005 (Kemp et al. 2017). From CE 1942 to 1993, a sandflat along the northern part of the relic flood-tidal delta accreted to an intertidal elevation and was subsequently colonized with salt marsh vegetation sometime between CE 1958 and 1993, based on historical aerial photos (Fig. 3.5). Core CI-17-1 was collected from this younger patch salt marsh and the thickness of salt marsh sampled is close to the 11 ±5 or 22 ±5 cm of RSL rise (linear MSL) measured from CE 1993-2017 and CE 1958-2017, respectively, at the NOAA 8656483 tide gauge, located 7.3 km from the patch salt marsh. Based on historical maps and aerial photos that document the time of sandflat formation and marsh thicknesses that match the increase in sea level that occurred since sandflat formation, it is likely that both patch salt marshes accreted at average rates of RSL rise. In addition, the timing of initial salt marsh colonization, extrapolated from sea-level records and maps, suggests that natural patch salt marsh formation can occur rapidly in less than a decade.

Similar to fringing salt marsh, backbarrier patch salt marsh can transgress the upland margin with RSL rise, expanding towards the center of the barrier island. This expansion towards the barrier, however, is in a seaward direction and will be short-lived because the planar contact between salt marsh and the overlying barrier-island migrates landward with RSL rise and storms. Ultimately, this will decrease the area of backbarrier patch salt marsh by burying it in sand through aeolian and overwash processes (Theuerkauf and Rodriguez, 2017). The formation of backbarrier intertidal substrate for colonization by salt marsh vegetation will have a net landward movement as barrier-island transgression continues, and the lagoon narrows and shallows. Eventually, the lagoon will decrease in size, fill with sediment, and transform into an extensive fringing salt marsh confined to the area between the barrier and the upland when the mainland fringing salt marsh and backbarrier patch salt marshes merge.

3.6 Deltaic Salt Marsh

Deltaic salt marsh exists at the heads of estuaries (bayhead deltas) and along river-dominated open-ocean shorelines (wave-, tide-, and river-dominated deltas; Galloway, 1975). Deltaic salt marsh formation is controlled by the deposition of a river's load at the margin of a basin, the associated accretion of intertidal flats, and subsequent changes in delta morphology and hydrology that make physiochemical conditions suitable for salt marsh vegetation to colonize (Fig. 3.1C). Salt marsh can be widespread in the delta-plain, delta-front, and interdistributary bay environments. The delta plain extends from the first distributary channel to the shoreline and the delta front encompasses the shoreline, distributary mouth bars, and adjacent basinward-sloping bed (Bhattacharya, 2006). Commonly, the upper delta plain is composed of freshwater wetland and the lower delta plain contains salt marsh, with the demarcation being strongly controlled by discharge, slope, and tidal range. The delta plain extends seaward as distributary-mouth bars accrete, become intertidal, and force the distributary channel to bifurcate and/or deflect in the direction of longshore transport. Delta lobes form and accrete by the coalescing of distributary-mouth bars and the formation of numerous orders of distributary channels (Olariu and Bhattacharya, 2006). Salt marsh forms in low-energy settings on distributary-mouth bars, crevasse-splays, and/or along the margins of interdistributary bays and back-bar lagoons.

Salt marsh formation at river dominated deltas, like the Mississippi Delta, is linked with delta-lobe building, abandonment, and deterioration (Penland et al. 1988; Roberts, 1997). Reed (2002) presented a conceptual model that incorporated ecological processes into the "delta cycle" (Roberts, 1997). As a delta lobe forms, salinity of the coalescing distributary bars is low and freshwater marshes form rapidly (Cahoon et al. 2011; Oliver and Edmonds, 2017). After the delta lobe is fully developed, the river avulses, sedimentation < RSL rise, the area becomes a brackish interdistributary bay, and intertidal freshwater wetlands convert into salt marsh. Most salt marsh in the modern Mississippi Delta is located within interdistributary bays (Howes et al. 2010). The interdistributary bay continues to expand and salt marsh area declines with RSL rise. Eventually, accommodation becomes high enough that the river shifts back into the area, sedimentation > RSL rise, and the autogenic delta cycle begins again (Penland et al. 1988; Roberts, 1997). Human activities, like fluid withdrawal, damming, and levee construction disrupt this cycle and offset the balance between marsh formation and destruction, leading to massive wetland loss and an increase in the vulnerability of coastal populations to flooding and inundation (Kennish, 2001; Blum and Roberts, 2009; Syvitski et al. 2009).

Bayhead deltas are located at the heads of estuaries, confined to incised valleys (Dalrymple et al. 1992). Deltaic salt marsh formation occurs on the lower delta plain and distributary mouth bars, which are protected in the estuary from erosive open-ocean waves and currents (Saintilan and Hashimoto, 1999; White et al. 2002). Deltaic salt marsh is less extensive at bayhead deltas than open-ocean deltas because river loads and the space available for intertidal channel-mouth bars is lower. Similar to the delta cycle of Penland et al. (1988), bayhead deltas and associated salt marsh also experience cycles of rapid transgression followed by stability and growth (Thomas and Anderson, 1994; Rodriguez et al. 2010; Anderson et al. 2016; Simms et al. 2018). These autogenic cycles are due to RSL rise forcing the bayhead delta to migrate landward through the irregular dendritic drainage network of the lower watershed. Tributary junctions are associated with an abrupt increase in river gradient, causing a decrease in the sediment accommodation/sediment supply ratio and bayhead delta growth and stability (Simms and Rodriguez, 2014).

Human activities, like impoundments, can decrease wetland accretion in bayhead deltas (White et al. 2002; McKee et al. 2006; Canuel et al. 2009; Jalowska et al. 2015; Jalowska et al. 2017); however, land-use changes can have the opposite effect. Clearing of land for farming during initial human settlement resulted in increased runoff, high river sediment load, and delta plain accretion from legacy sedimentation (James, 2013; Watson and Byrne, 2013; Jalowska et al. 2015). Bayhead deltas can respond rapidly to changes in land use. Sediment from landscape erosion at a silviculture operation in the Newport River watershed, North Carolina, USA, took <3 years to reach the bayhead delta. The addition of this new sediment source to the river load resulted in delta-plain salt marsh formation to

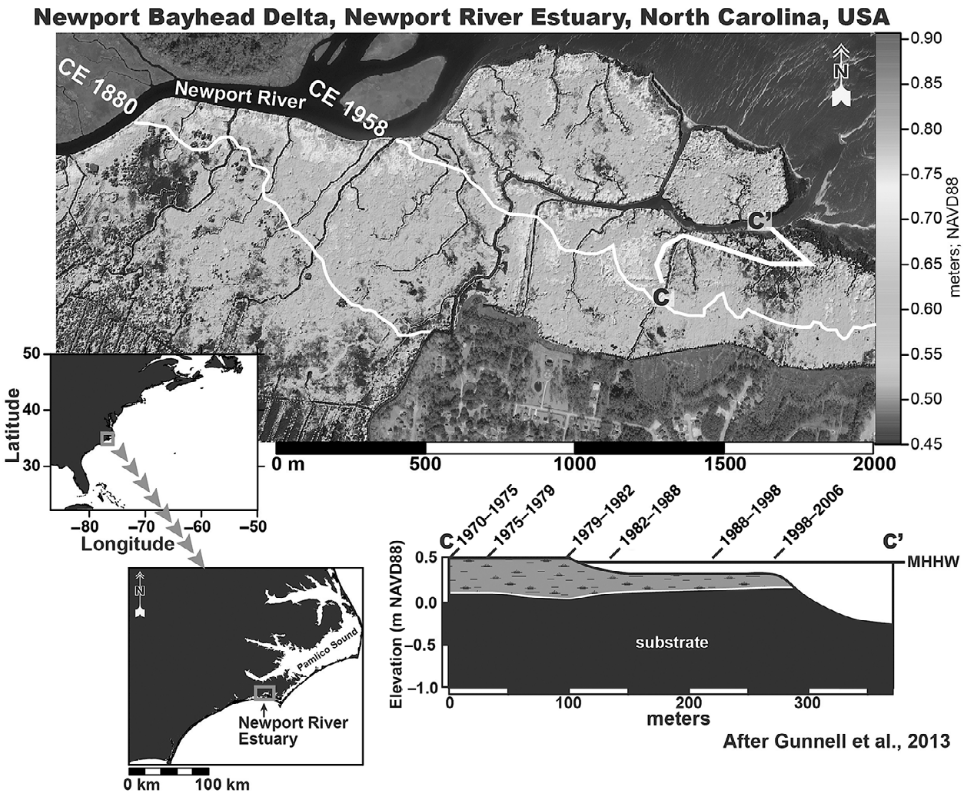

Figure 3.6 Digital elevation model of deltaic salt marsh at the Newport River Bayhead Delta, NC showing bayward accretion. Cross section C–C′ showing variations in deltaic salt marsh thickness and ages based on a time series of aerial photographs and cores (see Gunnell et al. 2013 for additional information) NAVD88 = North American Vertical Datum of 1988. (A black and white version of this figure will appear in some formats. For the color version, please refer to the plate section.)

abruptly increase from 10,000 m^2 yr^{-1} to 15,000 m^2 yr^{-1} (Mattheus et al. 2009; Fig. 3.6). Similar to the other types of salt marsh, the emergence of vegetation occurred rapidly, but was not punctuated by erosional events, like fringing salt marsh commonly is (Figs. 3.2, 3.3, and 3.6). Salt marsh formation at the Newport had an along-shore trajectory, promoted by marsh promontories that shielded adjacent mudflats from erosive forces in the already sheltered estuary (Fagherazzi, 2013; Gunnell et al. 2013; Fig. 3.6).

3.7 Salt Marsh Restoration and Rehabilitation

Salt marsh formation via restoration and rehabilitation is commonly carried out to counter losses from human impacts such as salt marsh reclamation, burial, excavation, pollution, development, and shoreline armoring, in addition to alteration of hydrology, sedimentation, and subsidence (Kennish, 2001; Adam, 2002). After restoration, the full development of salt marsh ecosystem function (vegetation, macroinvertebrate populations, fish and bird use) can take 20 years (Craft, 2000; Warren et al. 2002), while soil development and carbon accumulation can take more than a century before it is on par with a natural marsh (Craft et al. 2002). Salt marsh rehabilitation commonly takes the form of changing the physiochemical conditions around the margin of an estuary or delta, making them more conducive to salt marsh growth. Nearshore breakwater sills and groins protect salt marsh from erosion, decrease wave power, and increase sedimentation providing new intertidal substrate for salt marsh expansion (Adam 2002; Currin et al. 2008). Removal or modification of embankments, dikes, or walls at salt marsh sites returns tidal flow, sediment, and salt marsh vegetation to reclaimed areas (Bakker et al. 2002; Williams and Orr, 2002). In many deltaic salt marshes, thin-layer deposition of dredged material (Ford et al. 1999; Graham and Mendelssohn, 2013) and/or reconnection of the river to the delta plain (Day et al. 2007) conserves intertidal elevations and provides additional intertidal substrate for salt marsh colonization through increased sedimentation. Salt marsh restoration takes the form of planting salt marsh vegetation at sites where the requirements for salt marsh establishment are met (Broome, 1988). Restoration projects span a wide range of scales from <0.1 ha to >6,000 ha for the South Bay Salt Pond Restoration Project, San Francisco, California, USA (in progress). Salt marsh formation can also occur by introducing new species to an area, such as *Spartina alterniflora* that was introduced to China in 1979. *Spartina* was planted in the Yangtze Delta during the 1990s to increase reclamation by accelerating the accretion of tidal flats and offshore sands (Chung et al. 2004; Zhang et al. 2004; Xiao et al. 2010). This strategy worked well due to the high sediment load of the river, and a small restoration project in the Yangtze Delta expanded rapidly from 108 m^2 in 1988 to 12.74 km^2 by 2001 (Zhang et al. 2004).

3.8 Conclusions

Specific physiochemical conditions are required for salt marsh formation, the most important being brackish water, intertidal elevation, and low hydrodynamic energy. Most salt marsh falls into one of three different formation classifications, including fringing, patch, and deltaic. Fringing salt marsh expands landward with sea-level rise and bayward as tidal flats accrete and large expanses are colonized with salt marsh vegetation. Many fringing marshes are expanding landward through transgression and eroding along the shoreline, with the net growth or decay being determined by those relative rates. Patch salt marsh is initially disconnected from the upland and forms on intertidal substrates including flood-tidal deltas, washover fans, tidal flats, and oyster reefs. The conditions that promote patch salt marsh formation and degradation are tightly linked with the evolution of adjacent environments (e.g., barrier island rollover and oyster reef accretion) and changing hydrodynamics (e.g., a migrating tidal inlet) and storminess. Deltaic salt marsh forms on channel-mouth bars and within the lower delta plain including interdistributary bays. Deltaic salt marsh formation and degradation is coupled with autogenic delta processes (avulsion), sea-level rise, and human modifications to rivers (impoundments, levees, land-use change).

The conditions required for salt marsh formation and health often run counter to the needs and desires of humans. This has driven the massive historical decline in salt marsh area, which is projected to accelerate with climate change. Salt marsh restoration and rehabilitation is commonly employed to return some of the lost ecosystem services, but it is unlikely to keep pace with declining salt marsh area. Eventually, accelerating sea-level rise will adversely impact most salt marshes, the most resilient being in areas where sedimentation is high, such as flood tidal deltas, prograding river deltas, or along the margins of estuaries with high suspended-sediment concentrations. Landward migration will extend the life of some fringing salt marshes but increasing development along estuarine shorelines and steep upland gradients will be limiting factors.

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