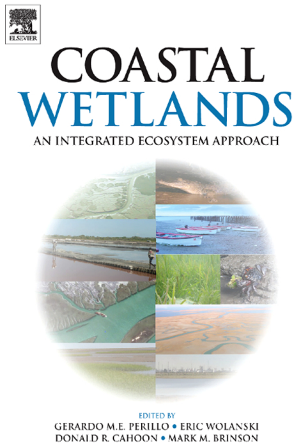

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CHAPTER 6

Tidal Courses: Classification, Origin and Functionality

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Contents

1. Introduction

Tidal courses (otherwise know as channels, creeks or gullies) are the most distinctive and important features of coastal environments even with a minimum tidal influence. They represent the basic circulatory system through which water, sediments, organic matter, nutrients as well as pollutants are transported in and out of these wetlands. The tide is the heart that pumps water through the arteries and veins of the system allowing the exchange with the open ocean and the exportation of materials (including pollutants) from it. As tide enters any of these environments, such as tidal flats, salt marshes or mangroves, it first becomes channelized until water level overflows channel banks and levees and develops a sheet flow. Ebbing is the reverse process, first water recedes as sheet flow but final drainage is through the courses.

Without tidal courses, life in coastal wetlands could not be possible as they provide nourishment, protection for local fauna and a place for reproduction, the growth of juveniles and, finally, a way out, when mature, to the open ocean for numerous species. In fact, tidal courses are one of the first features that appear upon the formation of a coastal wetland either by modification of earlier fluvial networks or by the direct action of the tides, groundwater and precipitation. As the wetland evolves in time and space, courses follow up and, many times, set the pace of this

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evolution since they, being the most energetic environment, are most sensitive to possible changes on the external variables that influence the systems.

Despite their importance, tidal courses have been taken for granted as a feature always present but little studied in comparison with the associated flat, marsh and mangrove areas. When studied, measurements focus on particular features such as biological or chemical composition and circulation (especially in meanders). Only recently, a relatively small group of researchers have started to address the geomorphologic issues regarding tidal courses intensively and, in most cases, their work is concentrated on specific wetlands mostly in western Europe (i.e., UK, Italy and the Netherlands), the Americas (i.e., Argentina, Canada and United States) and Australia, but their ecological relevance is undervalued as reflected by the lack of research on these systems compared to other parts of the wetlands.

Tidal courses are widespread and abundant in estuarine ecosystems (Mallin, 2004). There have been some initial attempts to describe the geomorphologic characteristics of tidal courses and even some basic assumption about the mechanism by which they originate. Some recent reviews (Eisma, 1997; Allen, 2000) provide an integrated approach from the classical geomorphologic point of view of our present knowledge about tidal courses. Work by the Italian group working on the salt marshes of Venetia Lagoon (Fagherazzi et al., 1999, 2004; Marani et al., 2002, 2003; Rinaldo et al., 1999, 2004) have introduced new concepts on how to study tidal course features and evolution.

Furthermore, knowledge of their geomorphologic and sedimentologic characteristics is essential for the interpretation of the stratigraphic record. Until recent, estuaries were seldom found in geological papers as there were no adequate correlations between the present-day conditions and what was observed in geological outcrops (Perillo, 1995). Estuaries in general and specifically tidal courses have relatively small area distributions which are difficult to find in stratigraphic outcrops. In many cases, their imprint may even be confused with unidirectional flows if other criteria (i.e., estuarine fossils) are not present.

The mechanisms that originate the courses are largely unknown. Although this seems a very simple concept, there is no real agreement as to how and when they initiate, not even if there are only one or multiple processes that develop them. In most cases, course initiation is an underwater process occurring over the intertidal area which is commonly occluded from direct observation by suspended sediments. Even if water transparency were not a problem, there is no way to predict when and where a rill will form, or if the courses will persist in time when they are flooded during the next tide.

Probably, many of the uncertainties arise from the simplistic although inaccurate idea that tidal courses are somewhat the bidirectional flow counterpart of fluvial streams. The original studies made by pioneers like Leopold et al. (1964) have specifically compared fluvial and tidal channels trying to apply various classification and statistical criteria employed in the terrestrial counterpart to the marine course networks. Preliminary studies, especially those related to drainage networks (Pethick, 1992; Pye and French, 1993) have considered fluvial terminology to describe tidal course networks although there are clear differences related to hydraulic and geomorphic considerations.

Despite over 150 years of research on fluvial systems, there are still no adequate explanations for many of their basic problems, including the origin of gullies and rills (Tucker et al., 2006), which are very important features compared with tidal gullies and rills, still less is known from tidal courses when the length of time that they have been studied is only a minimum portion of the previous one and, for the most parts, the field study conditions are much more difficult. In this regard, the difficulties to study both origin and evolution of tidal courses in physical models are much greater than fluvial ones due to the lack of widespread laboratories that have tidal-emulating facilities and also the additional problem of resolving the sediment scaling. This is not a minor issue since the large majority of coastal wetlands are dominated by fine (silt and clays) sediments whose behavior varies depending on the local environmental factors and interaction with the surrounding material.

Therefore, the objective of this chapter is to provide a review of the present knowledge of tidal course characteristics, networking and drainage systems as well as to present some ideas and examples of mechanism for course origin. We also provide a possible classification of the tidal courses as a way to simplify the existing confusion based on the indiscriminate use of various common names.

2. Proposed Tidal Course Classification

Upon scanning the literature on wetland morphology, there is a remarkable confusion about the variety of names given to the different valleys that intersect wetland. Names such as tidal channels and tidal creeks are very common and used interchangeably even in the same publication. Other common terms are gullies, rills, canals, and so on. A similar confusion occurs in the fluvial literature, especially regarding the differentiation between rivers and creeks. In the latter case, the problem appears to be related, in most cases, to local denominations without consideration of geomorphologic features notably distorted by the variety of definitions that exist in dictionaries and encyclopedias.

In the present context, we propose a basic definition and classification of tidal courses to provide a common descriptive ground based on size and persistence of water on the course during low tide conditions. We define a *tidal course as any elongated indentation or valley in a wetland either originated by tidal processes or some other origin, through which water flows primarily driven by tidal influence*. Tidal course is a general denomination that includes a series of indentations within a wide spectrum of sizes (width and depth) and with at least two levels of inundation (Table 1 and Figure 1). The classification based on size ranges proposed also provides a descriptive terminology for all tidal courses. Only depth and width of a cross section estimated to the bankfull level are considered since length could suffer large variations, especially if there are artificial or major geomorphologic constrains (i.e., dikes, cliffs, etc.).

Tidal rills (Figure 1a – tr) are very small superficial indentations developing in the later ebb stages along the unvegetated, sloping margins of larger courses or marsh fronts. In order to form, they require a small veneer of fine sediments which

Table 1 General classification and size range for tidal courses

Name	Water in low tide	Depth (cm)	Width (cm)	Cross-section area (cm^2)
Tidal rills	No	<1	<2	<2
Tidal grooves	No	1–5	2–10	<50
Tidal gullies	No	5–100	10–100	50–1,000
Tidal creeks	Yes	10–200	10–200	100–4,000
Tidal channels	Yes	>100	>200	>2,000

Depth is the mean vertical distance from the thalweg to the bankfull border of the indentation. Width is the mean horizontal distance measured across the indentation between the bankfull borders.

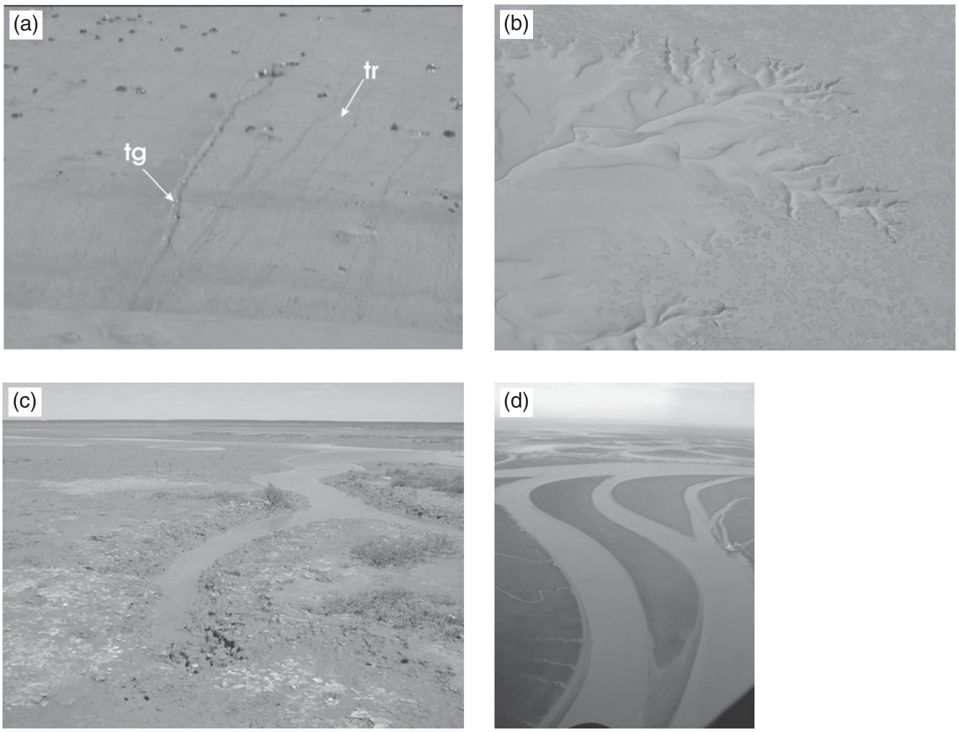

Figure 1 **Examples of end members in the tidal course classification. (a) Tidal rills (tr) and tidal grooves (tg), (b) tidal gullies, (c) tidal creeks and (d) tidal channels.**

is dissected by very low flowing waters. Depending on the surface slope and sediment characteristics (i.e., size, cohesiveness and depth), rills vary in shape from linear to sinuous and may even develop braided conditions (both distributary and convergent). Another common mechanism of rill formation, as is also frequent on sandy beaches, is due to groundwater discharge resulting in a large number of rills having a variety of shapes. Rills may be considered as the first step in the initiation of a tidal course and their preservation and further evolution depends on the depth of the indentation, the soil characteristics and most important, the processes occurring when tides inundate the course margin back. If the indentation progresses in depth to more than 1 cm, the rill becomes a tidal groove.

Grooves (Figure 1a – tg) tend to be 1–5 m long or exceptionally even longer courses, 1–5 cm deep and up to 10 cm wide. Normally, they are linear to sinuous in sectors. They develop along channel margins and marsh fronts with relatively high (3–7°) slopes due to strong ebb flows or groundwater outflows. Tidal grooves are more prone to resist the following tidal inundation as they produce a much deeper indentation in the soil. It is common to observe parallel groves along the margin of typical tidal channels (Figure 5a). Concentration of the flow in some of those grooves may produce the formation of larger courses such as gullies or creeks.

Tidal gullies (Figure 1b) are similar to those observed in continental drylands. They are deep indentations that may reach up to 1 m and much wider than the preceding courses. Gullies are preserved and enhanced by tidal inundation. They could easily superimpose in size with tidal creeks, the difference lies in the lack of tidal water during low tides although some flow may be observed, normally due to rainwater retained on the flats or marshes, or groundwater outflows. In all three cases, the larger relative relief is found near the head of the course becoming shallower as it approaches the mouth. Their course is also linear to sinuous and they seldom develop meanders, but if they do, those meanders appear at their mouth.

The major dynamic structures in wetlands are tidal creeks and channels. Probably, the main difference with the other courses, as indicated, is the permanent inundation by tidally driven water in at least part of the course even in the lowermost tides. Creeks range in size from few tens of centimeters in depth to up to 2 m while width has similar values. They normally have water during low tide; however, water depth during this time varies from practically none at the head to about 10–30% of bankfull depth at the mouth. Creeks, as well as the previous courses, are the tributaries of the tidal channels and form at their banks but they are distributed over tidal flats and marshes. Rills and grooves never reach the level of the tidal flats and marshes and gullies seldom do.

On the other hand, tidal channels are the largest features in most wetlands as they clearly stand out in maps and satellite images. Channels always have water along their whole course even during the lowest spring tides. Their depth is greater than 2 m and maximum values are highly dependent on the characteristics of the wetland, but on average they have as much as 10 m, in many cases reaching up to 30–40 m deep as in the case of coastal lagoon inlets. Regarding channel width, the degree of variability is much larger than in the case of depth. Tidal channel widths start at about 2 m and may reach several kilometers (i.e., Bahía Blanca Estuary and Ord Estuary).

3. Geomorphology of Tidal Courses

At first sight, tidal courses resemble concentrated fluvial networks having also similar course shapes both in plan view and in cross section. However, when analyzed in further detail, numerous differences appear that set them apart.

The most obvious ones are the flow characteristics (roughly unidirectional vs. bidirectional), relative relief, degree of course inundation, evolution pattern, among-course interactions and so on. Although there are wetlands developed in sandy substrates, most of them are dominated by fine sediments, therefore, braided patterns are rare and most valleys have single courses with roughly linear, sinuous or meandering patterns.

Various authors have partially analyzed the cross-section shape of courses in wetlands (Pestrong, 1965; French and Stoddart, 1992; Collins et al., 1998; Allen, 2000; Fagherazzi et al., 2004). The cross section form depends on wetland stratigraphy, tidal range, their associated plant cover and sinuosity (Allen, 2000), and they can be classified as V-shaped (Figure 2a), U-shaped (Figure 2b), asymmetric (Figure 2c), complex (Figure 2d) or with one or both banks significantly overhanging (Figure 2e and f). Beyond a general description of the shapes, there are no actual studies dedicated to relate cross-section forms and the course conditions, evolution and size.

Both V- and U-shaped (Figure 2a and b) are common in small courses and even in the linear portion of channels and creeks, whereas asymmetric cross sections are normally related to meandering reaches. Many linear channels also show clear asymmetries (Figure 2c), specially if they are subject to lateral migration (Ginsberg and Perillo, 2004). These types of cross sections appear during the initiation of the course, when the linear pattern is still present and most of the flow strength is used to deepen and lengthen the course.

Complex cross sections (Figure 2d) appear when bars are present along the course. Kjerfve (1978) described them as bimodal with two channels separated by a shallow area or bar having contrasting residual circulation associated with differential depths (flood and ebb dominance). However, there is still no clear indication whether the differential circulation is due to the bathymetry or the bathymetry is a consequence of the change in direction of the residual circulation often found on vertically homogeneous estuaries (Dyer, 1998) as there are no reports describing the formation process.

Overhanging banks (Figure 2e and f) are the result of differential sedimentary characteristics as they are the remnant of flow undercutting by tidal currents within the course. The overhanging portion, due to higher sediment compaction

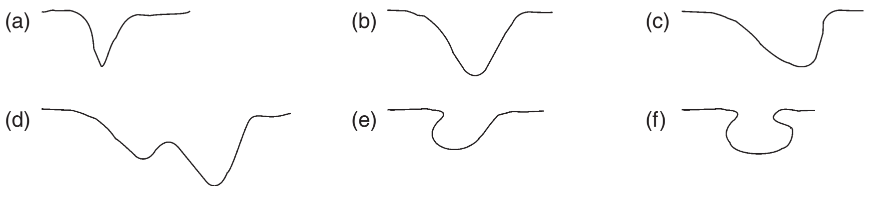

Figure 2 Description of the various possible cross-sections of tidal courses. V- (a) and U-shaped (b) normally corresponds to longitudinal portions of courses. Asymmetric (c) and complex (d) are found commonly in long meanders, whereas overhanging simple (e) and complex (f) represent differential erosion processes controlled by variations in sediment characteristics, water level in the course and/or differential material strength due to marsh plants.

or resistance or sustained by plant roots, has been less affected by the currents and preserved while a lower portion of the bank is undercut. Preservation of overhanging banks is poor as they are intrinsically unstable features that will collapse eventually, and the material is transported by the tidal currents.

In most cases, longitudinal profiles of tidal rills, grooves and gullies tend to be concave upward at the close-ended head becoming shallower and convex upward for most of their course (Figure 1a). At the head of these courses, there is normally fast erosion by headward retreat which induces a relatively high relief represented by a microcliff (Minkoff, 2007). Ebbing water cascades at the microcliff, resulting in a deeper thalweg at the head. In areas where sediment is compact, the head, flanks and bottom of the course at and near the head are marked by dislodged sediment clasts, flakes, crumbs and plant patches, remains of infauna burrows, and so on yielding a very irregular and even chaotic morphology (Figure 3a). Moving along the course toward the mouth, one will find that the relative relief diminishes in addition to the irregularities in the bottom and flanks, since the tidal flow tends to smooth them out by transporting and wearing out sediment blocks (Figure 3b). In recent deposited

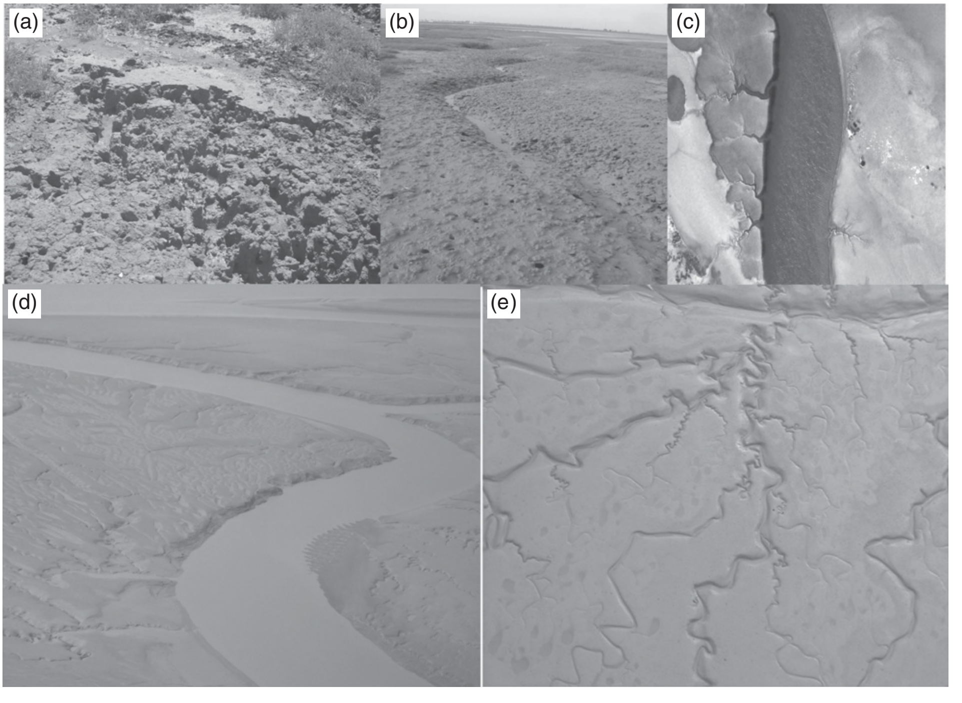

Figure 3 Various examples of tidal course characteristics. (a) Complex structure of a gully head, (b) view of a tidal creek morphology, (c) marked differences in tributaries along a tidal channel, (d) typical meander in a tidal channel at low tide showing the formation of a point bar in the inner flank of the curve and the cliff on the outer flank and (e) variability of the meander pattern over a tidal flat showing that longer curvature radius occur on the flatter portion of the flat.

sediments that have not achieved a certain level of compaction, preservation of small courses becomes difficult as tidal currents along creeks and channels resuspend and transport them and, as the courses are transversal to the flow, the potential to be filled up is high. In these cases, the flanks and bottom are smooth.

Tidal creeks and channels, when established, have relatively smooth flanks and bottoms, worn out by currents, and irregularities appear related to erosional processes at the outer bend of meanders, infauna activities, rotational slumps and deformed microdeltas deposited at the mouth of tributaries. Longitudinal profiles show a progressively higher relative relief toward the mouth. Tributary inlets often show sediment deposits that range in shape from typical deltas to extended banks parallel to the main channel. These deposits result from the sudden drop of the suspended and bedload material being transported during the ebb, especially at or near low water slack. Depending on the strength of the currents, these deposits are either carried away or deformed, normally following the channel current dominance. Most examples found at the Bahía Blanca Estuary (Figure 3c) and several others along the Argentine coast show a continuity of the inlet along the main channel flank bordered by a parallel shoal which, in most cases, is ebb-directed. Potentially, the displacement of the mouth induces an initial meander that may affect the circulation within the tributary. Although there are no studies that assert it, this is clearly an instability on the course plan morphology that could displace headward, resulting in a meandering pattern developed inversely to accepted theories in fluvial systems (Leopold et al., 1964).

Another interesting but often overlooked feature of tidal courses are the scour holes that form at the junction of two courses. Studies of scour holes have concentrated in fluvial junctions where their morphological features have been related to the angle of the junction, tributary/main river discharge ratio, sediment erodability, and so on (Best, 1987). Scour holes in tidal environments have only been mentioned by Shao (1977), Kjerfve et al. (1979) and Ginsberg and Perillo (1999). The results obtained demonstrate that scour holes, having large relative relief in excess of 2 m and up to 17 m, at tidal environments differ significantly in morphology from those found in fluvial systems. Those observed in the Bahía Blanca Estuary, for instance, although elongated in shape as in the case of fluvial holes, have the steeper side (3.5°) at the mouth of the confluent channel and the gentler side (1.5°) seaward. This structure is exactly opposite to scour holes found in river environments. Based on current measurements and sediment transport estimations, there is a clear flood dominance on the steep, inner face and ebb dominance over the gentler, outer flank (Ginsberg and Perillo, 1999). Another difference between holes in fluvial reaches and tidal scour holes is that the latter migrate headward.

Courses in marshlands have an additional mechanism to deepen their channels: levee accretion. Restricted by surrounding vegetation, sediment deposition along the course increases the relative relief between channel bottom and marsh surface. This process also inhibits water exchange, resulting in water and sediment becoming trapped at high water and during surges. Both in marshes and mangroves, vegetation plays an important role in sediment trapping and stabilizing the wetland (Wolanski, 2007). Lateral and headward erosion activities also widen and lengthen the channel (Chapman, 1960). The type of wetland vegetation and the vertical distribution of plant species have significant influence on the degree and rate of stream channel migration and meandering (Garofalo, 1980).

Although meandering flows have been extensively studied, especially in fluvial environments (Leopold and Wolman, 1960; Ikeda et al., 1981), an adequate theory that fully explains meanders is still lacking. The leading process in meander development in fluvial rivers is the redistribution of momentum due to channel curvature (Ikeda et al., 1981). Flow momentum concentrates along the outer bank originating its erosion. Secondary transversal flows driven by superelevation, transport the sediment which deposits on the inside bank resulting in the development of point bars. Furthermore, alternate bars can lead to the formation of meander bends at initial stages of meander development (Blondeaux and Seminara, 1985; Seminara and Tubino, 1992).

Beyond the specific mechanism of meander formation, the main hypothesis as to why a course meanders is based on the concept that meanders elongate the water path thus enabling it to contain more of the water discharge within the same valley distance. Although this is true for rivers and other contained flows, it appears somewhat difficult to transfer this concept to tidal environments where the water is contained within the course only for a part of the time, even though it is the period with the highest velocities, whereas a large percentage of the flow spreads out over the contiguous wetland as the bankfull stage is overcome. The only correlation could be that once the water is concentrated in the course during the ebb, the total water volume may be larger than the actual volume that the course can discharge, thus causing a situation similar to that found in rivers.

Tidal meanders have a cross-sectional morphology similar to fluvial meanders. The outer bank tends toward a steeper flank, sometimes a cliff develops (Figures 2d and 3d) and, eventually point bars evolve on the inner bank. Flow circulation on the meander appears, in general, similar to the circulation in river counterparts although this is a simplification that cannot go beyond the cases in which the meander is symmetric. The fact that in coastal wetlands tides control circulation implies that the flow along courses is bidirectional. Both the bottom topography and overflow are causes for common flow asymmetry which, eventually, could be enhanced by any river discharge. Therefore, flow around a bend in a tidal course is seldom symmetric and the resulting morphology of the bend suffers its consequences.

Meander shape in coastal wetlands, as in fluvial systems too, is controlled by the sediment properties and erosion–sedimentation history of the environment. Wetlands developed over older fluvial terraces or deltas being eroded may have layers of sediment of variable consistencies outcropping and, therefore, they can control the direction and planform shape of the whole course but most importantly that of the meanders. That is the case of the Bahía Blanca Estuary and Anegada Bay (Argentina), which were both part of the late Pleistocene-Early Holocene Colorado River Delta (Perillo and Piccolo, 1999), and it is common to find meanders with a "straight loop" (Figure 3e) following the orientation of the original delta sets. In many cases, meanders are also initiated by rotational slumps or "erosion cusps" (Ginsberg and Perillo, 1990).

Symmetrical currents may erode the banks differently resulting in meanders that are narrower at the bend apexes and wider between bends (Ahnert, 1960). Bend skewing, a normal feature in fluvial meanders due to their downflow asymmetry produces the "gooseneck loops" (Carson and LaPointe, 1983; Fagherazzi et al., 2004). However, due to the flow asymmetry, there are distinct planform morphologies depending on whether the course is ebb- or flood-dominated (Fagherazzi et al., 2004). As in rivers, wetland meanders also have point bars whose shape is controlled by the ebb-dominated tidal currents and the radius of curvature (Barwis, 1978).

The theoretical concept that equilibrium courses should have an exponential increase of their cross section from head to mouth could only apply to the large tidal channels and creeks. The degree of strong meandering paths that most small courses clearly have shows that equilibrium cannot be easily achieved in coastal wetlands. In these environments, the dynamical processes are constantly changing, flow seldom stays concentrated within the course and flow discharge may vary, even with the same tidal amplitude, by a simple change in wind direction or speed. Especially, higher order tidal channels which conduct most of the tidal prism may approach this equilibrium idea. A model for young marsh environments where sedimentation is active demonstrates, in contrast with terrestrial rivers, that salt marsh creeks experience a strong increase in width–depth ratio seaward due to the short duration time of the peak discharge (Fagherazzi and Furbish, 2001). Course widening results from the imbalance between erosion versus deposition and the course cross section does not have time to adjust itself. Furthermore, autoconsolidation of the bottom cohesive sediments and the consequent vertical gradients in resistance properties obstruct the formation of deep courses, yielding the formation of shallow wide courses (Fagherazzi and Furbish, 2001).

On the other extreme, course heads may have the same variability in shape than those observed in fluvial systems. Course heads can be defined as the upslope boundary of concentrated water flow and sediment transport between definable banks (Dietrich and Dunne, 1993). However, we follow the previous definition of course head given by Montgomery and Dietrich (1992) as the upstream limit of observable erosion and concentration of flow within definable banks. Minkoff and Perillo (2002) and Minkoff (2007) have differentiated three types of course heads: diffuse (with an undefined form, generally convergent with the surrounding land, Figure 4a), acute (ending in a point with a width <15 cm, Figure 4b) and open (wide, mostly round shaped with a width ≥ 15 cm, Figure 4c). In each of the head types, different processes act. In diffuse heads, the predominant process is the surface erosion of the sediment whereas other factors such as groundwater or headward erosion are of little consequence. In the case of acute heads, both subsurficial and headward retreat associated to water cascading are the most important factors, whereas subsurficial erosion and mass wasting are the major factors in the formation of open heads.

Studies of the time and spatial evolution of creek and gully heads indicate that their shapes varied as the courses were growing (Minkoff, 2007). For this particular case, the effects of the burrowing crab *Neohelice* (formerly *Chasmagnathus) granulatus* play an important role, and landward advance of the heads are most significant

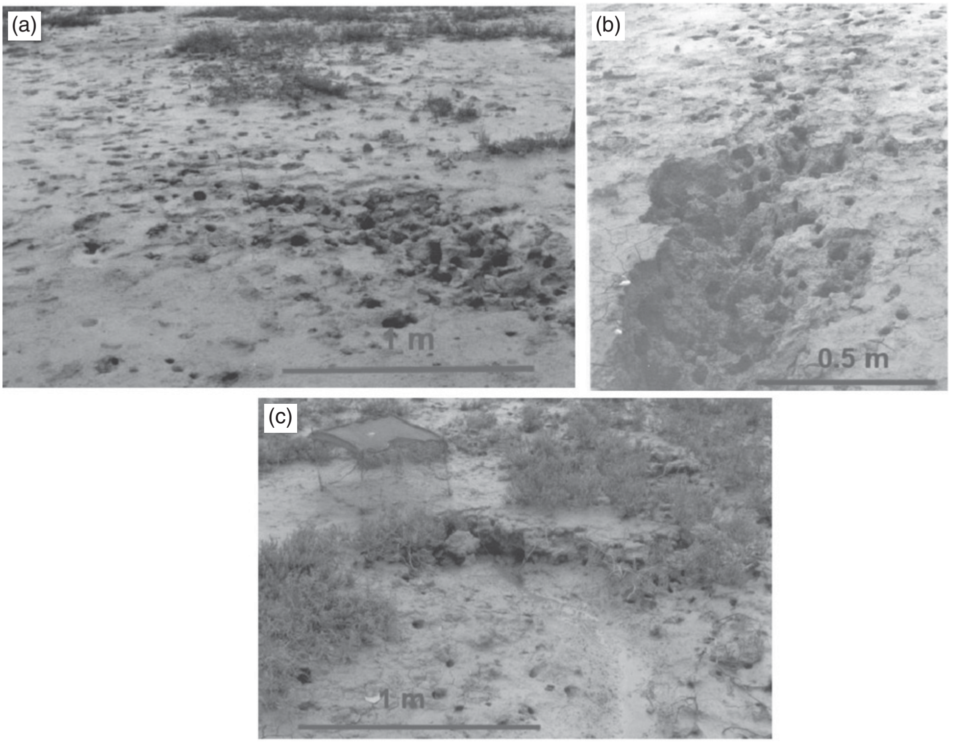

Figure 4 Examples of the three types of gully heads described by Minkoff (2007). (a) Diffuse, (b) acute and (c) open.

(about 2–3 times faster) during the warmer (late spring, summer and early autumn) than colder months due to the high crab activity and population increase (Escapa et al., 2007). For instance, diffuse heads do not grow directly at the head, but there is surficial erosion some distance from the border and then a sudden connection with the thalweg occurs by a depression on the surface. In all cases, Minkoff (2007) has demonstrated, after surveying more than 130 creek heads over a period of 4 years, that creek and gully landward growth is a pulsating process. Basically, there is a period of preparation after which the course retreats a certain distance in one single movement.

Minkoff (2007) seems to be the first detailed analysis of the retreat processes in coastal wetlands, except for the specific study made by May (2002) in which she related creek headward retreat on the level of mass wasting of the terrace between the head and the frontal marsh area incised by the creek. She divided it into strong terrace wasting (STW), weak terrace wasting (WTW) and no wasting (NW). The former occurs in sectors where there are thick deposits of organic-rich soil (at least 10 cm deep). The transition zone of these creeks had a steep slope and a large drop in elevation of the marsh surface (>15 cm). The transition zone presented various erosional features such as large holes (at least 5 cm in diameter) in the sediment surface that were coalescing, as well as undercutting and slumping of sod. Those of WTW appear in areas with thin deposits of organic-rich soil or overlying root mat with a smooth slope. The NW sectors have very little erosive features.

4. Course Networks and Drainage Systems

At first approach, drainage patterns on coastal wetlands resemble fluvial networks. But once they are analyzed in more detail, the first difference that appears is that the number of branches and channel hierarchy (in the Hortonian sense) is closer to mountain basins rather than to lowlands to which coastal wetlands could be related. This factor by itself indicates that the network is draining large amounts of water in short time as opposed to flatland networks. As also occurs with their fluvial counterpart, courses at the head are blind-ended channels for marshes (Ashley and Zeff, 1988) and tidal flats. Blind-ended courses are a mechanism to show local processes that control the network (Stark, 1991).

The second most striking difference is the lack of topographically defined basins. Although some geomorphologic control (likely to have developed after network inception) may exist, basin boundaries are controlled by hydrodynamic processes. Marani et al. (2003) found that the drainage density of a network, which was defined by Horton (1945), as proportional to the ratio of the basin's total channelized length divided by the watershed area, must in fact be defined by the statistical distribution and correlation structure of the lengths of unchanneled pathways. By modeling the current directions on the wetland surface, basin boundaries can be drawn and a network structure defined. A situation that arises because, a large percentage of the time, water on the wetland is moving as overmarsh flow rather than being channelized.

Flow directions under these conditions may be strongly affected by winds especially in areas where wind is a major factor, for instance, by having dominant directions or by storm surges. Also a wetland having different levels of inundation during the regular spring–neap cycle would be subject to different behavior for each sector resulting in modifications of the network. Therefore, drainage patterns at a high wetland could be rather different from that at the lower wetland.

Novakowski et al. (2004), on the other hand, have geomorphologically defined a marsh island as a section of marsh circumscribed by tidal channels that are deeper than 1 m at low tide. Although this definition is somewhat artificial, it provides an objective way for a first-level estimation of hydraulic conditions in a wetland. Marsh course watershed was estimated by connecting end points of first-order courses.

Network density could also be affected by anthropic actions such as ditching. Ditched marshes (a common practice in Europe and North America in the late 19th and early 20th centuries) have both less course density and pool density (Lathrop et al., 2000; Adamowicz and Roman, 2005). Pool density is an important factor in course initiation and development since pool interconnection may lead to course formation. Furthermore, some basin control may occur due to morphological/artificial changes such as dikes or raised continental features. In some cases, the hydraulic control is based on the fact that tides, even during storm surges, cannot inundate the backmarsh. Changes in vegetation patterns can often be a good indicator of hydraulic patterns (Pethick, 1992).

Figure 5 provides a series of examples of the various network patterns that can be observed in coastal wetlands. A variety of linear, sinuous, rectangular and dendritic patterns are common to all wetlands. Their diversity, unless artificially affected, is still a matter of discussion since there are no clear theories that can explain why a particular network pattern is present. Evidently sediment characteristics, previous sediment history and wetland slope all play a role, but to what extent each of them is actually responsible for a particular pattern is presently unknown.

The degree of control that plants have over the development of tidal course networks can hardly be better shown than by the case of Caleta Brightman (southernmost channel of BBE) (Figure 5e). The main channel is borderer in most of its extension by relatively newly developed *Spartina* marshes, which are from a few hundred meters to 1 km wide, and from there to the continent there are muddy tidal flats about 1–2 km wide. Channels and creeks crossing the marsh are mostly

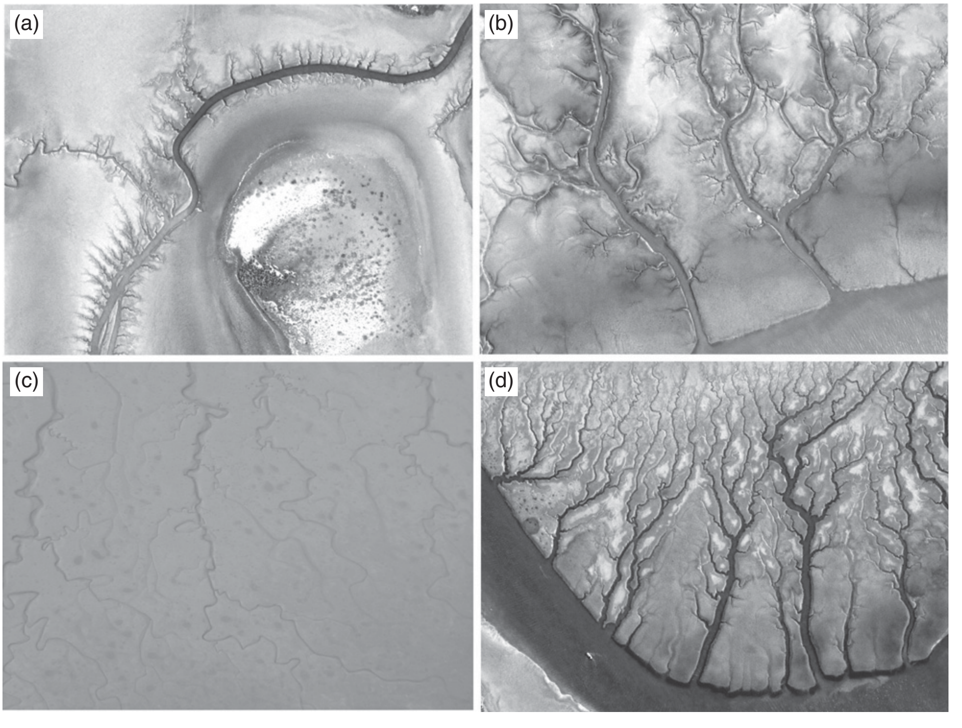

Figure 5 Examples of drainage patterns on a coastal wetland. (a) Rectangular, (b) linear dendritic, (c) sinuous and (d) sinuous dendritic.

linear or mildly sinuous with very few, parallel tributaries connected at right angles to the courses. As these courses are followed into the unvegetated areas, the network suddenly develops into a well-defined dendritic pattern with sinuous to meandering courses. The number of tributaries increases at least threefold, bearing in mind that the flats and marshes are subject to similar inundation regime.

5. Origin of Tidal Courses

After over 200 years of fluvial studies, detailed description of course initiation is still lacking (Montgomery and Dietrich, 1988, 1992, Istanbulluoglu et al., 2002; Kirkbya et al., 2003; Tucker et al., 2006). Obviously, there are many problems to reach consensus within the wetland community (Guilcher, 1957; Chapman, 1960; Pethick, 1969) but also in the well-documented fluvial literature. Based on Chapman (1960), Perillo et al. (1996) described the possible mechanism of tidal course formation as follows: during the ebb tide, sheet flows are guided by the topography, concentrating their discharge into depressions. These depressions become small courses by headward erosion. As the water velocity increases in the course during successive ebb tides due to the tendency to concentrate the retreating flow (Pestrong, 1965), these depressions become wider and deeper. Headward erosion is an important factor in further developing the course. As the course grows in size, meanders form, which are the final pattern that the channel acquires when fully developed. Headward erosion lengthens the course in combination with local annexation of nearby courses.

The main incognita still remains as to why a course forms at the place it does and not somewhere else. For instance, a probabilistic approach was proposed associated with the flank slope and the erodability conditions of the sediment for continental situations (Istanbulluoglu et al., 2002). Following the original Horton (1945) idea that course inception occurs at a certain distance downslope of an incline, many grooves appear at the flanks of tidal creeks and channels during low tide. Horton's idea is that overland flows may require certain distance to achieve enough bottom shear stress to overcome soil resistance. In other words, an erosion threshold controls the location of channel heads establishing that the same critical distance below a topographic divide is required for a sheet flow to erode sediment as for it to initiate a course. According to Horton's theory, courses, such as erosion features, may expand rapidly upslope in response to changed climate and land use conditions and can even form during individual storm events.

Although this is true for thin overflow layers, typical of continental processes, in coastal wetlands during ebb, the water depth along the flanks is at maximum and diminishes as low tide is reached. Nevertheless, at the initial stage of the ebbing phase of the tide, most of the flow may be concentrated along the main course and the flow cascading from the surrounding flats or marshes could be diverted toward the main course mouth and could not reach enough momentum to produce any erosion. As far as we were able to detect, there are no direct measurements of the flow dynamics very close to the surface along a channel margin during a tidal cycle;

therefore, the following analysis is to be considered only as a hypothesis yet to be demonstrated.

Course inception on channel flanks is produced only at the final stages of flow sheet as it moves downslope. At that moment, this sheet is only probably a few centimeter deep; therefore, the flow is both supercritical and most likely turbulent as well. Supercritical flows induce stationary waves that act damming the flow until enough depth is reached to reduce the Froude number at or below the critical value. Then the waves disappear and the flow is strongly accelerated which, in itself, can be strong enough to generate a bottom shear stress greater than the critical value for cohesive sediment erosion. As in most wetlands, this process occurs over fine sediments, which as they are eroded pass directly into suspension and thus increasing the flow density which further augments the erodability capacity of the flow. In the waters along channel borders, at low tide, one can commonly observe a 10 cm to up 1- to2-m wide streak (depending of the channel breadth) of high suspended sediment produced by this downslope erosion.

The incipient grooves have a remarkably constant spacing of the order of 1–10 m which may depend on the slope of the segment of tidal flat where they develop (Perillo et al., 2005). Potentially, all the grooves may develop further into gullies, creeks and channels, but only a few of them actually upgrade to the next level. However, Horton's general idea does not explain the presence of multiple, almost parallel grooves formed over a smooth muddy surface. Evidently, there must be other factors that induce course inception at specific points on the surface. The presence of the periodic grooves may be due to (1) periodic surface irregularities (i.e., along channel undulating surface); (2) presence of pebbles, infauna burrows or mounds; (3) nonuniform sediment (i.e., size, degree of compaction, mineralogy, bioturbation, etc.); (4) presence of vascular plants or roots; and (5) variability of the layer of microphytobentos and their excretions (i.e., EPS). Any or a sum of these items could be responsible for routing the flow, inducing hydraulic jumps. Eventhough all those conditions can be found individually or in groups acting together in helping the development of grooves, there is an important issue along channel banks that is often overlooked: groundwater seepage. Most studies of groundwater exchange between the flats and the marshes have been focused on the biogeochemical aspects (Findlay, 1995; Hollins et al., 2000; Osgood, 2000; Kelly and Moran, 2002; Duval and Hill, 2006).

Gardner (2005) and Gardner and Wilson (2006) modeled the seepage from channel banks indicating that this process is a major mechanism for water and nutrient exchange between the course and the wetland. The process is driven mostly by the tide but affected by rain, plant roots and infauna burrows, resulting in complex and highly transient variations in boundary conditions along creek banks and the marsh platform (Gardner, 2005), and the alternate development and disappearance of water tables, seepage faces and zones of saturated and unsaturated flow. Tidal rills are one of the main consequences of seepage (Figure 1a) as they mostly develop from the groundwater discharge through the saturated zone on channel and creek banks. Rill initiation associated with groundwater discharge, as also occurs on sandy beaches or drylands, is due to sediment dislodging at the inception point most likely produced by the concentration of percolines resulting in an acceleration of the seepage at specific points (Perillo et al., 2005).

In the case of rills, the longitudinal profile shows a maximum depth at the head but the course soon becomes shallower and only a minimum channelization is observed, being minimum at the water edge. The situation changes if several rills discharging groundwater converge into a single course, then course incision could be larger although discharge is seldom large enough to overcome a convex upward profile at the water edge.

Grooves are formed in channel margins of the Bahía Blanca Estuary (Figure 1a) by groundwater discharge but enhanced by the percoline concentration due to crab burrows (Perillo et al., 2005). As groundwater fills the burrows, it spills out over the burrow border, often producing an indentation on the sill that helps concentrate the flow and generate the groove. As these crabs are an important food source of the White Croaker fish, on occasions the mouth of the burrow is enlarged (generating a crater) as the fish hits the surface when it traps a crab resulting in a greater groundwater concentration and larger spill over and, consequently, wider and deeper grooves. In this case, groundwater comes from the tidal filling of dense concentration of crab burrows at the associated marsh platform.

While the various methods described are probably the basic mechanism for the initial development of courses on the lower tidal flats (hydroperiod = 360 days/year), where tidal penetration occurs frequently, it is probably not the dominant mechanism of channel formation within salt marshes or mangroves at higher elevations. Higher (and older) areas may have reached an elevation that limits tidal penetration to only the highest tides (hydroperiod <120 days/year). As a consequence, sheet flows are rarer.

For the case of courses in marshes, it has always been thought that they were inherited (and further modified) from the older tidal flats that they colonized. Some authors (Pethick, 1969, 1992; Pestrong, 1972; Pye, 1992) even suggested that channels were obliterated, rather than created in marshes. Yapp et al. (1917, in Allen, 2000) described the possible mechanism for the origin of courses in marshes based on the presence of hummocks formed by sediment retention due to the colonization of *Glyceria* (Puccinellia). Although Yapp et al. (1917) offered a valid explanation, it does not account for large gullies, creeks and channels. Perillo et al. (1996) proposed the first known mechanism for creek formation on a middle salt marsh in southern Argentina. Creeks develop from the interconnection of series of ponds enlarged by wave erosion of the pond walls as a consequence of the intense, direction-concentrated winds prevailing on Patagonia. Precreeks are finally connected to existing channels by cascading and seaward erosion as tides flood over the ponds. Further growth and maintenance of the creek depends on the intensity of the water exchange. If this exchange is poor, the reduction of soil salinity allows plant colonization which further reduces the exchange, enhancing sedimentation, and finally attaining the complete obliteration of the creek.

A similar mechanism for gully and creek initiation was described for a freshwater marsh on the southern coast of the Rio de la Plata Estuary (Perillo and Iribarne, 2003b). In this case, the ponds are much smaller (Figure 6a and b) and formed by soil marsh depression both by organic matter compaction and by groundwater washing of interlayered sands. Interconnection between ponds and creek formation (Figure 6c) is produced by wall erosion by waves, water cascading and seaward

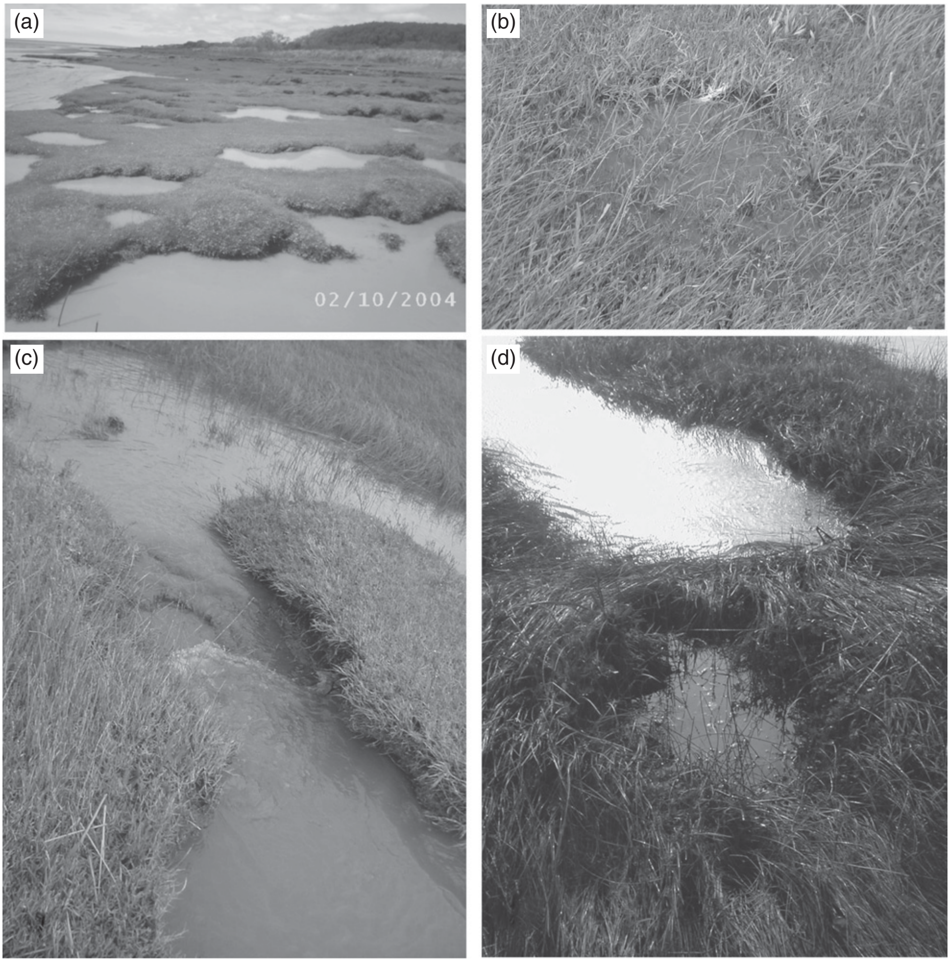

Figure 6 Formation of tidal creeks in a freshwater marsh. (a) View of the field of ponds along the southern coast of the Rio de la Plata, (b) view of a pond showing the variety of plants inside and outside the pond and (c) a threshold between two ponds. Cutting of the threshold interconnects the ponds as a step for the formation of a creek, (d) creek developed after the cutting of the threshold.

retreat. Once the ponds are connected among themselves and with the estuary, water flows through the creek driven by the microtide dominant in the estuary and also by storm surges that are frequent in the area.

Although there is no mention in the literature of processes similar to those described in the previous paragraphs, there are examples observed by the author in other marshes, that is, Minas Basin (Figure 6d) that lead to the formation of creeks by pond connection. This idea is exactly the contrary of the development of channel pans proposed by Pethick (1974, 1992) where changes in the marsh conditions (i.e., lowering of the sea level, marsh disconnection, etc.) result in a smaller hydroperiod followed by vegetation growth and lateral erosion which may isolate parts of the course, specially those with meanders that become ponds.

A fourth mechanism of creek and gully erosion found and extensively studied due to the interesting physical–biological interaction was observed in the Bahía Blanca Estuary (Perillo and Iribarne, 2003a,b). One of the most surprising and interesting findings is that creek formation can actually be a product of the intense action of crabs (*Neohelice granulatus*). In these settings, crabs first interact with a halophytic plant (*Sarcocornia perennis*) developing zones of high density of crab holes, which are then utilized by groundwater and tidal action to form tidal courses. When analyzing the spatial distribution of *Sarcocornia*, Perillo and Iribarne (2003a) discovered that the plants were mostly distributed in circles of up to 1.5–8 m in diameter, with the plants concentrated in a ring along the outer portion of the circle (Figure 7a). These rings vary in width from 0.5 to 1.5 m. The central part of the circle is an unvegetated salt pan, but it is densely excavated by the burrowing crab. Burrow density reaches from 40 to 60 holes/m^2 (Figure 7b). The holes made by the crabs have a diameter of up to 12 cm (Figure 7c), and they reach up to 70–100 cm into the sediment (Iribarne et al., 1997; Bortolus and Iribarne, 1999). An interesting feature of the plant rings and their interaction with the crabs is the effect on the erosive process of the salt marsh in an estuary which is generally in an erosional stage. The formation of the *Sarcocornia* rings and the crab activity plays a major role in the erosion of the marsh transferring 1,380 m^3 of sediment from a 270 ha marsh to the estuary (Minkoff et al., 2005, 2006; Escapa et al., 2007). In 2004 alone, more than 13% (183 m^3) of the total sediment was exported indicating that the process keeps exporting more sediment every year in a pronounced exponential curve.

The fact that the crab burrows are permanently inundated produces two effects. First, silty clay sediments (characteristic of this area) become partly loose and second, the water in the holes starts to migrate, breaking intercave walls and developing a groundwater stream that undermines soil. Under the pan surface, caves are formed where groundwater flows can be seen. A further stage in the development of the channel appears when soil surface fails. At this stage, the course has only a general structure filled with remnants of intercave walls (Figure 7c and d), around which water circulates (Perillo and Iribarne, 2003b). The final stage is reached when the walls are eroded after a number of tides and the channel presents smooth, low-slope banks (Figure 7d). Crabs can be found along the creek banks but no plants are observed. As a result, along the creeks, the original tidal flat moves landward at the expense of the salt marsh, and this mechanism has been described in the previous section. Application of a Cellular Automata model (Minkoff et al., 2006) that takes into account the various active interactions existing in the marsh even makes it possible to predict the changes in the geomorphology of the terrain and the formation and evolution of gullies and creeks.

At the present time, a new mechanism is being studied in the Bahía Blanca Estuary related to the development of shallow ponds (Figure 7e) over tidal flats which may either be or not be connected to creeks. These ponds may result from a complex interaction originated in the differential resistance that biofilms may produce to the flat erosion by short-period wind waves.

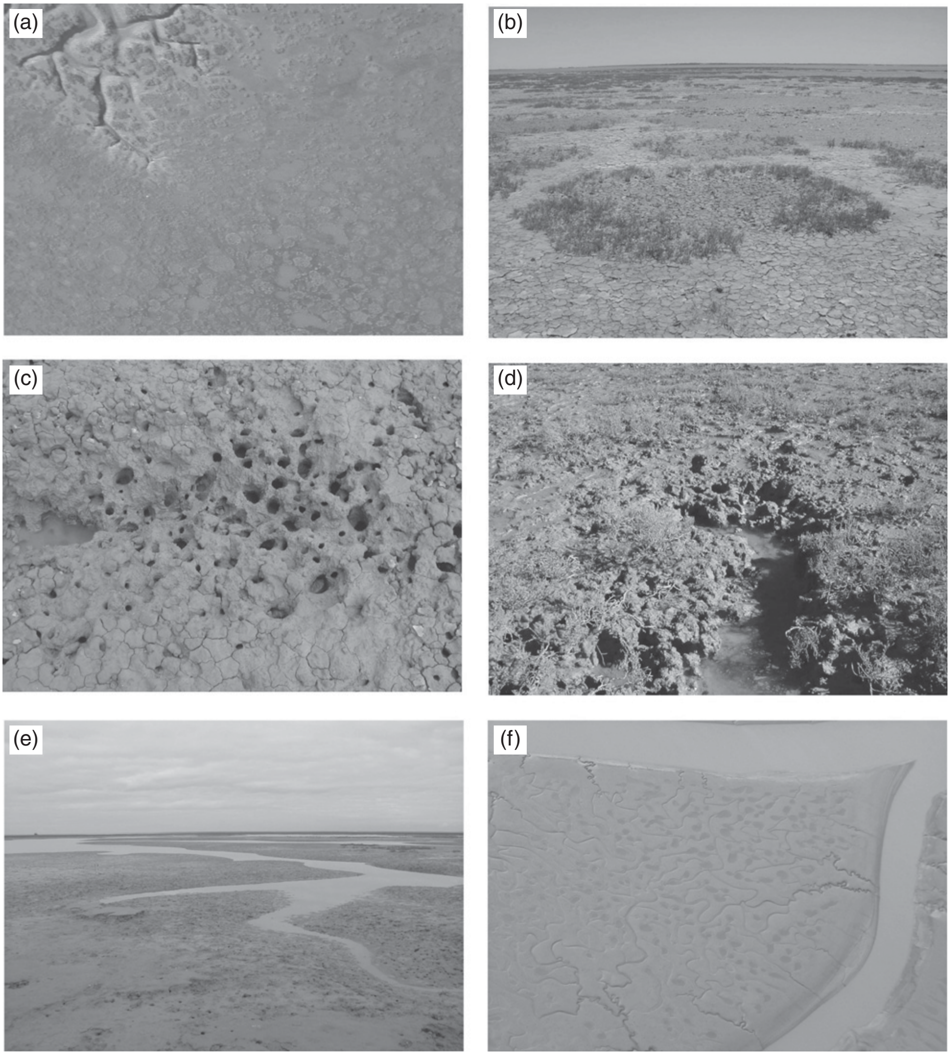

Figure 7 (a) Aerial view of the *Sarcocornia* rings, (b) an example of a *Sarcocornia* ring showing the dense distribution of crab burrows at the center, (c) head of a creek previous to the collapse of the surface soil where the large number of burrows mark the area that will be collapsed, (d) head of a creek after the collapse, (e) creek partially flooded and (f) aerial view of the distribution of small ponds over a tidal flat of the Bahía Blanca Estuary.

6. Course Evolution

The morphology of large rivers depends on many factors (i.e., tectonics, rock/sediment composition, slope, etc.). As size diminishes, local morphology plays an increasingly larger role in the valley and specifically in river channels.

Tidal streams have similar influences. For instance, rills are controlled by surface sediment characteristics; the less compacted the material the deeper the rill. Also slope is important since rills are more lineal with higher slopes. In most cases, rills are deeper near their heads than at their mouth.

In an evolving wetland, courses appear almost simultaneously with the deposition of sediments that forms a tidal flat (Wolanski, 2007). Rills, grooves and gullies are the first to appear and their persistence depends mostly on local factors such as some particular irregularity on the sediment surface or the depth of the incision. A deeper incision tends to occur in spring tides as low water level induces more erosion than neap low water. Once a course has been preserved through several tidal cycles, the possibility of it being erased by strong currents or wave activity diminishes proportionally with time. Thus, further evolution is now controlled by ebb and flood currents, the former being particularly important as they convey the water discharge from the overflat flooding.

Although tidal flats can be preserved, soil stability commonly helps in establishing pioneer plants which, in a longer run, allows the formation of salt marshes and/or mangroves. Levees tend to form along courses where sedimentation is predominant even in the case of tidal flats, being most common in marsh courses. However, they seldom occur along courses in erosive environments or when the concentration of suspended sediments is very low (<100 mg/l). The formation of levees further stabilizes the courses and allows their vertical growth.

When there is no levee dissection, after some extraordinary tides or rainfalls, levees may act as dams allowing the formation of ponds even if there are no significant depressions. Ponds, when persistent in places, can be major controllers of the geomorphologic evolution of an intertidal area. Ponds are affected by evaporation and more prone to be colonized by benthic fauna (i.e., crabs, poliquetes, etc.), which prefers areas with higher water content as it is easier for them to make their burrows further deepening the pond by sediment subsidence. Also diatoms tend to concentrate there, and especially during warmer periods, their oxygen productivity can be large enough to dislodge sediment and keeps it floating (Figure 8). Wind action is a major factor in enhancing ponds although this process has been rather overlooked in the literature. Typical short-period waves form due to small fetch and depth have high steepness which makes them a very erosive process. In places where some wind directions are very frequent (i.e., Argentine Patagonia), ponds can be enlarged along the wind path. Wind, in these cases, could play an important role in defining water circulation over the tidal flat (the effect on marshes and mangroves should be smaller) and because of the relatively low water depth and velocities, wind shear transfer may easily change the pressure gradient induced by the tide or, at the beginning and end of the inundation time, by the topographic gradient. Creeks can be easily formed by the interplay of previous factors as described by Perillo et al. (1996).

Levees are dissected by creeks and gullies but also by strong rain events and short-period, highly erosive waves. These processes help restore the exchange. Therefore, when considered as a whole, levees cannot grow indefinitely but must reach some kind of dynamic equilibrium based on tidal range, overflow velocities, suspended sediment concentration, plant species (involving all plant morphological characteristics affecting flow and sediment stability) and local geomorphology.

Figure 8 Example of the mechanism in which diatoms, by strong oxygen production, can dislodge sediment on a tidal flat.

Moving further inland, courses are also elongated by the interaction with freshwater. Fine sediments react in a different way in freshwater than in salt water. In the former, mud tends to remain in suspension longer than in saltwater conditions, thus allowing better conditions for sediment erosion (Wolanski, personal communication). Several examples in Australia (Mulrennan and Woodroffe, 1998; Cobb et al., 2007) show that freshwater affects saline water mud and, during continental floods, erosion of the salt marshes and flats is larger than during regular tidal inundation (Figure 9). Freshwater is not only introduced into the system by

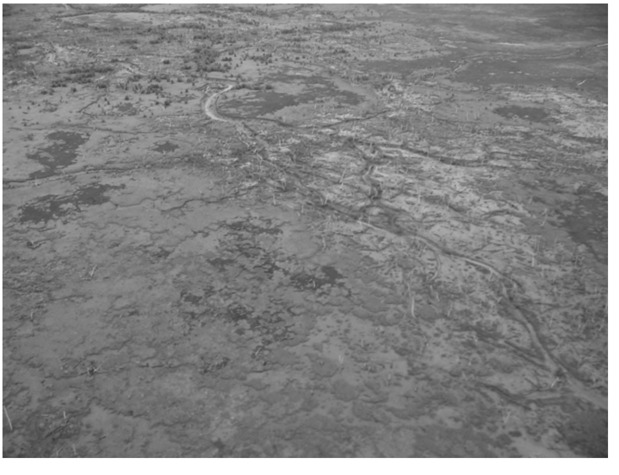

Figure 9 Aerial photograph showing the growth of saline tidal courses in the freshwater plains of the Mary River (Australia), seriously affecting the freshwater vegetation (courtesy of Eric Wolanski).

continental drainage but also in the form of rain which, when intense, also produces strong surface sediment erosion (Mwamba and Torres, 2002) which when conducted by the geomorphology can easily erode the sediment surface further and generate or prolong tidal courses. Then, further evolution of these features becomes a tidal process.

7. Summary

Tidal courses are an essential part of coastal wetlands as they play a major role in water and nutrient exchange. However, their origin and evolution is still a matter of discussion due to the complexities of the dynamic processes associated with their initiation. Some factors such as the role of overland flows versus bankfull flow in the evolution of the courses are unknown. A major question is what actually controls course meandering on tidal flats. Contradictory examples about the meandering distribution can be given when the cases of the Anse d'Aiguillion (Eisma, 1997) and Bahía Blanca Estuary are compared. In the former, creeks in the higher flats are strongly meandering and become straight and sinuous in the low flats, whereas in Bahía Blanca Estuary, what occurs is exactly the opposite; sinuosity increases significantly for creeks and gullies along the margins of tidal channel while on the surface of the flats they are much less sinuous.

A tidal course classification has been proposed to establish a unified description of these features and to avoid confusion. Based on geomorphologic information, this classification may evolve further by integrating other descriptors. Therefore, this classification is open to consideration and discussion.

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