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**Point-bar brink and channel thalweg trajectories depicting interaction
between vertical and lateral shifts of microtidal channels in the Venice Lagoon
(Italy)**

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ABSTRACT

Tidal point bars are generally described as laterally accreting bodies, generated by lateral shift of meander bends, in which the point-bar brink (i.e. the break between bar top and bar slope) and the channel thalweg (i.e. the deepest part of the channel) shift horizontally toward the outer bank. The present study applies the concept of trajectory analysis at the point bar scale, focusing on the trajectories of point-bar brink and channel thalweg, in order to understand how vertical aggradation can interact with lateral migration to shape geometries of tidal point bars developed in a microtidal and highly aggrading salt marsh setting. We selected eight study-case meander bends, located in the Venice Lagoon and characterized by different point-bar morphologies, whose widths and depths range from 2 to 11 m and from 0.5 to 1.6 m, respectively. All the point bars were investigated through a high resolution facies-analysis carried out on closely-spaced sediment cores, collected along the bar axis. Location of bar brink and channel thalweg at different times defined specific trajectories, which were classified either as ascending or descending, and linear or non-linear. All brink trajectories are ascending, and show evidence of lateral shift of the bar brink under aggradational conditions of surrounding marshes. Development of non-linear brink trajectories is linked with changes in the ratio between vertical and lateral shift rates of the brink, which is in turn dictated by changes in local base level due to substrate compaction. Conversely, the thalweg trajectories can be either ascending or descending, reflecting an interaction between rates of lateral shift and aggradation/degradation of the channel floor. The brink and thalweg can either shift consistently (e.g., both trajectories are ascending) or incongruously (e.g., ascending brink vs. descending thalweg trajectory), reflecting different attitudes of the channel to maintain or increase its cross-sectional area.

Keywords: tidal point bar, aggradation, channel thalweg, lateral migration

1. INTRODUCTION

Trajectory analysis focuses on the migration pattern of easily recognizable geomorphological features (e.g., breaks-in-slope) and associated sedimentary environments through time (cf. Helland-Hansen and Hampson, 2009). This approach is suitable to be applied to a large variety of scales, which can range from bedforms (cf. “climb angle” by Larue and Martinez, 1989) to continental margins (Henriksen et al., 2011). In addition, trajectory analysis has been widely employed at the shoreline or shelf-edge scale, shedding light on the distribution of different depositional systems through time in response to combined variations in rates of relative sea level changes and sediment supply (Henriksen et al., 2009, and reference therein). On the contrary, less attention has been paid at the scale of depositional elements, like deltas (Gobo et al., 2015) or fluvial bars (Ghinassi et al., 2014).

Point bars generated by fluvial and tidal meandering channels are mostly described as laterally accreting bodies (Bridges and Leeder, 1976; Colombero et al., 2017; D’Alpaos et al., 2017; Durkin et al., 2015; Fisk, 1944; Jackson, 1976; McGowen and Garner, 1970; Miall, 1985; Smith, 1987) with a tabular geometry. This geometry is generated by lateral shift of the inner channel bank, which implies a synchronous horizontal shift (Fig. 1A) of both the bar brink (i.e. the morphological break between bar top and bar slope) and the channel thalweg (i.e. the deepest part of the channel), which, in a cross section parallel to the bend axis, defines two parallel and horizontal trajectories (Fig 1A). Although the common occurrence of tabular point-bar bodies in

the stratigraphic record (Pranter et al., 2007; Puigdefabregas, 1973; Puigdefabregas and Van Vliet, 1977) supports this model, documentations of stable to slow-migrating channels in high-aggradational settings (Candel et al., 2017; Ielpi et al., 2015; Makaske, 2001; Nanson and Croke, 1992) suggest that a high ratio between vertical and lateral accretion rate is likely to promote a mixed, latero-vertical shift of meandering channel systems.

Fluvial meanders migrate laterally with a mean rate of meters per year (Bridge et al., 1995; Hudson and Kesel, 2000; Lagasse et al., 2004), although this rate can vary from being undetectable over decades (Beeson and Doyle, 1995; Candel et al., 2017; Hudson and Kesel, 2000) to reach values as high as tens of meters during a single flood event (Ghinassi et al., 2018b; Moody and Meade, 2014). The interaction between rate of vertical and lateral shift of fluvial meanders has been rarely documented in the stratigraphic record (Ghinassi et al., 2014; Ielpi et al., 2015; Rajchl and Uličný, 2005), and numerical models (van de Lageweg et al., 2016; Willis and Tang, 2010) show that only high overbank aggradation rates (i.e. cm/yr to dm/yr) might possibly affect geometries of point bar bodies. Fluvial meanders developed by high sinuosity, low-energy streams in peatlands of the Netherlands (Candel et al., 2017) represent an end-member of interaction between rate of vertical and lateral shift of meandering channels. In this setting, geometries of bar bodies are prominently shaped by vertical accretion through processes of oblique aggradation (cf. Candel et al., 2017), which cause the bar brink and channel thalweg to shift synchronously following steep to vertical trajectories. Rajchl and Uličný (2005) analyzed the influence of substrate compaction on geometries of point bar bodies recorded in the Neogene Hrabák fluvial system in the Most Basin (Czech Republic).

Figure 1 Models for point bar geometries. (A) Idealized model showing stratal geometries generated by lateral shift of the point bar brink and channel thalweg. **(B)** Lateral and vertical shift of a fluvial point bar of the Neogene Hrabák system, Czech Republic (modified after Rajchl and Uličný, 2005). **(C)** Channel thalweg migration under aggradational conditions in a modern tidal meander bend of the Drum Bay, Scotland

(modified after De Mowbray, 1983). (D) Brink and thalweg trajectories of a modern tidal point bar in the Venice Lagoon (modified from Brivio et al., (2016). (E) Channel thalweg trajectory under the effect of a changing tidal prism in a Mid-Holocene channel, Netherlands (modified from Rieu et al., (2005)

They showed that syndepositional compaction of peat deposits, due to sediment loading of the overlying growing point bars, leads to a local increase in subsidence, which causes development of steep brink and channel thalweg trajectories (Fig 1B).

Similarly to their fluvial counterparts, tidal point bars were commonly described (Barwis, 1978; Bridges and Leeder, 1976) as tabular sedimentary bodies generated by horizontal and synchronous shift of bar brink and channel thalweg. Nevertheless, as demonstrated for fluvial systems, lateral migration of tidal meanders can follow different styles, and the trajectories defined by both point bar brink and channel thalweg are not necessarily horizontal nor parallel. As a matter of example, De Mowbray (1983) showed that the channel thalweg of a tidal meander bend in the Drum Bay (Scotland) did not merely shifted horizontally, but migrated under gradual aggradation (Fig 1C) in order to keep pace with the build-up of the adjacent tidal flats. A similar behavior was also documented by Brivio et al. (2016) for the brink of a modern tidal point bar in the Venice Lagoon (Italy), where a combination between channel lateral shift and aggradation of the surrounding saltmarshes caused the bar brink to rise during the channel lateral shift (Fig 1D). Finally, using high-resolutions seismic data from western Netherlands, Rieu et al. (2005) described a lateral shift of a mid-Holocene tidal channel under the effect of varying tidal prism. In this case, the shift of the channel thalweg defined a descending and ascending trajectory during phases of increase and decrease in the tidal prism, respectively (Fig 1E).

It emerges that, during lateral migration of tidal channels, the shift of the bar brink and the channel thalweg may not be horizontal and in phase, being their behaviour

dictated by the interaction between different factors, which include rate of channel lateral shift (e.g. D'Alpaos et al., 2017; Fagherazzi et al., 2004; Finotello et al., 2018), rate of vertical aggradation of adjacent unchanneled areas (e.g. D'Alpaos et al., 2007), local variations of relative sea level and the related increase/decrease in tidal prism (e.g. D'Alpaos et al., 2010). The combinations between different types of bar brink and channel thalweg trajectories, along with the related changes of point bar thickness and geometry, can provide relevant insight on the identification of the depositional boundary conditions under which bend migration occurred. Therefore, describing different bar brink and thalweg trajectories, as well as of their possible combinations, is a critical step towards an improved understanding of the morphodynamic evolution of tidal meanders and related sedimentary products. These issues are mostly relevant in tidal landscapes, where highly aggradational (i.e. cm/yr) marshes host a complex network of slowly-migrating (i.e. cm to dm/yr) meandering channels (Allen, 2000; Fagherazzi et al., 2004; Gabet, 1998; Garofalo, 1980; Hughes, 2012; Marani et al., 2002), whose sedimentary and architectural features are still relatively unexplored (Boaga et al., 2018; D'Alpaos et al., 2017; Ghinassi et al., 2018a). The present work focuses on meander bends of the Venice Lagoon providing the first description of different styles of bar brink and channel thalweg trajectories in order to improve our understanding of tidal point-bar evolution and architecture. Specifically, this study aims at i) illustrating geometries of tidal point bars along 2D axial cross sections; and ii) defining the behavior of the bar brink and channel thalweg where point bar bodies accrete under aggradational conditions.

2. THE VENICE LAGOON AND THE STUDY SITES

The Venice Lagoon is located along the Northeastern coast of Italy, it has a total surface of about 550 km² and represents the largest brackish water body of the Mediterranean Basin (Fig 2A). The Lagoon is characterized by an elongated shape trending NE-SW and by a mean water depth of tidal flat and subtidal platform of about 1.5 m. It is connected to the Adriatic Sea through three inlets: Lido, Malamocco, and Chioggia (Fig 2A). The tidal regime is semidiurnal with an average range of about 1.0 m. The maximum water excursion at the inlets is ± 0.75 m around Mean Sea Level (MSL) (D'Alpaos et al., 2013) which can be increased by meteorological forcing (Carniello et al., 2016).

Figure 2 Study sites. (A) Location of the Venice Lagoon along the Northeastern coast of Italy. (B and C) Location of the study sites in the Northern (N1, N2, N3, and N4) and Southern (S1, S2, S3, and S4) Venice Lagoon, respectively.

The Venice Lagoon is part of a wider foreland basin located between the Apennine and the South Alpine chains (Massari et al., 2009). The Quaternary infill of the basin, in the Venice area, consists of a shallowing upward trend from deep marine hemipelagic mud and turbiditic succession to deltaic and shoreface deposits

(Massari et al., 2004), which is followed by a cyclic alternation of continental to shallow marine facies deposited under glacio-eustatic control in the uppermost succession (Kent et al., 2002). The last recorded cycle consists of the alluvial sediments of the Brenta River Megafan, developed in the area during the Last Glacial Maximum (Fontana et al., 2014), and of the lagoonal deposits related to the Holocene transgression, which prompted paralic deposition and the development of the Venice Lagoon (Amorosi et al., 2008; Zecchin et al., 2014, 2008). During the last five centuries, human interventions have strongly altered the natural evolution of the Lagoon, preventing its siltation (D'Alpaos, 2010; Gatto and Carbognin, 1981), and causing a human-induced transgressive phase (Zecchin et al., 2014, 2009, 2008), which is associated with a significant reduction in salt marsh surface (D'Alpaos, 2010; Day et al., 1998; Marani et al., 2007; Roner et al., 2016).

Nowadays, the great majority of salt marshes are found in the Southwestern and Northwestern areas of the Lagoon. Salt marshes located in the Northern Venice Lagoon have been characterized by accretion rates ranging between 0.1 and 0.5 cm/yr during the last century, with the highest values observed after the 1960s (Bellucci et al., 2007). Conversely, the accretion rates of salt marshes located in the Southern Venice Lagoon ranged between 0.1 and 1.2 cm/yr since the 14th century, when palustrine sedimentation ceased and salt marshes colonization began (Roner et al., 2017).

The present study focuses on eight different meander bends, four of which (N1 –N4, Fig 2B) cut through the San Felice salt marshes in the Northern Lagoon, whereas the others are located in the Southwestern marshes of Punta Cane (S1 – S4, Fig 2C).

Aggradation and lateral migration in tidal point bars

Figure 3 Satellite images (from Google™Earth) showing the study meander bends in the Northern (N1, N2, N3, and N4) and Southern (S1, S2, S3, and S4) Lagoon. Dots indicate the position of the recovered cores in each meander.

The study case bends are 2 – 11 m wide and 0.5 –1.6 m deep (Fig 3), and cut through salt marshes colonized by a dense halophytic vegetation species such as *Limonium narbonense*, *Spartina maritima*, *Sarcocornia fruticosa*, *Juncus maritimus*, and *Salicornia veneta*. All the study channel bends are still active, made exception for that at site N4 (Fig 3), which is now completely filled following a cutoff occurred during the 1950s (D’Alpaos et al., 2017). None of the active channels is completely dried out at the lowest tides, and point bar top can be covered by few decimeters of water at the highest tides. Meanders at sites N2 and S2 show a “simple asymmetric” (*sensu* Brice, 1974) planform geometry, whereas all the other channels can be labelled as “simple symmetric” (*sensu* Brice, 1974). The radius of curvature of the bends ranges between 6 and 21 m (Fig 3).

3. METHODS AND TERMINOLOGY

A total of 50 cores were recovered along axial transects of the study bends. Each transect extended from the channel thalweg to the line connecting inflection points of the bend (Fig 3). The position and elevation of the cores was determined using differential GPS TOPCON GR-3 receivers – dual frequency (L1/L2) and dual-constellation (NavStar/Glonass) with integrated Tx/Rx UHF radio. Cores were up to 3 m deep and were recovered using an Eijkelkamp hand auger, through a gouge sampler with a length of 1 m and a diameter of 30 mm, which prevented sediment compaction. PVC liners were used to keep the cores humid, which were successively cut longitudinally, photographed and logged. Sedimentological

analyses were carried out on the study cores following the basic principles of facies analyses. Identification of different types of deposits (Fig 5 and 6) was based on the integration between core location and related sedimentary features, including sediment grain size and color, presence of sedimentary structures, vertical grain-size trends, degree of bioturbation and occurrence of plant and/or shell remains. Correlation among cores allowed to disclose 2D point bar sections along the bend axis (Fig 7 and 8).

Figure 4 Terminology used in the present work. (A) Key terms used to define different elements of tidal meander bends. (B) Examples of point bar brink (left) and channel thalweg (right) deposits in sedimentary cores. (C) Tidal point bar architecture and definition of point bar brink and thalweg trajectories. s_t : maximum thickness of bar top, salt-marsh deposits accumulated during bar migration. d : distance covered by the lateral shift of the channel.

Terminology used in the present work is reported in figure 4, and follows that used for fluvial point bars (cf. Ielpi and Ghinassi, 2014), although some modifications have been introduced due to the bidirectional nature of tidal currents (e.g. seaward/landward instead of downstream/upstream). The point bar brink is defined here as the bar rim (Fig 1A and 4C), characterized by a break in the angle of deposition from the flat bar top deposits to the inclined bar slope. Accordingly, in cores, the point bar brink was pointed out where lamination changed from horizontal (i.e. bar top deposits) to inclined (i.e. bar slope deposits). This change in inclination occurs within an interval spanning between 5 and 10 cm in thickness (Fig 4B), which represents the uncertainty for detecting the brink position in the study deposits. The channel thalweg is defined as the deepest part of the active channel (Fig 1A and 4C). In cores, it corresponds to the surface flooring channel lag deposits and can be precisely located (Fig 4B).

Minimum and maximum lateral migration rates (ζ_{min} , ζ_{max} , *respectively*) of each bend have been approximately estimated on the basis of the ratio between the distance covered by the lateral shift of the channel (d in figure 4C) and the minimum and maximum estimated time span over which migration occurred (Δt_{min} and Δt_{max}). The latter have been determined as $\Delta t_{min}=s_t/s_{a,min}$ *and* $\Delta t_{max}=s_t/s_{a,max}$, where s_t is the maximum thickness of bar top salt-marsh deposits accumulated during bar migration (Fig 4C) and $s_{a,min}$ and $s_{a,max}$ are the minimum and maximum vertical accretion rates of salt marshes reported in literature for the study area (Bellucci et al., 2007; Roner et al., 2017).

4. RESULTS

4.1 Study deposits

4.1.1 Substrate deposits

These deposits represent the lowermost stratigraphic unit of each study case, and are radically different in the Northern and Southern Venice Lagoon.

In the Northern Lagoon, they consist of an alternation between sandy and muddy layers. The sandy layers, that can locally exceed 1 m in thickness, consist of well-sorted dark gray sand, medium to very fine in grain size, with abundant shells, shell fragments and plant debris (Fig 5B). Sandy deposits can locally contain millimetric muddy laminae. The muddy layers consist of dark, organic-rich mud with scattered shells and plant fragments, and are up to 10 cm thick (Fig 5A). Bioturbation is common in both the sandy and muddy layers, and prevents detection of sedimentary structures. Substrate deposits of the Northern Lagoon are interpreted to be formed in a tidal flat/subtidal platform environment. Sandy deposits were likely originated during wind-induced storm events, when wave winnowing entrained fine-grained sediments and sand concentrated on the lagoon floor (Carniello et al., 2009). Muddy deposits were settled down from suspension during the waning stage of storm events.

In the Southern Lagoon, substrate deposits consist of peat with abundant fragments of reeds (Fig 5D). In some of the cores from sites S2 and S3, a dark-gray mud interval with sparse reed fragments (Fig 5C) occurs on top of peat deposits. Peat consists of comminuted dark brown to black plant debris with a minimum amount of dispersed mud. The presence of peat deposits and abundant reeds is consistent with a wetland setting with freshwater input (Bondesan and Meneghel, 2004; Tosi et al., 2007). The mud on top of peat deposits could be related to local protected ponds, where vegetation cannot grow and mud can settled down (Roner et al., 2016; Silvestri et al., 2005). Although peat is widespread in the Southern Lagoon (Roner et

al., 2017), substrate deposits at site S4 (Fig 2) consist of an alternation between mud and sand similar to that occurring in the Northern Lagoon.

Figure 5 Substrate deposits of the Northern (A and B) and Southern (C and D) Lagoon. (A) Organic-rich mud with shells and plant fragments; (B) Dark-gray sand, medium to very fine in grain size, with abundant shells, shell fragments and plant debris. (C) dark-gray mud with scattered reed fragments; (D) Peat with abundant fragments of reeds.

4.1.2 Point bar and channel lag deposits

These deposits are up to 2.25 m thick, and show a lateral extent of 10-20 m, and 3-15 m for the Northern and Southern sites, respectively (Fig 9). These deposits pinch out toward both the terminations of all the study transects (Fig 8). They are erosively-based and grade upward into salt marsh or channel-fill deposits. The basal erosional surface is draped by a shell-rich, 5-20 cm thick massive layer (Fig 6 A.7 and B.5), which consists of fine to medium sand and of silt to fine sand in the Northern and Southern sites, respectively. Shells are commonly fragmented and include both gastropods (e.g. *Bittium scabrum*, *Tritia neritea*, *Gibbula* sp.) and bivalves (e.g. *Loripes orbiculatus*, *Acanthocardia tuberculata*, *Scrobicularia plana*). Plant debris and pebble-sized, rounded mud clasts are also common (Fig 6 A.6).

298 The shell-rich layer is covered by clinostratified deposits, which are up to 2.2 m thick
 299 and show an overall fining-upward grain size trend (Fig 6 A and B). Clinostratified
 300 deposits consist of two intervals. The lower interval is made of fine sand to silt-rich
 301 mud with scattered shell fragments and mud clasts. These deposits are mainly
 302 massive (Fig 6 A.4), although a local inclined lamination is slightly visible (Fig 6 B.3).
 303 Bioturbation is really common and often prevent the identification of primary bedding
 304 (Fig 6 A.5). The upper interval ranges in grain size between sandy silt and mud and
 305 shows a clear inclined heterolithic stratification, with laminae dipping channelward
 306 between 5° to 25° (Fig 6 A, A.3, B and B.2). Laminae consist of well-sorted very-fine
 307 to fine sand (Fig 6 A.3) or comminuted plant debris (Fig 6 B.2), which are more
 308 common in the Northern and Southern sites, respectively. Inclination of laminae
 309 decreases moving upward in the clinostratified deposits, and becomes insignificant
 310 at the boundary with the overlying salt-marsh deposits (Fig 6 A, A.3 and B).

311 The shell-rich basal layer and the overlying clinostratified silty sand are interpreted
 312 as channel lag and point bar deposits, respectively. Channel lag deposits
 313 accumulated in the deepest part of the channel, where the coarser sediments and
 314 shell fragments concentrate, as finer grains are entrained by currents (Terwindt,
 315 1988). Shelly assemblages consist of mainly marine and subordinately brackish
 316 molluscan species that roughly reflect the benthic fauna currently inhabiting mud and
 317 sandy substrates of the Venice Lagoon. Pebble-sized mud clasts are the product of
 318 fragmentation of blocks collapsed from the channel banks (Terwindt, 1988).

319 Clinostratified bar deposits accumulated as a consequence of lateral shift of the
 320 channel, and their fining-upward grain size trend, reflect the overall decrease of the
 321 bottom shear stress moving from channel thalweg to bar top in the axial zone of a
 322 meander bend (Dietrich et al., 1979; Dietrich and Smith, 1984; Frothingham and

323 Rhoads, 2003; Hooke, 1975). In the lower bar deposits (i.e. lower part of
324 clinostratified sediments), intense bioturbation, and related paucity of sedimentary
325 structures, is consistent with the permanent presence of water on this part of the bar.
326 In the upper bar deposits (i.e. upper part of clinostratified sediments), the dominance
327 of mud points out to prevalence of fallout processes, which were probably dominant
328 during prolonged slack water periods or after storm events (Carniello et al., 2011).
329 Inclined sandy laminae of the upper bar were produced during storms at high tides,
330 when both salt marshes and bar tops were winnowed by waves (Carniello et al.,
331 2011; Choi and Jo, 2015; Fruergaard et al., 2011; Green and Coco, 2007). The
332 decrease in the inclination of laminae reflects the progressive upward flattening of
333 bar slope and heralds the transition into overlying salt-marsh deposits.

Heterolithic bar deposits consisting of silt with inclined sandy laminae. Note the upward decrease in dip angle of sandy laminae; (A.4) and (A.5) Structureless sand and bioturbation in the lower part of the point bar; (A.6) Pebble-sized mudclast in massive channel-lag sand; (A.7) Shell-rich channel-lag sand; (B) Southern Case. Example from site S3; (B.1) Salt-marsh mud deposits with roots; (B.2), (B.3) and (B.4) Silty to muddy bar deposits with a variable amount of inclined plant debris and sandy laminae; (B.5) Massive channel-lag sand with shells and shells fragments, flooring the channel base.

4.1.3 Salt marsh deposits

Salt-marsh sediments commonly cover point bar bodies, but they can also overlay substrate deposits. Salt-marsh deposits consist of brownish, oxidized mud with 1-3 mm thick horizontal well-sorted sandy laminae (Fig 6 A.1, A.2 and B.1). In situ roots, wood fragments and bioturbation are common. Along each transect, the thickness of these deposits decreases toward the channel (Fig 8). Maximum thicknesses of salt marsh deposits is 0.7 and 2 m in the Northern and Southern sites, respectively (s_t in figure 9).

Salt-marsh deposits accumulate in the upper part of the intertidal zone, where subaerial exposure is frequent (Silvestri et al., 2005). This is in agreement with widespread oxidation and occurrence of abundant roots. These deposits are mainly accreted through mud settling and organic matter accumulation (Allen, 2000; Mudd et al., 2010; Roner et al., 2016). Muddy sediments are deposited by fallout during high water slacks, at the transition between flood and ebb tide. Sorting of sand forming thin laminae suggests their formation during storm events, when salt-marsh surface was flooded and wave-winnowing suspended mud particles and concentrated coarser sediments (Choi and Jo, 2015).

4.1.4 Channel-fill deposits

These deposits occur only at sites N2, N4 and S1 (Fig 8), are up to 1.5 m thick and overlie point-bar and channel-lag sediments. At the abandoned channel bend of site N4 they are thick as the adjacent point bar, whereas at sites N2 and S1 they are thinner than the bar. These deposits consist of dark gray, massive mud with dispersed bivalves in life position. At sites N4 and S1 (Fig 8), the transition between these deposits and underlying point bar sand is transitional.

These muddy deposits settled down from suspension, silting up channels affected by a progressive decrease in water discharge. In case of site N4, deactivation of the channel followed a neck cutoff, which allowed the whole channel to be filled with mud (D'Alpaos et al., 2017). At sites N2 and S1, the increase of fallout processes can be ascribed to a general decrease in the discharge shaping the tidal network in the area, and the gradual transition from underlying bar deposits points out to a progressive deactivation of the channel (cf. Toonen et al., 2012).

4.2 Point bar brink and channel thalweg trajectories

The stratigraphic relationship between different sedimentary facies marked location of point bar brink (i.e. transition between point-bar and salt-marsh deposits) and thalweg (i.e. base of channel lag deposits) zone in all the study cores. Correlation between adjacent cores depicted the point bar brink and channel thalweg trajectories, which were generated during lateral shift of the channel and accumulation of related point-bar bodies (Fig 7). Brink and thalweg trajectories start from a shared point (i.e. “nucleation point” in figure 7) and follow specific paths, which are summarized in figure 8, where the vertical scale has been doubled in order to emphasize different trends.

Figure 7 Correlation between cores along the axis of point bar N1. Bar brink and channel thalweg trajectories are highlighted.

All the reconstructed bar brink trajectories show an ascending pattern (Fig 8), although some of them show low steepness (less than 2°) and can be considered sub-horizontal. In sites located in the Northern Lagoon, the trajectories appear to be gentle and linear, with a constant slope ranging between 0.3 and 1.7° (Fig 8 N1-N4). In the Southern Lagoon, a linear and gentle (2.9°) trajectory occurs only at site S4, whereas sites S1 – S3 are characterized by curve and steep trajectories. At sites S2 and S3 (Fig 8), the trajectories show a concave-upward profile, with a slope inclination that changes from sub-horizontal to ca. 41.5° and 9.4° , respectively (Fig 8). At site S1 the brink trajectory defines a convex-upward profile, with the slope decreasing from 16.2° to 6.6° .

The thalweg trajectories appear to be more complex and result from the combination of descending, horizontal and ascending shifts. In all the study cases (Fig 8), with the exceptions of the site N2 (where nucleation point has not been detected), the thalweg trajectories show an initial downward shift. The steepness of this descending segment varies from 6.3° to 31.6° and shows a linear or faintly concave-upward geometry (Fig 8). After this first descending reach, the following portion of the trajectories shows either gentle (N1 – N4) or steep (S1 – S4) slopes. In the Northern sites, these trajectories are commonly sub-horizontal with slightly ascending (N1) or descending (N2-N3) shifts. In the Southern cases, they rise up with inclination

410 ranging between 4.8° and 23.3° (Fig 8). None of the ascending reach show convex-
 411 or concave-upward geometries.

412 **Figure 8** Point bar sections of the northern (N1 – N4) and southern (S1 – S4) study cases. Point bar brink and
 413
 414 thalweg trajectories are marked with light blue and red dotted lines, respectively. Slopes of bar brink and
 415 channel thalweg trajectories are shown on the right side of the figure. Vertical exaggeration x2.

Different combinations of bar brink and thalweg trajectories occur at different sites (Fig 11), although all the study cases share an early stage where brink and thalweg trajectories diverge (Fig 7 and 8), causing an increase of point bar thickness. This pattern has not been tested at site N2, since deposits of the early channel evolution were not cored (Fig 8). This early stage of divergence occurs over distances spanning between 1 and 7 m, and the angle of divergence ranges between 7.8° (N3) and 36.5° (S1). After the stage of divergence, different patterns of brink and thalweg trajectories are combined. At Northern sites (N1 – N3), gently ascending brink trajectories are combined with similar thalweg trajectories. At site N4, after an initial similar pattern (ascending brink and thalweg) a phase of divergence of brink and thalweg trajectories is recorded. At Southern sites, the stage of divergence is also followed by a stage in which brink and thalweg trajectories show a similar ascending pattern (S1 – S3). Site S4 does not show this trend and, after the stage of divergence, the thalweg trajectory rises converging toward the corresponding brink trajectory.

4.3 Lateral migration rates

The lateral migration rate of each bend has been estimated through a comparison between the maximum thickness of salt-marsh deposits overlying different bars and aggradation rates documented in the study areas (Bellucci et al., 2007; Roner et al., 2017). Figure 9 shows that maximum migration rates do not exceeds 0.2 m/yr (N2). Minimum migration rates obtained for the Northern and Southern sites are in the range of 0.02-0.04 m/yr and 0.002-0.01 m/yr, respectively; maximum migration rates span between 0.11-0.19 m/yr in the Northern meanders, and between 0.02-0.16 m/yr in the Southern meanders. These values are consistent with other studies carried out

for meandering channels of the Venice Lagoon (Brivio et al., 2016; D’Alpaos et al., 2017; Donnici et al., 2017; Finotello et al., 2018; Ghinassi et al., 2018a; McClennen and Housley, 2006) and in other microtidal environments worldwide (Gabet, 1998; Garofalo, 1980).

ACCRETION RATE ($s_{a,min} - s_{a,max}$)

Northern Venice Lagoon (Bellucci et al., 2007): 0.1 - 0.5 cm/yr

Southern Venice Lagoon (Roner et al., 2017): 0.1 - 1.2 cm/yr

	site	s_t (m)	d (m)	ζ_{min} (m/yr)	ζ_{max} (m/yr)
NORTHERN CASES	N1	0.70	15	0.02	0.11
	N2	0.40	15	0.04	0.19
	N3	0.30	10	0.03	0.16
	N4	0.70	20	0.03	0.14
SOUTHERN CASES	S1	2.00	9	0.005	0.05
	S2	1.65	3	0.002	0.02
	S3	1.10	15	0.01	0.16
	S4	0.47	6	0.01	0.15

Migration Rate (ζ) = $d / \Delta t = (d * s_a) / s_t$

$\Delta t = s_t / s_a$

Figure 9 Estimation of minimum (ζ_{min}) and maximum (ζ_{max}) lateral migration rates of the different meanders considering maximum ($s_{a,max}$) and minimum ($s_{a,min}$) salt marsh accretion rates reported in the literature.

5. DISCUSSION

Trajectory analysis has been applied here at the point-bar scale, showing the migration patterns of the point bar brink and the channel thalweg along axial cross sections of different tidal bends. Trajectories are labelled as ascending, descending and horizontal (Fig 10). Ascending and descending trajectories are distinguished in linear and non-linear (Fig 10). While linear trajectories document a constant ratio between vertical and lateral shift rates ($d(v/\zeta)/dt = 0$), non-linear trajectories, which

457 can be concave- or convex-upward, document temporal changes in the ratio
458 between vertical and lateral shift rates ($d(v/\zeta)/dt \neq 0$).

459

460 Figure 10 Different types of point bar brink and channel thalweg trajectories.

461

462 Figure 11 Combinations of point bar brink and channel thalweg trajectories registered in northern and
463 southern cases.

5.1 Point bar brink trajectories

Ascending point bar brink trajectories point out that the lateral shift of the brink was influenced by aggradation of surrounding salt-marshes, which kept pace with the progressive sea-level rise by accreting both inorganic and organic sediments (D'Alpaos et al., 2007; Morris et al., 2002; Mudd et al., 2010). Steepness of brink trajectories reflects a different ratio between vertical accretion and lateral migration, and, where lateral shift rate dominates over that of vertical aggradation, the brink trajectories will appear sub-horizontal (i.e. less than 2°). Conversely, where vertical aggradation is higher, it forces the brink to define steep trajectories (i.e. up to 41.5°). The remarkable difference in the flat and steep trajectories observed in the Northern and Southern Lagoon, respectively, can hardly be ascribed to the slightly difference between aggradation rates of these two areas (Bellucci et al., 2007; Roner et al., 2017). Therefore, a further constraint is considered to be responsible of this remarkable difference, namely the nature of substrate deposits hosting different bars. Peat deposits, like those hosting S1-S3 bars, are the most compressible of all natural soils (Allen, 1999) and their compaction leads to a substantial increase of local subsidence (Long et al., 2006; Rajchl and Uličný, 2005; Törnqvist et al., 2008; van Asselen, 2011; van Asselen et al., 2010). Peat deposits can be compacted up to 43% within few centuries (van Asselen et al., 2010), with compaction rates that can exceed 10 mm/yr, on decadal to centenary timescales (Törnqvist et al., 2008). The dominant factors influencing peat compaction are the organic matter content and the sediment loading of overlying deposits (Elliott, 1985; van Asselen, 2011; van Asselen et al., 2010). The peat of the Southern Lagoon was progressively compacted by the increasing sediment load due to the growth of the overlying point bars, with a significant increase of local subsidence and related aggradation (Fig 12). The lack of

a steep brink trajectory at bar S4, which covers a sandy substrate, supports this hypothesis. Progressive compaction of peat substrate is also consistent with the occurrence of non-linear and concave-upward trajectories (sites S2 and S3; Fig 8), which point out a progressive increase of the ratio between vertical and lateral shift rate. This increase was due to intensification of local subsidence/aggradation triggered by loading exerted by the growing of bar deposits over peat deposits. The non-linear convex-upward trajectory at site S1 documents a progressive decrease in the ratio between vertical and lateral shift rate. Progressive flattening of this trajectory points out to a relative increase in lateral channel shift, which would be caused by the occurrence of erodible deposits (e.g. sand) along the outer bank of the channel (cf. Ghinassi et al., 2016; Smith et al., 2009). Linear and gently-sloping trajectories of the Northern sites indicate that the ratio between vertical and lateral shift rate remained constant during lateral shift of the channel, and that rate of lateral migration dominated over that of vertical shift. The presence of a sand-prone substrate in the Northern Lagoon nullifies the effects of local substrate compaction, hindering the development of curvilinear trajectories.

sediment loading
 regional subsidence
 total subsidence
 Point bar brink
 Channel thalweg

Figure 12 Diagram illustrating the influence of substrate compaction on developing of bar brink and channel thalweg trajectories. The progressive increase of bar sediment loading on the peaty substrate, causes a local increase in subsidence/accretion, which are documented by development of steep brink trajectories.

5.2 Channel thalweg trajectories

Channel thalweg trajectories include ascending, descending and horizontal trends, indicating that the channel base shifted laterally under aggradational, degradational and stable conditions, respectively. The lack of clear convex- or concave-upward profile of ascending and descending trajectories could point to a stable ratio between lateral and vertical shift rates, although the erosive (i.e. irregular) nature of the channel basal surface could prevent a clear detection of these trends.

The descending thalweg trajectories which characterize the early part of all the study trajectories, document the establishment of the channels, which shift laterally and cut down into the substrate in order to reach a cross sectional depth that is in equilibrium with the local, formative tidal prism. The formative depth in the bend axial zone is function of the channel discharge and bend geometry. Channel discharge is related to the tidal watershed area, and can vary over temporal scales of decades to centuries as consequence of changes in the tidal prism, watershed area (e.g. due to channel piracies and/or meander cutoff) and channel network evolution (Allen,

2000; D'Alpaos et al., 2005; Dalrymple et al., 1991; Garofalo, 1980; Lanzoni and Seminara, 2002; Stefanon et al., 2012). Bend geometry can influence the capability of the channel to cut through its substrate (Crosato, 2009; Hooke, 1984; Hudson and Kesel, 2000; Lagasse et al., 2004; Nanson and Hickin, 1986). Specifically, a progressive increase in bend sinuosity enhances the erosive power of the secondary helical flow, causing a progressive deepening of the pool in the axial zone of the bend (Willis and Tang, 2010). The slope of descending thalweg trajectories indicates the rate of channel incision, which reflects the amount of time required to reach the equilibrium depth.

The thalweg trajectories, which characterize the evolution of the meander after the achievement of the formative depth, show either sub-horizontal (N1 – N4) or steep (S1 – S4) ascending patterns. Sub horizontal thalweg trajectories (inclination $< 2^\circ$) point out that the channel mainly shifted laterally with a minor influence of aggradational or degradational processes. Rising thalweg trajectories indicate that the channel shifted laterally under aggradational conditions, and, as for bar brink trajectories, gently- and steeply-rising trajectories are associated with a low and high values of the ratio between vertical and lateral shift rate, respectively. In-channel aggradation can either occur when its transport capability decreases, as in consequence of avulsion or cutoff (Toonen et al., 2012), or when, under aggradational and short-term dynamic equilibrium conditions (*sensu* Allen, 2000), sediments are stored in the channel in order to maintain a constant equilibrium depth. The first scenario is documented at sites N2, N4 and S1 (Fig 8) and implies the progressive decrease of channel cross-sectional area through deposition of fine-grained sediments on the channel floor. The similarity between depth of active channels (N1, N3 and S2 – S4) and thickness of the associated bar indicate that

they migrated depositing sediments at their base in order to keep the equilibrium with the surrounding environment.

5.3 Brink and thalweg trajectory combinations

The study sites document an early stage where brink and thalweg trajectories show an ascending and descending trends, respectively. This stage, named “formative phase” in figure 13, documents the onset of bar depositions and the achievement of the channel depth equilibrium. The angle between brink and thalweg trajectory controls the increase in point bar thickness, and is higher where channels cut rapidly into highly-aggrading marshes. Minor modifications to this process are represented by changes in climbing angle of the brink trajectory, which can be triggered by changes in ratio between vertical and lateral shift rate of the channel.

Once the equilibrium depth is reached, six of eight study channels show that thalweg and brink trajectories move following an almost parallel pathway (“equilibrium phase” in figure 13). It arises that, once the channel reached its equilibrium depth, this was kept constant shifting the thalweg in parallel with the bar brink zone. Minor variations from this overall behavior are documented at sites N4 and S4. Thalweg trajectory at site N4 slightly cuts down before the onset of the abandonment stage (Fig 8). This downward shift would be consistent either with an increase in water discharge, possibly due to a piracy in the upstream reaches of the tidal network, or with an increase in channel sinuosity due to adjustment of its planform shape. Differently, the rising thalweg trajectory at site S4 suggests that after reaching its formative depth, the channel decreased its discharge, possibly as consequence of changes in channel network structure.

BRINK TRAJECTORY VARIABILITY

Depending on vertical / lateral shift ratio

Processes

- salt marsh aggradation
- low substrate compaction
- low resistance deposits in the outer bank

Products

LINEAR & GENTLE

constant ratio in time
predominance of lateral migration

- salt marsh aggradation
- high substrate compaction

NON-LINEAR (concave) & STEEP

increasing ratio in time
aggradation becomes predominant

- salt marsh aggradation
- high substrate compaction
- low resistance deposits in the outer bank

NON-LINEAR (convex) & STEEP

decreasing ratio in time
lateral migration becomes predominant

FORMATIVE PHASE	EQUILIBRIUM PHASE
DIVERGENT brink and thalweg trajectories	PARALLEL brink and thalweg trajectories
Growth of the channel toward its equilibrium	The channel keeps the equilibrium cross section
Increase in point bar thickness	Constant point bar thickness

Figure 13 Summary of the main architectural features associated with development and combination of different point bar brink and channel thalweg trajectories.

6. CONCLUSIONS

The concept of trajectories analysis has been applied at the point bar scale in order to improve our understanding of the behavior of tidal point bars in micro-tidal and highly aggradational landscapes.

The main conclusions, as highlighted in figure 13, can be summarized as follows:

i) Bar brink trajectories are all ascending and show linear or non-linear trends whether the ratio between vertical and lateral shift rate is constant or not through time. When this ratio increases or decreases through time, non-linear trajectories show a concave- or convex-upward profile, respectively (Fig 10). The different shapes and steepness of these trajectories suggest that the bar brinks register different aggradational conditions while shifting laterally.

ii) The local increase in subsidence, due to bar loading on a soft (i.e. peaty) substrate can be a key factor for the development of steep brink trajectories. In particular, our results suggest that the differential compaction of sandy and peaty substrates plays a significant role in the creation of local accommodation, which controls the steepness of the trajectory.

iii) Thalweg trajectories can be both ascending and descending. The different inclination of the descending portion of the thalweg trajectory can be ascribed to the incision rate of the channel while shifting laterally. This is influenced by local tidal prism, bend geometry and evolution of the channel network. Ascending thalweg trajectories suggest that the channel shift laterally under aggradational conditions.

iv) The combination of brink and thalweg trajectories reveals the variation of point bar thickness during its evolution and allows one to differentiate an early formative phase followed by a late equilibrium phase. During the formative phase, the two trajectories diverge from one another because of the growth of the channel toward its

equilibrium depth (formative depth). In the following equilibrium phase, the two trajectories shift in parallel, allowing the channel to keep its equilibrium depth. Variations to this evolution model can be triggered by modification of channel discharge, possibly triggered by piracy events and channel network evolution.

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REFERENCES

- Allen, J.R.L., 2000. Morphodynamics of Holocene salt marshes: a review sketch from the Atlantic and Southern North Sea coasts of Europe. *Quat. Sci. Rev.* 19, 1155–1231.
- Allen, J.R.L., 1999. Geological impacts on coastal wetland landscapes: Some general effects of sediment autocompaction in the Holocene of northwest Europe. *The Holocene* 9, 1–12. <https://doi.org/10.1191/095968399674929672>
- Amorosi, A., Fontana, A., Antonioli, F., Primon, S., Bondesan, A., 2008. Post-LGM sedimentation and Holocene shoreline evolution in the NW Adriatic coastal area.

- 631 GeoActa 7, 41–67.
- 632 Barwis, J.H., 1978. Sedimentology of some South Carolina tidal-creek point bars,
633 and a comparison with thier fluvial counterparts, in: Miall, A.D. (Ed.), *Fluvial*
634 *Sedimentology*. Canadian Society of Petroleum Geologists Memoir 5, pp. 129–
635 160.
- 636 Beeson, C.E., Doyle, P.F., 1995. Comparison of bank erosion at vegetated and non-
637 vegetated channel bends. *Water Resour. Bull.* 31, 983–990.
- 638 Bellucci, L.G., Frignani, M., Cochran, J.K., Albertazzi, S., Zaggia, L., Cecconi, G.,
639 Hopkins, H., 2007. ²¹⁰Pb and ¹³⁷Cs as chronometers for salt marsh accretion
640 in the Venice Lagoon - links to flooding frequency and climate change. *J.*
641 *Environ. Radioact.* 97, 85–102. <https://doi.org/10.1016/j.jenvrad.2007.03.005>
- 642 Boaga, J., Ghinassi, M., D’Alpaos, A., Deidda, G.P., Rodriguez, G., Cassiani, G.,
643 2018. Geophysical investigations unravel the vestiges of ancient meandering
644 channels and their dynamics in tidal landscapes. *Sci. Rep.* 8, 1–8.
645 <https://doi.org/10.1038/s41598-018-20061-5>
- 646 Bondesan, A., Meneghel, M., 2004. Geomorfologia della provincia di Venezia. Note
647 illustrative della carta geomorfologica della provincia di Venezia.
- 648 Brice, J.C., 1974. Evolution of Meander Loops. *Geol. Soc. Am. Bull.* 85, 581–586.
- 649 Bridge, J.S., Alexander, J., Collier, R.E.L., Gawthorpe, R.L., Jarvis, J., 1995.
650 Ground-penetrating radar and coring used to study the large-scale structure of
651 point-bar deposits in three dimensions. *Sedimentology* 42, 839–852.
- 652 Bridges, P.H., Leeder, M.R., 1976. Sedimentary model for intertidal mudflat
653 channels, with examples from the Solway Firth, Scotland. *Sedimentology* 23,
654 533–552. <https://doi.org/10.1111/j.1365-3091.1976.tb00066.x>
- 655 Brivio, L., Ghinassi, M., D’Alpaos, A., Finotello, A., Fontana, A., Roner, M., Howes,

- N., 2016. Aggradation and lateral migration shaping geometry of a tidal point bar: An example from salt marshes of the Northern Venice Lagoon (Italy). *Sediment. Geol.* 343, 141–155. <https://doi.org/10.1016/j.sedgeo.2016.08.005>
- Candel, J.H.J., Makaske, B., Storms, J.E.A., Wallinga, J., 2017. Oblique aggradation: A novel explanation for sinuosity of low-energy streams in peat-filled valley systems. *Earth Surf. Process. Landforms* 42, 2679–2696. <https://doi.org/10.1002/esp.4100>
- Carniello, L., D’Alpaos, A., Defina, A., 2011. Modeling wind waves and tidal flows in shallow micro-tidal basins. *Estuar. Coast. Shelf Sci.* 92, 263–276. <https://doi.org/10.1016/j.ecss.2011.01.001>
- Carniello, L., D’Alpaos, A., Botter, G., Rinaldo, A., 2016. Statistical characterization of spatiotemporal sediment dynamics in the Venice lagoon. *J. Geophys. Res. Earth Surf.* 121, 1049–1064. <https://doi.org/10.1002/2015JF003793>.Received
- Carniello, L., Defina, A., D’Alpaos, L., 2009. Morphological evolution of the Venice lagoon: Evidence from the past and trend for the future. *J. Geophys. Res. Earth Surf.* 114, 1–10. <https://doi.org/10.1029/2008JF001157>
- Choi, K., Jo, J., 2015. Morphodynamics and stratigraphic architecture of compound dunes on the open-coast macrotidal flat in the northern Gyeonggi Bay, west coast of Korea. *Mar. Geol.* 366, 34–48. <https://doi.org/10.1016/j.margeo.2015.05.002>
- Colombera, L., Mountney, N.P., Russell, C.E., Shiers, M.N., McCaffrey, W.D., 2017. Geometry and compartmentalization of fluvial meander-belt reservoirs at the bar-form scale: Quantitative insight from outcrop, modern and subsurface analogues. *Mar. Pet. Geol.* 82, 35–55. <https://doi.org/10.1016/j.marpetgeo.2017.01.024>

- 681 Crosato, A., 2009. Physical explanations of variations in river meander migration
 682 rates from model comparison. *Earth Surf. Process. Landforms* 34, 2078–2086.
 683 <https://doi.org/10.1002/esp.1898>
- 684 D’Alpaos, A., Carniello, L., Rinaldo, A., 2013. Statistical mechanics of wind wave-
 685 induced erosion in shallow tidal basins: Inferences from the Venice Lagoon.
 686 *Geophys. Res. Lett.* 40, 3402–3407. <https://doi.org/10.1002/grl.50666>
- 687 D’Alpaos, A., Ghinassi, M., Finotello, A., Brivio, L., Bellucci, L.G., Marani, M., 2017.
 688 Tidal meander migration and dynamics: A case study from the Venice Lagoon.
 689 *Mar. Pet. Geol.* 87, 80–90. <https://doi.org/10.1016/j.marpetgeo.2017.04.012>
- 690 D’Alpaos, A., Lanzoni, S., Marani, M., Fagherazzi, S., Rinaldo, A., 2005. Tidal
 691 network ontogeny: Channel initiation and early development. *J. Geophys. Res.*
 692 *Earth Surf.* 110, 1–14. <https://doi.org/10.1029/2004JF000182>
- 693 D’Alpaos, A., Lanzoni, S., Marani, M., Rinaldo, A., 2010. On the tidal prism-channel
 694 area relations. *J. Geophys. Res. Earth Surf.* 115, 1–13.
 695 <https://doi.org/10.1029/2008JF001243>
- 696 D’Alpaos, A., Lanzoni, S., Marani, M., Rinaldo, A., 2007. Landscape evolution in tidal
 697 embayments: Modeling the interplay of erosion, sedimentation, and vegetation
 698 dynamics. *J. Geophys. Res. Earth Surf.* 112, 1–17.
 699 <https://doi.org/10.1029/2006JF000537>
- 700 D’Alpaos, L., 2010. Fatti e misfatti di idraulica lagunare. La laguna di Venezia dalla
 701 diversione dei fiumi alle nuove opere alle bocche di porto, Venezia: Istituto
 702 Veneto di Scienze, Lettere ed Arti.
- 703 Dalrymple, R.W., Makino, Y., Zaitlin, B., 1991. Temporal and spatial patterns of
 704 rhythmite deposition on mud flats in the macrotidal Cobequid Bay-Salmon River
 705 estuary. Bay of Fundy, Canada. *Clastic Tidal Sedimentol.* 16, 137–160.

- 706 Day, J.W., Rismondo, A., Scarton, F., Are, D., Cecconi, G., 1998. Relative sea level
707 rise and Venice lagoon wetlands. *J. Coast. Conserv.* 4, 27–34.
- 708 De Mowbray, T., 1983. The genesis of lateral accretion deposits in recent intertidal
709 mudflat channels, Solway Firth, Scotland. *Sedimentology* 30, 425–435.
710 <https://doi.org/10.1111/j.1365-3091.1983.tb00681.x>
- 711 Dietrich, W.E., Smith, J.D., 1984. Processes controlling the equilibrium bed
712 morphology in river meanders, in: *River Meandering: Proceedings of the*
713 *Conference Rivers '83*. pp. 759–769.
- 714 Dietrich, W.E., Smith, J.D., Dunne, T., 1979. Flow and Sediment Transport in a Sand
715 Bedded Meander. *J. Geol.* 87, 305–315. <https://doi.org/10.1086/628419>
- 716 Donnici, S., Madricardo, F., Serandrei-Barbero, R., 2017. Sedimentation rate and
717 lateral migration of tidal channels in the Lagoon of Venice (Northern Italy).
718 *Estuar. Coast. Shelf Sci.* 198, 354–366.
719 <https://doi.org/10.1016/j.ecss.2017.02.016>
- 720 Durkin, P.R., Hubbard, S.M., Boyd, R.L., Leckie, D.A., 2015. Stratigraphic
721 Expression of Intra-Point-Bar Erosion and Rotation. *J. Sediment. Res.* 85,
722 1238–1257. <https://doi.org/10.2110/jsr.2015.78>
- 723 Elliott, R.E., 1985. Quantification of peat to coal compaction stages, based especially
724 on phenomena in the East Pennine Coalfield, England. *Proc. Yorksh. Geol. Soc.*
725 45, 163–172.
- 726 Fagherazzi, S., Gabet, E.J., Furbish, D.J., 2004. The effect of bidirectional flow on
727 tidal channel planforms. *Earth Surf. Process. Landforms* 29, 295–309.
728 <https://doi.org/10.1002/esp.1016>
- 729 Finotello, A., Lanzoni, S., Ghinassi, M., Marani, M., Rinaldo, A., D'Alpaos, A., 2018.
730 Field migration rates of tidal meanders recapitulate fluvial morphodynamics.

- 731 Proc. Natl. Acad. Sci. 115, 1463–1468.
- 732 <https://doi.org/10.1073/pnas.1711330115>
- 733 Fisk, H.N., 1944. Geological Investigation of the Alluvial Valley of the Lower
- 734 Mississippi River. US Army Corps Eng. Mississippi River Comm. 1–78.
- 735 Fontana, A., Mozzi, P., Marchetti, M., 2014. Alluvial fans and megafans along the
- 736 southern side of the Alps. Sediment. Geol. 301, 150–171.
- 737 <https://doi.org/10.1016/j.sedgeo.2013.09.003>
- 738 Frothingham, K.M., Rhoads, B.L., 2003. Three-dimensional flow structure and
- 739 channel change in an asymmetrical compound meander loop, Embarras River,
- 740 Illinois. Earth Surf. Process. Landforms 28, 625–644.
- 741 <https://doi.org/10.1002/esp.471>
- 742 Fruergaard, M., Andersen, T.J., Nielsen, L.H., Madsen, A.T., Johannessen, P.N.,
- 743 Murray, A.S., Kirkegaard, L., Pejrup, M., 2011. Punctuated sediment record
- 744 resulting from channel migration in a shallow sand-dominated micro-tidal
- 745 lagoon, Northern Wadden Sea, Denmark. Mar. Geol. 280, 91–104.
- 746 <https://doi.org/10.1016/j.margeo.2010.12.003>
- 747 Gabet, E.J., 1998. Lateral Migration and Bank Erosion in a Saltmarsh Tidal Channel
- 748 in San Francisco Bay, California. Estuaries 21, 745–753.
- 749 <https://doi.org/10.2307/1353278>
- 750 Garofalo, D., 1980. The Influence of Wetland Vegetation on Tidal Stream Channel
- 751 Migration and Morphology. Estuaries 3, 258–270.
- 752 <https://doi.org/10.2307/1352081>
- 753 Gatto, P., Carbognin, L., 1981. The lagoon of Venice: Natural environmental trend
- 754 and man-induced modification. Hydrol. Sci. Bull. 26, 379–391.
- 755 <https://doi.org/10.1080/02626668109490902>

- 756 Ghinassi, M., Brivio, L., D'Alpaos, A., Finotello, A., Carniello, L., Marani, M., Cantelli,
757 A., 2018a. Morphodynamic evolution and sedimentology of a microtidal
758 meander bend of the Venice Lagoon (Italy). *Mar. Pet. Geol.* 96, 391–404.
759 <https://doi.org/10.1016/j.marpetgeo.2018.06.011>
- 760 Ghinassi, M., Ielpi, A., Aldinucci, M., Fustic, M., 2016. Downstream-migrating fluvial
761 point bars in the rock record. *Sediment. Geol.* 334, 66–96.
762 <https://doi.org/10.1016/j.sedgeo.2016.01.005>
- 763 Ghinassi, M., Moody, J.A., Martin, D., 2018b. Influence of extreme and annual floods
764 on point-bar sedimentation: Inferences from Powder River , Montana , USA.
765 *Geol. Soc. Am. Bull.* 1–13.
766 <https://doi.org/10.1130/B31990.1/4292202/b31990.pdf>
- 767 Ghinassi, M., Nemec, W., Aldinucci, M., Nehyba, S., Ozaksoy, V., Fidolini, F., 2014.
768 Plan-form evolution of ancient meandering rivers reconstructed from longitudinal
769 outcrop sections. *Sedimentology* 61, 952–977.
770 <https://doi.org/10.1111/sed.12081>
- 771 Gobo, K., Ghinassi, M., Nemec, W., 2015. Gilbert-type deltas recording short-term
772 base-level changes: Delta-brink morphodynamics and related foreset facies.
773 *Sedimentology* 62, 1923–1949. <https://doi.org/10.1111/sed.12212>
- 774 Green, M.O., Coco, G., 2007. Sediment transport on an estuarine intertidal flat:
775 Measurements and conceptual model of waves, rainfall and exchanges with a
776 tidal creek. *Estuar. Coast. Shelf Sci.* 72, 553–569.
777 <https://doi.org/10.1016/j.ecss.2006.11.006>
- 778 Helland-Hansen, W., Hampson, G.J., 2009. Trajectory analysis: Concepts and
779 applications. *Basin Res.* 21, 454–483. [https://doi.org/10.1111/j.1365-](https://doi.org/10.1111/j.1365-2117.2009.00425.x)
780 [2117.2009.00425.x](https://doi.org/10.1111/j.1365-2117.2009.00425.x)

- 781 Henriksen, S., Hampson, G.J., Helland-Hansen, W., Johannessen, E.P., Steel, R.J.,
 782 2009. Shelf edge and shoreline trajectories, a dynamic approach to stratigraphic
 783 analysis. *Basin Res.* 21, 445–453. [https://doi.org/10.1111/j.1365-](https://doi.org/10.1111/j.1365-2117.2009.00432.x)
 784 [2117.2009.00432.x](https://doi.org/10.1111/j.1365-2117.2009.00432.x)
- 785 Henriksen, S., Helland-hansen, W., Bullimore, S., 2011. Relationships between
 786 shelf-edge trajectories and sediment dispersal along depositional dip and strike:
 787 a different approach to sequence stratigraphy. *Basin Res.* 23, 3–21.
 788 <https://doi.org/10.1111/j.1365-2117.2010.00463.x>
- 789 Hooke, J.M., 1984. Changes in river meanders: a review of techniques and results of
 790 analyses. *Prog. Phys. Geogr. Earth Environ.* 8, 473–508.
 791 <https://doi.org/10.1177/030913338400800401>
- 792 Hooke, R.L.B., 1975. Distribution of sediment transport and shear stress in a
 793 meander bend. *J. Geol.* 83, 543–565.
- 794 Hudson, P.F., Kesel, R.H., 2000. Channel migration and meander-bend curvature in
 795 the lower Mississippi River prior to major human modification. *Geology* 28, 531–
 796 534. [https://doi.org/10.1130/0091-7613\(2000\)28<531:CMAMCI>2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28<531:CMAMCI>2.0.CO;2)
- 797 Hughes, Z.J., 2012. Tidal Channels on Tidal Flats and Marshes, in: Davis, R.A.,
 798 Dalrymple, R.W. (Eds.), *Principles of Tidal Sedimentology*. pp. 269–300.
 799 <https://doi.org/10.1007/978-94-007-0123-6>
- 800 Ielpi, A., Ghinassi, M., 2014. Planform architecture, stratigraphic signature and
 801 morphodynamics of an exhumed Jurassic meander plain (Scalby Formation,
 802 Yorkshire, UK). *Sedimentology* 61, 1923–1960.
 803 <https://doi.org/10.1111/sed.12122>
- 804 Ielpi, A., Gibling, M.R., Bashforth, A.R., Dennar, C.I., 2015. Impact of Vegetation On
 805 Early Pennsylvanian Fluvial Channels: Insight From the Joggins Formation of

- 806 Atlantic Canada. *J. Sediment. Res.* 85, 999–1018.
 807 <https://doi.org/10.2110/jsr.2015.50>
- 808 Jackson, R.G.I., 1976. Depositional model of point bars in the Lower Wabash River.
 809 *J. Sediment. Petrol.* 46, 579–594.
- 810 Kent, D. V., Rio, D., Massari, F., Kukla, G., Lanci, L., 2002. Emergence of Venice
 811 during the Pleistocene. *Quat. Sci. Rev.* 21, 1719–1727.
 812 [https://doi.org/10.1016/S0277-3791\(01\)00153-6](https://doi.org/10.1016/S0277-3791(01)00153-6)
- 813 Lagasse, P.F., Zevenbergen, L.W., Spitz, W.J., Thorne, C.R., 2004. Methodology for
 814 Predicting Channel Migration. NCHRP Proj. No. 24-16 . Ayres Assoc. Inc. - Fort
 815 Collins, Color. <https://doi.org/10.17226/23352>
- 816 Lanzoni, S., Seminara, G., 2002. Long-term evolution and morphodynamic
 817 equilibrium of tidal channels. *J. Geophys. Res. Ocean.* 107, 3001.
 818 <https://doi.org/10.1029/2000JC000468>
- 819 Larue, D.K., Martinez, P.A., 1989. Use of bed-form climb models to analyze
 820 geometry and preservation potential of clastic facs and erosional surfaces. *Am.*
 821 *Assoc. Pet. Geol. Bull.* 73, 40–53.
- 822 Long, A.J., Waller, M.P., Stupples, P., 2006. Driving mechanisms of coastal change:
 823 Peat compaction and the destruction of late Holocene coastal wetlands. *Mar.*
 824 *Geol.* 225, 63–84. <https://doi.org/10.1016/j.margeo.2005.09.004>
- 825 Makaske, B., 2001. Anastomosing rivers: A review of their classification, origin and
 826 sedimentary products. *Earth Sci. Rev.* 53, 149–196.
 827 [https://doi.org/10.1016/S0012-8252\(00\)00038-6](https://doi.org/10.1016/S0012-8252(00)00038-6)
- 828 Marani, M., D’Alpaos, A., Lanzoni, S., Carniello, L., Rinaldo, A., 2007. Biologically-
 829 controlled multiple equilibria of tidal landforms and the fate of the Venice lagoon.
 830 *Geophys. Res. Lett.* 34, 1–5. <https://doi.org/10.1029/2007GL030178>

- 831 Marani, M., Lanzoni, S., Zandolin, D., Seminara, G., Rinaldo, A., 2002. Tidal
832 meanders. *Water Resour. Res.* 38, 7-1-7–14.
833 <https://doi.org/10.1029/2001WR000404>
- 834 Massari, F., Grandesso, P., Stefani, C., Jobstraibizer, P.G., 2009. A small polyhistory
835 foreland basin evolving in a context of oblique convergence: the Venetian basin
836 (Chattian to Recent, Southern Alps, Italy), in: Allen, P.A., Homewood, P. (Eds.),
837 *Foreland Basins*. pp. 141–168. <https://doi.org/10.1002/9781444303810>
- 838 Massari, F., Rio, D., Serandrei Barbero, R., Asioli, A., Capraro, L., Fornaciari, E.,
839 Vergerio, P.P., 2004. The environment of Venice area in the past two million
840 years. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 202, 273–308.
841 [https://doi.org/10.1016/S0031-0182\(03\)00640-0](https://doi.org/10.1016/S0031-0182(03)00640-0)
- 842 McClennen, C.E., Housley, R.A., 2006. Late-Holocene Channel Meander Migration
843 and Mudflat Accumulation Rates, Lagoon of Venice, Italy. *J. Coast. Res.* 22,
844 930–945. <https://doi.org/10.2112/03-0113.1>
- 845 McGowen, J.H., Garner, L.E., 1970. Physiographic Features and Stratification Types
846 of Coarse-Grained Pointbars: Modern and Ancient Examples. *Sedimentology*
847 14, 77–111. <https://doi.org/10.1111/j.1365-3091.1970.tb00184.x>
- 848 Miall, A.D., 1985. Architectural-element analysis: A new method of facies analysis
849 applied to fluvial deposits. *Earth-Science Rev.* 22, 261–308.
850 [https://doi.org/10.1016/0012-8252\(85\)90001-7](https://doi.org/10.1016/0012-8252(85)90001-7)
- 851 Moody, J.A., Meade, R.H., 2014. Ontogeny of point bars on a river in a cold semi-
852 arid climate. *Bull. Geol. Soc. Am.* 126, 1301–1316.
853 <https://doi.org/10.1130/B30992.1>
- 854 Morris, J.T., Sundareshwar, P. V, Nietch, C.T., Kjerfve, B., Cahoon, D.R., 2002.
855 *Responses of Coastal Wetlands To Rising Sea Level. Ecology* 83, 2869–2877.

- 856 Mudd, S.M., D'Alpaos, A., Morris, J.T., 2010. How does vegetation affect
857 sedimentation on tidal marshes? Investigating particle capture and
858 hydrodynamic controls on biologically mediated sedimentation. *J. Geophys.*
859 *Res. Earth Surf.* 115, 1–14. <https://doi.org/10.1029/2009JF001566>
- 860 Nanson, G.C., Croke, J.C., 1992. A genetic classification of floodplains.
861 *Geomorphology* 4, 459–486.
- 862 Nanson, G.C., Hickin, E.J., 1986. A statistical analysis of bank erosion and channel
863 migration in western Canada. *Geol. Soc. Am. Bull.* 97, 497–504.
864 [https://doi.org/10.1130/0016-7606\(1986\)97<497:ASAOBE>2.0.CO;2](https://doi.org/10.1130/0016-7606(1986)97<497:ASAOBE>2.0.CO;2)
- 865 Pranter, M.J., Ellison, A.I., Cole, R.D., Patterson, P.E., 2007. Analysis and modeling
866 of intermediate-scale reservoir heterogeneity based on a fluvial point-bar
867 outcrop analog, Williams Fork Formation, Piceance Basin, Colorado. *Am.*
868 *Assoc. Pet. Geol. Bull.* 91, 1025–1051. <https://doi.org/10.1306/02010706102>
- 869 Puigdefabregas, C., 1973. Miocene point bar deposits in the Ebro Basin, Northern
870 Spain. *Sedimentology* 20, 133–144.
- 871 Puigdefabregas, C., Van Vliet, A., 1977. Meandering stream deposits from the
872 Tertiary of the Southern Pyrenees. *Fluv. Sedimentol. Mem.* 5 5, 469–485.
- 873 Rajchl, M., Uličný, D., 2005. Depositional record of an avulsive fluvial system
874 controlled by peat compaction (Neogene, Most Basin, Czech Republic).
875 *Sedimentology* 52, 601–625. <https://doi.org/10.1111/j.1365-3091.2005.00691.x>
- 876 Rieu, R., van Heteren, S., Van der Spek, A.J.F., De Boer, P.L., 2005. Development
877 and preservation of a Mid-Holocene tidal-channel network offshore the Western
878 Netherlands. *J. Sediment. Res.* 75, 409–419.
879 <https://doi.org/10.21110/jsr.2005.032>
- 880 Roner, M., D'Alpaos, A., Ghinassi, M., Marani, M., Silvestri, S., Franceschinis, E.,

- 881 Realdon, N., 2016. Spatial variation of salt-marsh organic and inorganic
882 deposition and organic carbon accumulation: Inferences from the Venice
883 lagoon, Italy. *Adv. Water Resour.* 93, 276–287.
884 <https://doi.org/10.1016/j.advwatres.2015.11.011>
- 885 Roner, M., Ghinassi, M., Fedi, M., Liccioli, L., Bellucci, L.G., Brivio, L., D'Alpaos, A.,
886 2017. Latest Holocene depositional history of the southern Venice Lagoon, Italy.
887 *The Holocene* 27, 1731–1744. <https://doi.org/10.1177/0959683617708450>
- 888 Silvestri, S., Defina, A., Marani, M., 2005. Tidal regime, salinity and salt marsh plant
889 zonation. *Estuar. Coast. Shelf Sci.* 62, 119–130.
890 <https://doi.org/10.1016/j.ecss.2004.08.010>
- 891 Smith, D.G., 1987. Meandering river point bar lithofacies models: modern and
892 ancient examples compared, in: Ethridge, F.G., Flores, R.M., Harvey, M.D.
893 (Eds.), *Recent Developments in Fluvial Sedimentology: Contributions from the*
894 *Third International Fluvial Sedimentology Conference*. SEPM Society for
895 *Sedimentary Geology*, pp. 83–91.
- 896 Smith, D.G., Hubbard, S.M., Leckie, D.A., Fustic, M., 2009. Counter point bar
897 deposits: Lithofacies and reservoir significance in the meandering modern
898 peace river and ancient McMurray formation, Alberta, Canada. *Sedimentology*
899 56, 1655–1669. <https://doi.org/10.1111/j.1365-3091.2009.01050.x>
- 900 Stefanon, L., Carniello, L., D'Alpaos, A., Rinaldo, A., 2012. Signatures of sea level
901 changes on tidal geomorphology: Experiments on network incision and retreat.
902 *Geophys. Res. Lett.* 39, 1–6. <https://doi.org/10.1029/2012GL051953>
- 903 Terwindt, J.H.J., 1988. Palaeotidal reconstructions of inshore tidal depositional
904 environments, in: De Boer, P.L., van Gelder, A., Nio, S.-D. (Eds.), *Tide-*
905 *Influenced Sedimentary Environments*. pp. 233–263.

- 906 Toonen, W.H.J., Kleinhans, M.G., Cohen, K.M., 2012. Sedimentary architecture of
 907 abandoned channel fills. *Earth Surf. Process. Landforms* 37, 459–472.
 908 <https://doi.org/10.1002/esp.3189>
- 909 Törnqvist, T.E., Wallace, D.J., Storms, J.E.A., Wallinga, J., Van Dam, R.L., Blaauw,
 910 M., Derksen, M.S., Klerks, C.J.W., Meijneken, C., Snijders, E.M.A., 2008.
 911 Mississippi Delta subsidence primarily caused by compaction of Holocene
 912 strata. *Nat. Geosci.* 1, 173–176. <https://doi.org/10.1038/ngeo129>
- 913 Tosi, L., Rizzetto, F., Bonardi, M., Donnici, S., Serandrei-Barbero, R., Toffoletto, F.,
 914 2007. Note Illustrative della Carta Geologica d'Italia alla scala 1:50.000. foglio
 915 148-149, Chioggia-Malamocco.
- 916 van Asselen, S., 2011. The contribution of peat compaction to total basin
 917 subsidence: Implications for the provision of accommodation space in organic-
 918 rich deltas. *Basin Res.* 23, 239–255. [https://doi.org/10.1111/j.1365-](https://doi.org/10.1111/j.1365-2117.2010.00482.x)
 919 [2117.2010.00482.x](https://doi.org/10.1111/j.1365-2117.2010.00482.x)
- 920 van Asselen, S., Stouthamer, E., Smith, N.D., 2010. Factors Controlling Peat
 921 Compaction in Alluvial Floodplains: A Case Study in the Cold-Temperate
 922 Cumberland Marshes, Canada. *J. Sediment. Res.* 80, 155–166.
 923 <https://doi.org/10.2110/jsr.2010.015>
- 924 van de Lageweg, W.I., Schuurman, F., Cohen, K.M., van Dijk, W.M., Shimizu, Y.,
 925 Kleinhans, M.G., 2016. Preservation of meandering river channels in uniformly
 926 aggrading channel belts. *Sedimentology* 63, 586–608.
 927 <https://doi.org/10.1111/sed.12229>
- 928 Willis, B.J., Tang, H., 2010. Three-Dimensional Connectivity of Point-Bar Deposits. *J.*
 929 *Sediment. Res.* 80, 440–454. <https://doi.org/10.2110/jsr.2010.046>
- 930 Zecchin, M., Baradello, L., Brancolini, G., Donda, F., Rizzetto, F., Tosi, L., 2008.

- 931 Sequence stratigraphy based on high-resolution seismic profiles in the late
932 Pleistocene and Holocene deposits of the Venice area. *Mar. Geol.* 253, 185–
933 198. <https://doi.org/10.1016/j.margeo.2008.05.010>
- 934 Zecchin, M., Brancolini, G., Tosi, L., Rizzetto, F., Caffau, M., Baradello, L., 2009.
935 Anatomy of the Holocene succession of the southern Venice lagoon revealed by
936 very high-resolution seismic data. *Cont. Shelf Res.* 29, 1343–1359.
937 <https://doi.org/10.1016/j.csr.2009.03.006>
- 938 Zecchin, M., Tosi, L., Caffau, M., Baradello, L., Donnici, S., 2014. Sequence
939 stratigraphic significance of tidal channel systems in a shallow lagoon (Venice,
940 Italy). *The Holocene* 24, 646–658. <https://doi.org/10.1177/0959683614526903>
941