A flow-curvature-based model for channel meandering in tidal marshes

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Abstract

Channel meandering is ubiquitous in tidal marshes, yet it is either omitted or weakly implemented in morphodynamic models. Here we propose a novel numerical method to simulate channel meandering in tidal marshes on a Cartesian grid. The method calculates a first-order flow by considering the balance between pressure gradient and bed friction. To account for flow momentum shift towards meander outer banks, the flow is empirically modified. Unlike previous simplified methods that relied on the curvature of the bank, this modification is based on the curvature of the flow, making the model suitable for use in dendritic channel networks. The modified flow intrinsically accounts for the topographic steering effect, which tends to deflect the momentum toward the outer bank. As a result, the outer bank becomes steeper and erodes due to soil creep. Additionally, the outer bank experiences erosion proportional to the flow curvature. This erosion mechanism parameterizes the direct erosion caused by flow impacting the bank through a proportionality coefficient, which modulates the rate of lateral channel migration. Deposition on the inner bank is automatically simulated by the model, triggered by reduced bed shear stress in that area. The model accurately reproduces channel lateral migration and sinuosity development, and associated processes such as meander cutoffs, channel piracies, and network reorganizations. The model provides an efficient tool for predicting marsh landscape evolution from decades to millennia, which will enable exploring how lateral migration and meandering of tidal channels affect marsh ecomorphodynamics, carbon and nutrient cycling, drainage efficiency, and pond dynamics.

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Key Points:

- Novel, depth-averaged, Cartesian-grid-based numerical model to simulate channel meandering in tidal marshes
- Realistic tidal channel morphologies and dynamics are reproduced, including cuspate bends, meander cutoffs, and channel piracies
- Model simulates the ecomorphodynamic evolution of tidal marshes with branching and meandering channel networks over decades to millennia

Keywords

Tidal marshes, Tidal Meanders, Channel Migration, Ecomorphodynamics, Numerical Modelling

Abstract

Channel meandering is ubiquitous in tidal marshes, yet it is either omitted or weakly implemented in morphodynamic models. Here we propose a novel numerical method to simulate channel meandering in tidal marshes on a Cartesian grid. The method calculates a first-order flow by considering the balance between pressure gradient and bed friction. To account for flow momentum shift towards meander outer banks, the flow is empirically modified. Unlike previous simplified methods that relied on the curvature of the bank, this modification is based on the curvature of the flow, making the model suitable for use in dendritic channel networks. The modified flow intrinsically accounts for the topographic steering effect, which tends to deflect the momentum toward the outer bank. As a result, the outer bank becomes steeper and erodes due to soil creep. Additionally, the outer bank experiences erosion proportional to the flow curvature. This erosion mechanism parameterizes the direct erosion caused by flow impacting the bank through a proportionality coefficient, which modulates the rate of lateral channel migration. Deposition on the inner bank is automatically simulated by the model, triggered by reduced bed shear stress in that area. The model accurately reproduces channel lateral migration and sinuosity development, and associated processes such as meander cutoffs, channel piracies, and network reorganizations. The model provides an efficient tool for predicting marsh landscape evolution from decades to millennia, which will enable exploring how lateral migration and meandering of tidal channels affect marsh ecomorphodynamics, carbon and nutrient cycling, drainage efficiency, and pond dynamics.

1 Introduction

Branching and meandering tidal channel networks are widespread in low-lying coastal areas, from open coasts to back-barrier lagoons, as well as in tidally-influenced fluvio-deltaic and estuarine plains (Fagherazzi et al., 1999; Kearney and Fagherazzi, 2016; Rinaldo et al., 1999a, 1999b) (Figure 1). Besides governing tide propagation, thereby mediating changes in local mean sea level and tidal range, tidal channel networks control the ecomorphodynamic evolution of the wetlands they drain by regulating fluxes of water, sediments, nutrients, pollutants, and particulate matter (Coco et al., 2013; Hughes, 2012; Marani et al., 2003). Given their importance, there is a pressing need to understand the formation and evolution of these landscape features, especially when considering climatic changes and anthropogenic disturbances (Stefanon et al., 2012; Zhou et al., 2014a). Even though some major progress has occurred over the last two decades in terms of numerical modeling and laboratory experiments, the long-term prediction of tidal channel network morphological evolution remains a major challenge both at the theoretical and practical levels (Belliard et al., 2015; Belliard, 2014; Cleveringa and Oost, 1999; Cosma et al., 2020; Geng et al., 2020; Hood, 2014; Kleinhans et al., 2015; Stefanon et al., 2012; van de Vijsel et al., 2023; Xu et al., 2017; Zhou et al., 2014b).

A major limitation in understanding the long-term evolution of tidal marshes has been the widely accepted notion that tidal channel networks are relatively stable geomorphological features (Choi and Jo, 2015; Fagherazzi et al., 2004; Gabet, 1998; Garofalo, 1980; Kleinhans et al., 2009). This notion caused the dynamics of mature networks to be largely disregarded in conventional ecogeomorphological practices, projections of local sea-level changes, and estimates of biogeochemical fluxes in highly-productive tidal wetland ecosystems. Although channel network ontogeny occurs on timescales considerably shorter than those involved in other relevant ecomorphodynamic processes such as sea-level changes and vegetation dynamics (D'Alpaos et al., 2005; Fagherazzi and Sun, 2004; Kirwan and Murray, 2007), field observations have shown that channels are highly dynamic systems that can change over decadal timescales (Rizzetto and Tosi, 2012; Shimozono et al., 2019; Wilson et al., 2017).

Recent research provided proof of concept that tidal channel lateral migration and adjustment to changing external forcings are key, yet largely unexplored, drivers for marsh morphodynamics (Cosma et al., 2019, 2020; Finotello et al., 2018, 2019a, 2020c, 2020b; Gao et al., 2023). In particular, channel meandering, coupled with the characteristically high drainage density of channel networks, is likely to produce frequent piracies and network reorganizations (Cosma et al., 2020; Letzsch and Frey, 1980; Litwin et al., 2013). These changes drive profound modifications of hydrological connectivity, thus ultimately affecting the morphology, sedimentology, ecology, and carbon storage capacity of the marsh system. For instance, the continued reworking of tidal channel banks is likely to have important effects on vegetation dynamics and blue carbon fluxes because it removes established vegetation along the channel margins and allows the process of ecological succession to begin anew; moreover, it exposes organic-rich substrates, thereby affecting carbon fluxes (Elsey-Quirk et al., 2019; Hopkinson et al., 2018; Kalra et al., 2021; Leonardi et al., 2016). Meandering is also important from sedimentological perspectives since tidal channels are typically preserved in the sedimentary record through both laterally accreting deposits, which occur mostly at meander bends, and channel infilling as the tidal prism decreases when channels are partially abandoned via either avulsion or meander cutoff (Brivio et al., 2016; Cosma et al., 2019, 2020; Gao et al., 2023).

In spite of its prominence and widespread occurrence, meandering in tidal channels lacks the detailed inspection that has been devoted to their fluvial relatives (Güneralp et al., 2012; Hooke, 2013; Ikeda et al., 1981; Leopold and Wolman, 1960; Parker et al., 1982). Of the various numerical models employed to reproduce the planimetric development of tidal channel networks (e.g., D'Alpaos et al., 2005; Fagherazzi and Sun, 2004; Kirwan and Murray, 2007; Van Maanen et al., 2013; van de Vijsel et al., 2023), none has insofar accounted for meander bend evolution. Numerical models have been proposed to simulate the evolution of individual, single-thread, sinuous tidal channels, assuming that a close morphodynamic similarity exists between tidal and fluvial meanders. Hence, some authors directly employed modified versions of meandering-river centerline-migration models (e.g., Coulthard and Van De Wiel, 2012; Howard and Knutson, 1984; Ikeda et al., 1981; Nicholas, 2013; Seminara et al., 2001; Zolezzi and Seminara, 2001) to investigate the evolution of single-thread tidal meandering channels under the influence of changing tidal asymmetry (Fagherazzi et al., 2004) and in highly cohesive mudflat environments (Kleinhans et al., 2009). Solari et al. (2002) developed the only existing numerical model capable of forecasting the flow patterns and bed topography in mildly sinuous, single-thread tidal channels through the implementation of linear stability theory derived for meandering rivers.

Despite their appeal, river meander models have major limitations when applied to tidal channels. First, meander models are difficult to implement in a 2D cartesian domain, i.e., a gridded-based domain, which is generally the structure of salt-marsh ecomorphodynamic models (e.g., Belliard et al., 2015; D'Alpaos et al., 2007; Kirwan and Murray, 2007; Mariotti, 2020). Second, meandering tidal channels do not occur as isolated single-thread streams that freely wander through an otherwise empty alluvial plain. Rather, they are found within morphologically complex tidal channel networks where lateral channel migration leads to interactions with neighboring channels, producing frequent reorganization of the whole network structure and the formation of multichannel loops (e.g., Konkol et al., 2022) in addition to the classic meander neck cutoffs observed in rivers (Gao et al., 2023). Third, tidal channels are characterized by highly variable widths due to channel funneling (Lanzoni and D'Alpaos, 2015), a phenomenon that can be disregarded when dealing with meandering rivers, which display limited width variations along their courses (Finotello et al., 2020a). Finally, in contrast to coastal wetlands, where sediments are predominantly fine and cohesive, conventional river meander models consider coarse non-cohesive sediments (e.g., Bogoni et al., 2017).

Existing models for tidal marsh morphodynamics do not properly include the dynamics associated with channel meandering (Boechat Albernaz et al., 2023; D'Alpaos et al., 2005; Gourgue et al., 2022; Mariotti, 2020; van de Vijsel et al., 2023; Xu et al., 2022), either because they do not account for the flow modification due to the curvature (which is associated with the secondary flow) or because they do not include bank erosion. The channel networks simulated with these models display some levels of sinuosity, which is mostly acquired in the early stage of channel network development and is driven by changes in tidal watersheds as the network evolves. Channel sinuosity is however smaller than the one observed in the field and, more importantly, the channels do not actively migrate once mature.

Here we propose a novel, computationally efficient numerical model capable of reproducing channel migration in tidal marshes and, more generally, the free evolution of tidal channel networks (i.e., including cutoffs and piracies). The model is informed and developed based on available field data and can simulate the evolution of salt-marsh landscapes on time scales ranging from decades to millennia.

2 Materials and Methods

2.1 MarshMorpho2D

Meandering in tidal channels was simulated using MarshMorpho2D, a numerically efficient marsh evolution model that includes a variety of hydrodynamic and sedimentary processes. A summary of its components is reported below, and more details can be found in previous publications (Mariotti, 2020; Mariotti et al., 2016, 2019).

The model calculates a base tidal flow velocity field $\boldsymbol{U} = \left(U_x, U_y\right)$, representing the average flow during half a tidal cycle (i.e., either the ebb or flood phase), based on a balance between pressure gradient and linearized bed friction, and considering a tide-averaged water depth h for each cell (Mariotti, 2018, 2020):

$$\nabla \cdot (h\boldsymbol{U}) = S \qquad (1)$$

$$\boldsymbol{U} \propto \frac{h^{4/3}}{n^2} \nabla\eta \qquad (2)$$

The term η denotes the elevation of the free surface, whereas S is the source term for the tidal prism, and it is equal to $r/2 - \max\,(-r/2, \min\,(z, r/2))]/(T/2)$, where z is the bed elevation, r is the tidal range, and T is the tidal period. For numerical stability reasons, the minimum water depth is set equal to h=1 cm.

Sediment, which is here assumed to be only comprised of mud, can be resuspended from the bottom by both tidal currents and wind waves (see Mariotti, 2020 for details). A sinusoidal modulation of the tidal currents is empirically introduced (Mariotti, 2020), thus allowing a more realistic estimate of the peak velocity (which reads $\pi/2\boldsymbol{U}$) and its associated sediment resuspension.

For simplicity, the sediment resuspension by currents and waves is summed linearly. The average tidal velocities (***U***) are used to estimate tidal dispersion (Di Silvio et al., 2010), which is employed to transport sediment in suspension. A constant baseline sediment diffusivity, equal to 10 m^2/s, is also included to account for all transport processes not directly simulated, such as wind-driven currents (by local forcings), wave-driven currents, shear-induced dispersion, and turbulent mixing.

Soil creep (i.e., gravity-driven diffusion of the elevation) is used to simulate marsh bank erosion (Mariotti et al., 2016, 2019). The creep coefficient is set equal to 0.1 m^2/yr and 3 m^2/yr in vegetated and unvegetated areas, respectively. Soil creep at the marsh banks (i.e., the interface between vegetated and unvegetated cells) is treated in a special way, given that bank erosion is driven by a complex interplay of hydrodynamic and geotechnical processes, and because its implementation is highly sensitive to the resolution of the computation (dx). Specifically, the constant diffusion term at the bank is set equal to $0.05dx$ ([m^2/yr], with dx in meters), and an additional transport term is set proportional to the velocity at the bank and equal to $0.2dx|\boldsymbol{U}|$ ([m^2/yr], with dx in meters). Overall, the interplay of these downslope transport mechanisms determines the shape of the channel cross-section and the related width-to-depth ratio.

Marsh vegetation is modeled as a function of the hydroperiod, which in turn is a function of bed elevation and tidal fluctuations. The upper limit for vegetation growth is set equal to the monthly median high water level, which depends on both the astronomic and the meteorological tides (Mariotti and Zapp, 2022). In addition to modifying soil creep (as described above), vegetation also modifies the bed drag, which in turn affects the flow. Specifically, Manning's friction coefficient is set equal to 0.02 $s/m^{1/3}$ and 0.1 $s/m^{1/3}$ in unvegetated and vegetated areas, respectively. Vegetation also drives in situ organic accretion, which is modeled as a parabolic function of elevation (Mariotti and Zapp, 2022; Morris et al., 2002), with a maximum accretion rate set equal to 5 mm/yr.

Settling velocity is equal to 0.2 mm/s in both the vegetated and unvegetated areas since net sediment deposition on the marsh is already triggered by the reduction in bed shear stress. This low value of settling velocity on the marsh platform is consistent with field estimates (Duvall et al., 2019; Lacy et al., 2020). Critical shear stress (τ_{cr}) for bed sediment erosion is assumed to be equal to 0.2 Pa for both unvegetated and vegetated areas to enable channel incision on the marsh platform. Notably, even with a relatively low value of marsh τ_{cr}, unchanneled flow is not able to induce marsh erosion, which is consistent with observations (Christiansen et al., 2000; Kastler and Wiberg, 1996). The 0.2 Pa critical shear stress is also in range with values measured in the field and used in models, generally between 0.05 and 1 Pa (Mengual et al, 2017). For small water depths (h<0.5 m), the maximum bed shear stress by tidal currents is set equal to 1 Pa in order to filter out unrealistic erosion.

Ponding dynamics include pond formation by random seeding and impoundment, pond lateral expansion and deepening, and pond merging (Mariotti, 2020; Mariotti et al., 2020). Wind waves are calculated based on fetch, water depth, and wind speed (Young and Verhagen, 1996). Waves contribute to both bed sediment resuspension and marsh edge erosion. The latter is directly proportional to the wave power at the edge and implemented through a probabilistic approach (Mariotti and Canestrelli, 2017), assuming a proportionality coefficient of 0.2 $my^{-1}/(Wm^{-1})$ and a fixed fraction of oxidized material equal to 25%, assuming that on average 50% of the marsh volume is composed of organic material (e.g., Mariotti et al., 2020), and that 50% of the material is oxidized. Bed shear stress by waves – and more importantly, its associated sediment resuspension – is set equal to zero in the vegetated marsh, which is consistent with various field observations (e.g., Christiansen et al., 2000; Kastler and Wiberg, 1996). Hence, marshes cannot be eroded by waves from above, even though they can be eroded by waves laterally at their edge.

2.2 Rationale for simulating stream meandering in tidal marshes

Compared to rivers, meandering in tidal streams might be perhaps treated more simply. Sediments are typically fine and cohesive and are mostly transported in suspension (Finotello et al., 2019a), such that the cross-stream bedload is not as important as it is for rivers. Hence, we argue that the main feature that needs to be reproduced is the lateral shift in flow momentum, as opposed to the entire cross-stream secondary circulation. We suggest a middle-ground approach to recreate this effect by developing a simplified model based on the curvature of the flow rather than of the banks, which bears similarities to Nicholas's (2013) model for meandering rivers.

A modification of the base tidal flow field is then needed to induce two basic mechanisms: erosion at the outer concave bank (i.e., bank pull) and accretion at the inner convex bank (i.e., bar push) (e.g., Eke et al., 2014; van de Lageweg et al., 2014). Both mechanisms are critical, especially in tidal channels where banks do not necessarily move parallel to each other as a result of potential mutually evasive paths followed by the ebb and flood flows (Ahnert, 1960). Eroding the outer bank is needed to move the meander, whereas accreting the point bar at the inner bank is necessary to maintain the channel width in dynamic equilibrium and perpetuate channel meandering.

Special emphasis is given to formulating the model in a manner that enables its execution at coarse spatial resolutions, that is, for channels that have a width of just a few computational cells. Because channel width varies continuously, any marsh model would necessarily include channels that are one cell wide (unless spatial resolutions smaller than the smallest possible channels, on the order of 1 m, are used). When using coarse resolution (5-10 m), a large fraction of channels could still be only a few cells wide. Hence, it is important that the meandering model also works in this case.

2.3 Hydrodynamic model for curved flows

The curvature-driven modification should theoretically differ between ebb and flood (Finotello et al., 2020c; Gao et al., 2022). However, given that the model computes the same flow field for both ebb and flood ($\boldsymbol{U}$) (Eqs. 1,2), any difference in curved flow between ebb and flood is likely a second-order effect compared to the assumptions already present in the model. Hence, to retain an appropriate level of complexity, we propose an empirical modification to the main flow field that is identical for both ebb and flood.

The first step to account for the curvature effect is to identify the active flow region (Nicholas, 2013). For simplicity, we assume that the active flow region coincides with the unvegetated cells, although a distinction could also be made based on water depth and flow velocity. All other cells are considered hydrodynamically inactive. Inactive cells adjacent to active cells are considered bank cells. In practice, the flow domain coincides with the area lying below the lower limit for vegetation growth, which includes both the main channels and the mudflats.

The second step is to calculate, within the active flow region, the flow curvature strength (Φ):

$$\Phi = \left[\nabla x \widehat{\boldsymbol{U}}\right] |\boldsymbol{U}| h \tag{3}$$

where $\widehat{\boldsymbol{U}}$ is the velocity versor equal to $\boldsymbol{U}/|\boldsymbol{U}|$. The curvature is spatially smoothed along the cross-flow direction (i.e., perpendicular to the vector $\boldsymbol{U}$) within the active flow domain, with the purpose of creating a relatively uniform curvature field across the channel (cf. Lazzarin and Viero, 2023). The rationale of this step is that the cross-stream flow (responsible for carrying flow momentum outward) is produced as the result of a force balance within the whole channel cross-section (Solari et al., 2002). In addition, this smoothing tends to remove the curvature in channels that are not bounded by banks (i.e., mudflat channels). Notably, the curvature is not averaged in the along-stream direction to avoid the suppression of short wavelength instabilities that are the likely culprits of meander development (Seminara, 2006).

The third step is to use the flow curvature strength as a proxy for the modification of the flow field. Qualitatively, curvature tends to shift water flows from the inner to the outer meander bank. This effect is simulated by the parameter f:

$$f = min\left(1, \widehat{\boldsymbol{U}}\ \left[I + \nabla \cdot \boldsymbol{T}\right]^{-1}\right) \tag{4}$$

where $\boldsymbol{T} = a\Phi^{b}\widehat{\boldsymbol{U}}^{\perp}$ is a vector perpendicular to the main velocity $\boldsymbol{U}$, a and b are calibration parameters, and I is the identity matrix. The parameter f can be thought of as the velocity versor $\widehat{\boldsymbol{U}}$) being advected laterally by a fictitious flow ($\boldsymbol{T}$) proportional to the flow curvature strength. In the absence of such lateral advection, the parameter f is everywhere equal to one. When the flow is in a divergent zone of the lateral advection $\boldsymbol{T}$ (e.g., the meander inner bank), f is smaller than one, whereas when the flow is in a convergent zone (e.g., the meander outer bank), the parameter is larger than one (Figure 2). The a and b parameters are empirical calibration factors that cannot be derived from first principle. We found that a value of b smaller than one (e.g., equal to 0.5) provided a realistic flow field. The parameter a was calibrated by trial-and-error to obtain a modification of the flow field on the order of 10-50%, and a value equal to 250 was selected for all simulations.

The final step is to recalculate the velocity within the whole domain by introducing f in the momentum equation:

$$\boldsymbol{U} \propto f\frac{h^{4/3}}{n^2}\nabla\eta \tag{5}$$

and solving Eqs. 1 and 5 together. In practice, including the parameter f decreases the bed friction in areas with f>1 and increases it in areas with f<1. Compared to the flow calculated without the curvature correction (Eqs. 1-2), the discharge passing through the cross-section will increase in areas with f>1 and it will decrease in areas with f<1. Notably, the factor f only modifies the bed friction, and since the continuity equation (Eq. 1) is recalculated, mass conservation is automatically satisfied.

Preliminary tests indicated that, for complex geometry such as the intersection of channels, positive values of f occurred at unrealistic locations. On the other hand, negative values of f always occurred in the interior of bends. Furthermore, positive values of f (at bend outer edges) were highly sensitive to grid resolution, whereas negative values of f (at bend inner edges) were not. We found that the flow slowdown at the inner part of the bend (where f<1) sufficed to create a realistic flow over the whole domain. This is because velocity reduction at the inner bend automatically enhances velocities at the outer bank, given that Eqs. 1 and 5 recalculate the flow in the whole domain. As such, values of f greater than one are set equal to one.

2.4 Curvature-induced outer bank erosion (bank pull)

The model includes two distinct mechanisms for bank erosion driven by curved flows. The first mechanism originates from a modification of the bank erosion process that is already present in the model, even without curvature correction. Flow is deflected to the outer banks, which increases bed scour in the proximity of the bank. This renders the bank steeper and consequently enhances elevation diffusion by soil creep (Mariotti et al., 2016, 2019). This mechanism phenomenologically represents bank collapse mediated by gravity (i.e., mass wasting), and automatically reproduces the observed effect that bank retreat increases with bank height (Zhao et al., 2019, 2022). In reality, bank collapse is mediated by the water table and hence the tidal stage (Gasparotto et al., 2022; Mariotti et al., 2019; Zhao et al., 2019), but given that the model does not explicitly simulate different tidal stages, this dependence is not considered.

The second mechanism entails bank erosion solely caused by the impact of curved flow on the bank, without any influence from the bed slope or bank height. Conceptually, this mechanism involves an active stripping of sediment from the bank, rather than a passive mass-wasting process. The rate of lateral bank erosion is set proportional to the flow curvature strength (Φ) at the outer bank:

$$E = k\Phi \tag{6}$$

This conceptualization is similar to previous centerline migration models used for meandering rivers (e.g., Bogoni et al., 2017; Frascati and Lanzoni, 2009; Seminara et al., 2001; Zolezzi and Seminara, 2001) and tidal creeks (e.g., Gong et al., 2018; Kleinhans et al., 2009), where bank erosion is linked to the near-bank excess velocity, the latter being however driven by the curvature of the channel axis rather than the flow. The coefficient of proportionality k is considered to be dependent on the soil strength and, more broadly, on the whole suite of hydro-mechanical processes that contribute to bank erosion. At this stage, however, k cannot be quantitatively linked to soil properties and is thus treated as an empirical parameter. Since banks are identified as vegetated cells adjacent to a non-vegetated cell (see Section 2.3), bank erosion due to curved flows (Eq. 6) only occurs along salt-marsh channels where banks are vegetated. In theory, this mechanism could also be implemented in non-vegetated banks, which would, however, require defining the flow domain and the banks based on characteristics other than vegetation (e.g., a velocity threshold).

Operationally, lateral erosion predicted by Eq. 6 is often implemented through a reservoir method: partial erosion of a cell is stored until it reaches the size of a cell, at which point the entire cell is eroded by reducing its elevation (Nicholas, 2013). Here the lateral erosion is implemented with a probabilistic algorithm akin to that used to simulate wave-induced edge erosion (Mariotti and Canestrelli, 2017). This does not imply that bank erosion is a stochastic process but rather serves as a numerical approach to simulate lateral erosion in a domain represented by fixed cell locations and elevations. Once a cell has been eroded, a given portion of the material (25% in this case) is assumed to be removed through oxidation, while the remaining portion is redistributed among the neighboring cells.

2.5 Curvature-induced inner bank deposition (bar push)

Inner bend deposition is automatically reproduced in the model. Indeed, the curvature-induced modification of the basic flow field (Eq. 5) not only increases flow velocity at meander outer banks, but also reduces it at inner banks, thus promoting sediment deposition. This latter process does not require any ad hoc formulation, as it rather emerges from the sediment transport mechanism already present in the model.

The model also automatically includes a topographic steering effect (Lancaster and Bras, 2002). Indeed, the first-order flow already tends to allocate more flow where channels are deeper. Thus, if the scouring tends to occur toward the outer bank and deepens that part of the channel, more flow would be conveyed through that section by the first-order flow model. Then, the deposition in the inner bank increases the flow on the outer bank, which in turn increases the outside bank erosion through both mechanisms.

2.6 Representative numerical simulations

Tests of the hydrodynamic model were performed considering a simplified and fixed geometry. We considered an idealized tidal channel with a sinusoidal planform (amplitude of 150 m and a wavelength of either 500 or 1000 m), a constant width of 40 m, and a constant depth of 3 m (Figure 2). For simplicity, the tidal range was imposed equal to zero, and a constant tidal prism was imposed at the landward end so that the discharge was constant along the channel.

We also performed some idealized simulations inspired by two natural salt marshes (Figure 1): Barnstable (Massachusetts, USA) and Bishops Head (Maryland, USA). The water level and wind inputs for each site are calculated as in Mariotti and Zapp (2022). Briefly, this accounts for the temporal variability in tidal range and sea level anomalies. Barnstable has a 10^{th} and 90^{th} tidal range of 2.21 and 3.63 m, a 10^{th} and 90^{th} sea level anomaly of -0.14 m and 0.13 m, and an upper limit for vegetation growth equal to 2.16 m above mean sea level. Bishops Head has a 10^{th} and 90^{th} tidal range of 0.41 and 0.67 m, a 10^{th} and 90^{th} sea level anomaly of -0.19 m and 0.19 m, and an upper limit for vegetation growth equal to 0.7 m above mean sea level. Both marshes have a semidiurnal tide, with a period of 12.5 hours. In both cases, we consider a marsh that is 2 km wide and 3 km long, with an additional 1 km long mudflat in front of it and an additional fetch of 1 km (Figure 1). For the Barnstable case, we considered a characteristic suspended sediment concentration of 20 mg/L and a rate of relative sea level rise of 2.5 mm/yr, whereas for Bishops Head we considered a characteristic suspended sediment concentration of 40 mg/L and a rate of relative sea level rise of 3.5 mm/yr, in agreement with field measurements (Wasson et al., 2019).

The reference simulation for both Barnstable and Bishops Head analogs used a grid resolution dx=10 m, the flow correction to account for the effect of curvature (Eqs. 4,5), and a bank erosion coefficient k=0.2. Five additional simulations were conducted to test the model sensitivity by varying the grid resolution (to dx=5 m and dx=20 m, respectively), the value of the bank erosion coefficient (assuming k equal to 0 and 0.5), and by deactivating the flow correction based on curvature. A fixed computation timestep dt=0.5 years was used in all the simulations, with a total duration equal to 3000 years. The standard simulations (with a grid resolution dx=10 m) took about 2 hours on a 3.2 GHz single-core processor.

The results of numerical simulations were analyzed to quantify the morphometric characteristics and planform dynamics of numerically modeled tidal channels. Details of this procedure are described in the Supplementary Information.

3 Results

3.1 Model Performance and Sensitivity

Simulations performed on individual meandering channels with constant depth demonstrate how the model provides a realistic curvature-induced modification of the flow (Figure 2). Without such modification, the highest velocities are consistently found at inner convex banks, as expected by the balance between pressure gradient and bed friction. Indeed, the flow path is shorter in the inner than in the outer part of meander bends, resulting in larger pressure gradients in the former. Given that this gradient is assumed to be fully balanced by friction, and given that the water depth is constant, this is achieved by higher velocities in the inner bend. When the curvature-induced flow modification is included, threads of maximum velocity are shifted toward outer concave banks. Curvature-adjusted velocities are modified by up to 0.1 m/s compared to the base flow field (Figure 2), which is generally on the order of 0.3 m/s. As expected, the flow modification is progressively stronger as the curvature increases (Figure 2). Furthermore, the results suggest that the flow fields are similar regardless of the grid resolution, although higher resolutions yield more detailed results (Figure 3).

The application of flow correction based on curvature yields satisfactory outcomes even when considering complex meandering networks rather than individual sinuous channels, indicating the model's suitability for use in such hydro-morphodynamically complex environments (Figure 2). Importantly, the modified flow does not necessarily produce the highest velocity close to the outer bank, which is generally the case in riverine settings and it is often assumed in simplified models (Kleinhans et al., 2024; Motta et al., 2012). Rather, maximum flow velocities can be found in the middle of the channel or even closer to the inner bank (Figure 2B), typically aligned with the thalweg. These conditions tend to occur in channels that are wide and moderately sinuous (Figure 2B).

The morphodynamic effect of curvature-induced flow modification on the network structure is readily visible when comparing it to cases where the correction is not implemented (Figure 4). Without flow modification, channels are straight to mildly curved, consistent with previous simulations performed using the same numerical model (Mariotti, 2018, 2020) or other models that do not consider curvature-driven flow modifications (e.g., Belliard et al., 2015; Gourgue et al., 2022; Kirwan and Murray, 2007). In contrast, with the inclusion of flow-curvature correction, channels become highly sinuous, and closely resemble those observed at the reference sites (Figure 1).

According to the model, the marsh platform in Bishops Head takes a few hundred years to establish, whereas the marsh platform in Barnstable takes about 1000 years, likely because of the lower sediment supply (Figure 5). In both cases, the model predicts that the channels have a low sinuosity when they initially form on the marsh platform, and that the sinuosity increases over the course of about 1000 years before stabilizing even though the channels keep migrating.

In addition to curvature-driven flow modifications, bank erodibility also plays a critical role in driving marsh morphodynamic evolution (Figure 6). Interestingly, channels can become highly sinuous even without including lateral bank erosion (i.e., k=0). This is because channel banks can retreat and modify through bank creep alone. However, in this case, the lateral mobility of the channels is considerably reduced compared to scenarios where bank erodibility is considered. Overall, increasing the bank erodibility coefficient results in accelerated rates of channel meandering, regardless of channel size and environmental conditions (Figure 6).

3.2 Quantitative analysis of meander morphology and dynamics

The model reproduces not only realistic meander planform shape and cross-sectional morphology , but also (and more importantly) channel lateral migration rates and dynamics – with the progressive erosion of concave outer banks and the formation of point bars on convex inner banks. Specifically, a comparison with existing data (Finotello et al., 2018, 2020a) highlights similar scaling of meander wavelengths, amplitudes, and radii of curvature as a function of meander width for both field and numerically modeled meanders (Figure 7 and Figure S6 in Supporting Information). Close similarity to field data is also observed in terms of channel width-to-depth ratios and cross-sectional shape. A comparison with the numerical results obtained by Solari et al. (2002) also demonstrates that our model is comparably better at reproducing meander wavelengths observed from field data (see SI Text, SI references, and Figures S1 to S4). Notably, similar channel morphometrics are obtained when considering different spatial resolutions (Figures 7 and 8). Meander cross-sections are generally asymmetric and deeper toward the outer bank, which is often the case in fluvial and tidal channels alike (Figure 9). Nonetheless, the channel thalweg is at times located closer to the inner bend (Figure 9), which aligns with field observations (Figure 10). Noticeably, both modeling results and the field observations indicate that this condition is typically associated with meanders exhibiting cuspate inner banks (Figures 9 and 10), which are widely recognized as a diagnostic feature of tidal influence (Finotello et al., 2020c; Hughes, 2012; Woodroffe et al., 1989) (Figure S7 in Supporting Information).

In terms of meander planform dynamics, the reference simulations produce migration rates M_R= 0.039±0.14 m/yr for Barnstable and M_R=0.058±0.18 m/yr for Bishops Heads (Figure 8). For comparison, the migration rates at those sites, estimated through aerial images considering the apex migration of about 30 meanders in the last 40 years, are on the order of 0.01-0.1 m/yr. Furthermore, width-adjusted rates of lateral meander migration (M_R^*=M_R/W) are equal to 0.13±0.54 % yr^{-1} at Barnstable and 0.23±0.77 % yr^{-1} at Bishops Head (Figure 8), which broadly aligns with M_R^* observed in other tidal settings worldwide (Finotello et al., 2018; Gabet, 1998; Garofalo, 1980; Jarriel et al., 2021). Model results show that migration rates increase with increasing bank erosion coefficient (k). Numerical simulations carried out without curvature-flow correction exhibit the lowest migration rates, while the highest rates are observed when factoring in curvature correction and assuming a bank erodibility coefficient k=0.5 (Figure 6). Different grid resolutions result in remarkably similar migration rates (Figure 8).

4 Discussion

4.1 Model Performance and Comparison to Field Sites

The proposed model reproduces the ecomorphodynamic evolution of tidal marshes dissected by complex, branching, and meandering networks of tidal channels. Specifically, the model replicates meander cutoff (which is driven by intra-channel interactions) and channel piracies (which is driven by inter-channel interactions), both of which are observed in real marshes (Figure S8 in Supporting Information).

Unlike most centerline migration models for meandering rivers – for which banks invariably move parallel to each other – our model allows channel banks to evolve independently. As a consequence, our model replicates the characteristic funneling of tidal channels, resulting from the progressive landward reduction of the tidal prism (Finotello et al., 2020a; Lanzoni and D'Alpaos, 2015). In addition, our model is more flexible than the few existing models for channel migration that allow the inner and outer banks to move independently (e.g., Eke et al., 2014; Lopez Dubon and Lanzoni, 2019; Parker et al., 2011; Zhao et al., 2021). Hence, our model recreates non-trivial meandering morphologies, such as meander bends characterized by cuspate inner banks (Figure 9 and Figures S7 and S8 in Supporting Information). Although these morphologies have traditionally been associated with mutually evasive paths followed by the ebb and flood flows (Ahnert, 1960; Dalrymple et al., 2012; Hughes, 2012), our model does not differentiate between ebb and flood. Therefore, the emergence of cuspate meanders from our numerical simulations suggests that the formation of cuspate bends may not necessarily be related to the above-described mechanism.

The gradual increase in channel sinuosity over time (Figure 5, S3) supports the hypothesis that mature marshes feature more extensive and sinuous tidal channels compared to juvenile marshes (Allen, 2000; Pethick, 1969), which is also supported by field examples. For example, the Barnstable marsh, which formed ca. 4000 years ago (Redfield, 1972), hosts a highly sinuous channel network (Figure 1). In contrast, the Saeftinghe marsh (Western Scheldt Estuary, The Netherlands), which has a similar tidal range (about 4 m) but was only formed in the last 300 years, features lower channel sinuosities (Jongepier et al., 2015). On the other hand, previous studies showed that sinuosity might also be inherited from pre-existing sinuous mudflat channels that are progressively colonized by halophytic vegetation as marshes expand laterally (Belliard et al., 2015; Pestrong, 1972; Schwarz et al., 2014). Hence, we propose that elevated channel sinuosity does not always indicate marsh maturity, whereas the presence of meander cutoffs and channel piracies (Figures S7, S8, and S9 in Supporting Information) might serve as more reliable indicators.

4.2 Curvature-driven Meander Dynamics

Although many factors have been proposed to control meander migration, previous studies emphasized the prominent role of curvature in determining not only migration rates but also the evolution of meander patterns in both tidal and fluvial landscapes (Fagherazzi et al., 2004; Finotello et al., 2018, 2022; Hooke, 2013; Lagasse et al., 2004; Sylvester et al., 2019). It is generally accepted that migration rates increase nonlinearly with curvature and tend to saturate when the radius of curvature becomes larger than 2-4 times the channel width (i.e., for width-adjusted radii of curvature R^*=R/W>2-4; Finotello et al., 2019; Hickin and Nanson, 1975; Hooke, 2013). These nonlinearities between migration and curvature are believed to arise from two contrasting effects. On the one hand, stronger channel curvature enhances both the secondary flow and the phase lag between curvature and near-bank velocity (Crosato, 2009; Lanzoni and Seminara, 2006; Parker et al., 1983; Seminara et al., 2001), resulting in more pronounced outer-bank erosion and migration. On the other hand, in sharp bends, the increase in curvature-induced effects is limited by the growth of hydrodynamic nonlinearities, such as saturation of secondary flow, enhanced secondary outer-bank flow cells, and flow separation at the outer bank (Blanckaert, 2011; Finotello et al., 2020c; Hooke, 2013; Lazzarin and Viero, 2023), all of which contribute to slowing down bank erosion as curvature increases.

The relationships between modeled width-adjusted migration rates (M_R^*) and channel curvature (expressed through R^*= R/W) show positive correlations that tend to saturate for increasing values of R^* (Figure 8), in accordance with empirical observations. Since the model does not directly account for flow nonlinearities described above, other mechanisms are likely to explain the saturation of M_R^* at large curvature predicted by the model. First, curvature strength (Φ) affects the basic flow field in a nonlinear fashion, with curvature-induced flow modification being proportional to $\Phi^{0.5}$ (see Eq. 4). Second, some of the sediment eroded from the bank deposits in close proximity to the bank itself (see section 2.4), acting as a sheltering barrier and reducing the pace at which the bank retreats laterally (e.g., Eke et al., 2014; Motta et al., 2014; Posner and

Duan, 2012). Indeed, field observations show that collapsed bank soil in salt marsh creeks could persist for several years (Fagherazzi et al., 2004), contributing to the channel erosion paradox (Gabet, 1998) whereby marsh creeks are likely to migrate laterally at a quite slow rate despite the widespread occurrence of collapsed bank blocks. Third, migration of the channel entails not only bank erosion but also channel thalweg migration, which occurs through bed erosion. Given that this process is not controlled by the flow curvature, it tends to slow down the overall migration for large flow curvatures.

Variability in bank erodibility is thought to explain the large scatter present in the curvature-migration relationship (Hooke, 2013; Nanson and Hickin, 1986; Schwendel et al., 2015). Despite the model having a spatially constant bank erodibility, however, there remains a considerable scatter between width-adjusted migration rate and channel curvature (Figure 8). This is explained by considering that the channel curvature is only a proxy for the process driving channel migration, the real cause being the curved flow. For a given channel width and curvature, the flow can be different depending on the channel morphology (i.e., the depth and the position of the thalweg) and on the tidal prism passing through the cross-section. Indeed, because of the frequent reorganization of the tidal network, a channel cross-section might suddenly experience an increase or decrease in tidal prism, and hence change its migration rate.

Grid resolution can also alter the curvature-migration relationship by both affecting the numerical computation of curvature (Crosato, 2007) and setting the lower bound for migration rates that can be detected. Our analyses revealed that numerical simulations conducted with a grid resolution of dx=20 m yield poor correlations between M_R^* and R^*(especially for the Bishops Head case), suggesting that there exists a critical dx threshold beyond which the model struggles to accurately compute curvature and replicate its morphodynamic effects. The upcoming section provides a comprehensive discussion of the implications of grid resolution on modeled channel morphology and dynamics.

4.3 Effects of Computational Grid Resolution

The introduction of curvature-based flow modification leads to an increase in velocity near the outer bank and a decrease in velocity near the inner bank (Figure 2). Consequently, this flow modification is not present in channels that are only one cell wide, where f=1 by definition (see Eq. 5). As a result, if the model creates channels that are one cell wide, these channels experience virtually no curvature-induced migration (except where they form 90 degree angles).

This limitation highlights the importance of selecting an appropriate grid resolution for numerical simulations. In the case of Barnstable, where most channels are wider than 10 m, resolutions $dx \leq 10$ m are suitable for accurately representing the overall channel network structure (Figure 11). With a resolution dx=20 m, in contrast, small channels become only one cell wide and thus do not exhibit significant migration (Figure 11). Nevertheless, the channels in the main network branches are still replicated reasonably well. For the Bishops Head case, which features narrower channels than Barnstable due to a smaller tidal range, a spatial resolution dx=20 m fails to reproduce most of the channels. Because the largest channels at Bishops Head are approximately 20 m wide, even the channels in the main network branches experience minimal migration and, as a result, exhibit poor sinuosity (Figure 11).

To verify that the lack of migration in the Bishops Head case is due to reduced channel width rather than other factors (e.g., tidal range) or model shortcomings, an additional simulation was performed using a 10 km long domain, and replicated twice using a resolution dx=10 m and *dx*=20 m, respectively (Figure 12). This extended domain increased the tidal prism in the seaward portion of the marsh, allowing for the formation of wider channels (Figure 12). Even with *dx*=20 m, the model successfully replicated channel migration in the most seaward 3 km of the marsh, where channels were approximately 100 m wide. Yet, in the more landward regions, tidal channels remained narrower than 20 m and did not exhibit significant migration (Figure 12), thus confirming our initial hypothesis.

4.4 Model Limitations and Future Developments

Modification of hydrodynamics in the model is currently implemented using an empirical approach. Future research could explore the possibility of deriving a simplified flow correction from a theoretical perspective, and to extensively compare the simplified model with models that fully resolve the 3D flow (e.g., Fig. S5 in Supporting Information). This would contribute to enhancing the accuracy and reliability of the model in simulating meander migration processes.

The model utilizes a base flow field that does not account for tidal asymmetries. However, asymmetries in tidal flow velocities critically affect meander planform evolution by governing net in-channel sediment transport (Dronkers, 1986; Finotello et al., 2019a; Tambroni et al., 2017). Specifically, tidal meanders develop distinct depositional patterns based on the dominant flow direction, with flood-dominated flows leading to the formation of point bars upstream of bend apexes, and vice versa for ebb-dominated flows (Ghinassi et al., 2018; Tambroni et al., 2017). Although the model is capable of reproducing strongly skewed meander bends characterized by an asymmetry index ($\mathcal{A}$) significantly different from zero, the probability distributions of $\mathcal{A}$ for numerically modelled bends differ from the field data (Figure S6 in Supporting Information). This is however a limitation of marginal importance. While local variations in tidal asymmetries can impact meander planform over relatively short timescales (i.e., decades), the overall evolution of the network structure on timescales ranging from centuries to millennia is more prominently influenced by inter and intra-channel dynamics (Figures S7 and S8 in Supporting Information). This is because meander cutoffs and channel piracies can lead to significant reconfigurations of the network structure and related hydrological connectivity, and our model serves as a computationally efficient tool to replicate such dynamics.

Asymmetries in tidal flows can also lead to mutually evasive paths of ebb and flood currents, resulting in segregated ebb and flood channels separated by mid-channel bars (e.g., Leuven and Kleinhans, 2019; Robinson, 1960; Shimozono et al., 2019) (Figure 10). These hydrodynamic nonlinearities, together with flow separation in sharp bends (Finotello et al., 2020c; Leeder and Bridges, 1975), are currently not accounted for. As a consequence, the model may generate narrower channels near the marsh seaward margin, which deviate from the channel characteristics observed in field studies (e.g., Barnstable; see Figure 1). These limitations might also contribute to explaining the less pronounced saturation of migration rates beyond a critical $R*$ threshold compared to empirical field data, as well as the slightly shorter wavelengths of modeled meanders in comparison to those observed in nature (Figure 7). It is worth noting that previous models attempting to extend the theory of river bend instability to tidal meandering channels also encountered similar shortcomings (Solari et al., 2002).

Bank erodibility is currently calibrated to reproduce realistic rates of lateral migration. It would be interesting to explore in detail the relationship between the bank migration parameter (k) and soil properties such as bulk density, organic matter content, and vegetation root characteristics. In addition, a more realistic implementation of bank collapse – here simulated as a diffusion process – might be implemented. For example, the geotechnical forces acting on the bank could be estimated at different tidal stages (which would require introducing a simplified model for tidal stage variability), and the resulting mode of bank failure could be predicted. Then, simple criteria based on the ratio between bank height and near-bank water depth (Zhao et al., 2019) could be implemented to predict the rate of bank collapse. Alternatively, these criteria for bank collapse could be used to empirically adjust the bank creep coefficient.

A combination of modeling and field investigations can provide insights into the variation of bank erodibility (associated with flow-induced erosion or mass collapse) among different marshes. Bank erodibility varies with heterogeneous marsh platform sedimentology, including fine-grained deposits sequestered in meander cutoffs and abandoned channels (Güneralp and Rhoads, 2011). Hence, future modeling efforts should aim at enabling the autogenic capacity of tidal channels to modify the surrounding tidal platforms as they undergo lateral migration (e.g., Bogoni et al., 2017). In addition, future model development should consider the case of multiple sediment classes, e.g., both mud and sand, which would be particularly important to simulate sandy point-bar deposits.

5 Conclusions

We developed a novel numerical model to simulate channel meandering in tidal marshes. The key feature of the model is to correct the basic hydrodynamic field based on the curvature of the flow rather than the curvature of the bank. The model realistically recreates erosion of the outer bank (bank pull) and deposition in the inner bend (bar push). The model is highly numerically efficient and allows for simulating the ecomorphodynamic evolution of complex, branching and meandering networks of tidal channels over timescales ranging from centuries to millennia.

Our model represents a significant advancement in the study of tidal marshes as it enables the investigation of periodic reorganization in tidal channel networks, and hence marsh hydrological connectivity, at practically relevant spatial and temporal scales. This could be of particular importance for marsh conservation and restoration efforts.

Model applications to field and synthetic case studies will improve our understanding of the relationships between function and form in tidal channel networks. For instance, the model will facilitate the exploration of how lateral migration and meandering of tidal channels influence marsh ecology (e.g., vegetation), carbon and nutrient cycling, sedimentary dynamics, and stratigraphy.

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Open Research

The numerical model source code is freely available at Mariotti (2024).

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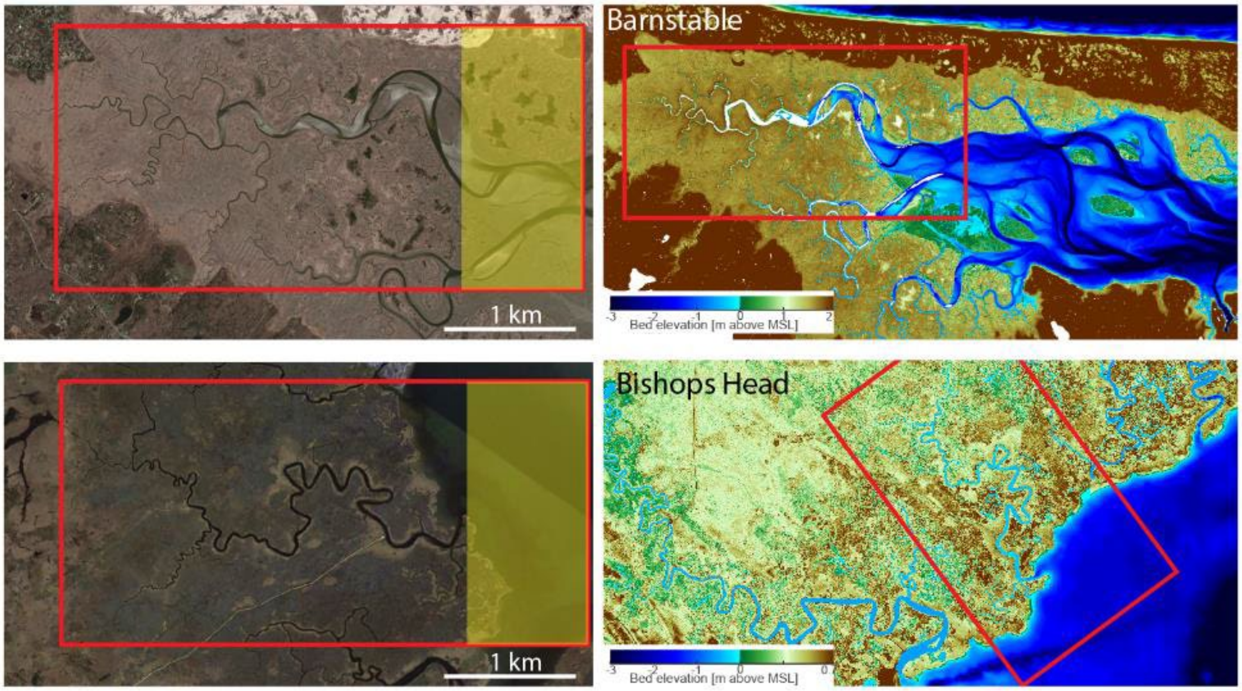


Figure 1. Aerial images and topobathymetric maps of salt marshes found in Barnstable (Massachusetts, USA) and Bishops Head (Maryland, USA). The red rectangles indicate the size of the computational domain used in numerical simulations, which also includes the adjacent mudflat highlighted in yellow. Note that channel depths in Bishops Head are underestimated as they are based on elevations of the water surface rather than of channel bed.

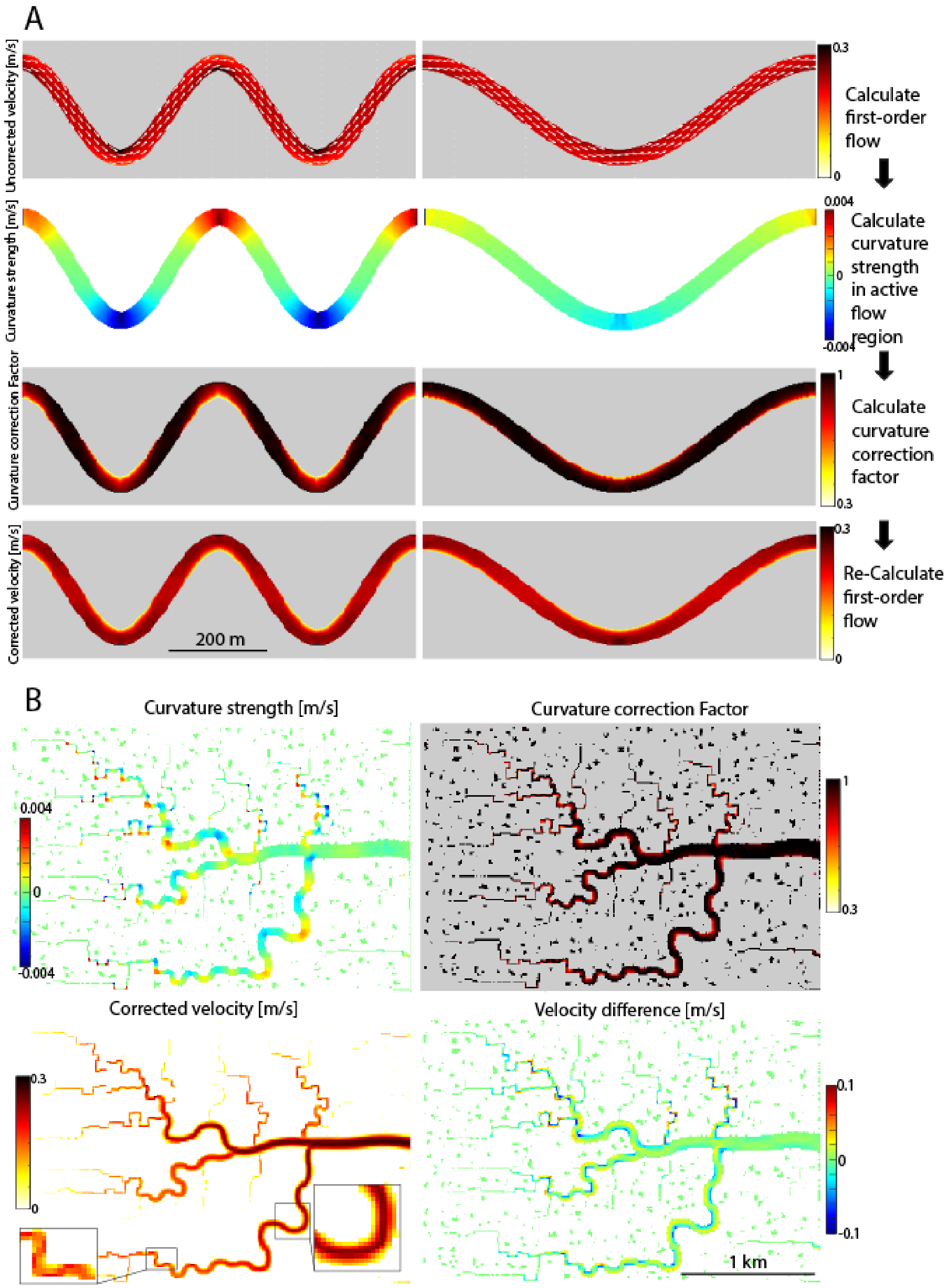

A
Uncorrected velocity [m/s]
Calculate first-order flow
0.3
0
Curvature strength [m/s]
Calculate curvature strength in active flow region
0.004
0
-0.004
Curvature correction Factor
Calculate curvature correction factor
1
0.3
Corrected velocity [m/s]
Re-Calculate first-order flow
0.3
0
200 m
B
Curvature strength [m/s]
0.004
0
-0.004
Curvature correction Factor
1
0.3
Corrected velocity [m/s]
0.3
0
Velocity difference [m/s]
0.1
0
-0.1
1 km

Figure 2. Example of the calculated flow field. A) Application of the model to a regular sinusoidal channel characterized by an amplitude of 150 m, a constant width of 40 m, a fixed depth of 3 m, and a wavelength of either 500 m (left column) or 1000 m (right column). The resolution of the computational grid is equal to *dx*=2.5 m. Moving from top to bottom, each panel illustrates a different computational step to account for curvature-induced hydrodynamic adjustments, as indicated by the legend to the right. Note that the curvature correction also modifies the flow direction, but the changes in direction are small and hence the flow vectors are not reported in the re-calculated flow. B) Application of the model in the case of a morphologically complex tidal channel network simulated by the model, similar to the one depicted in Figure 4A, using a tidal range equal to 3 m.

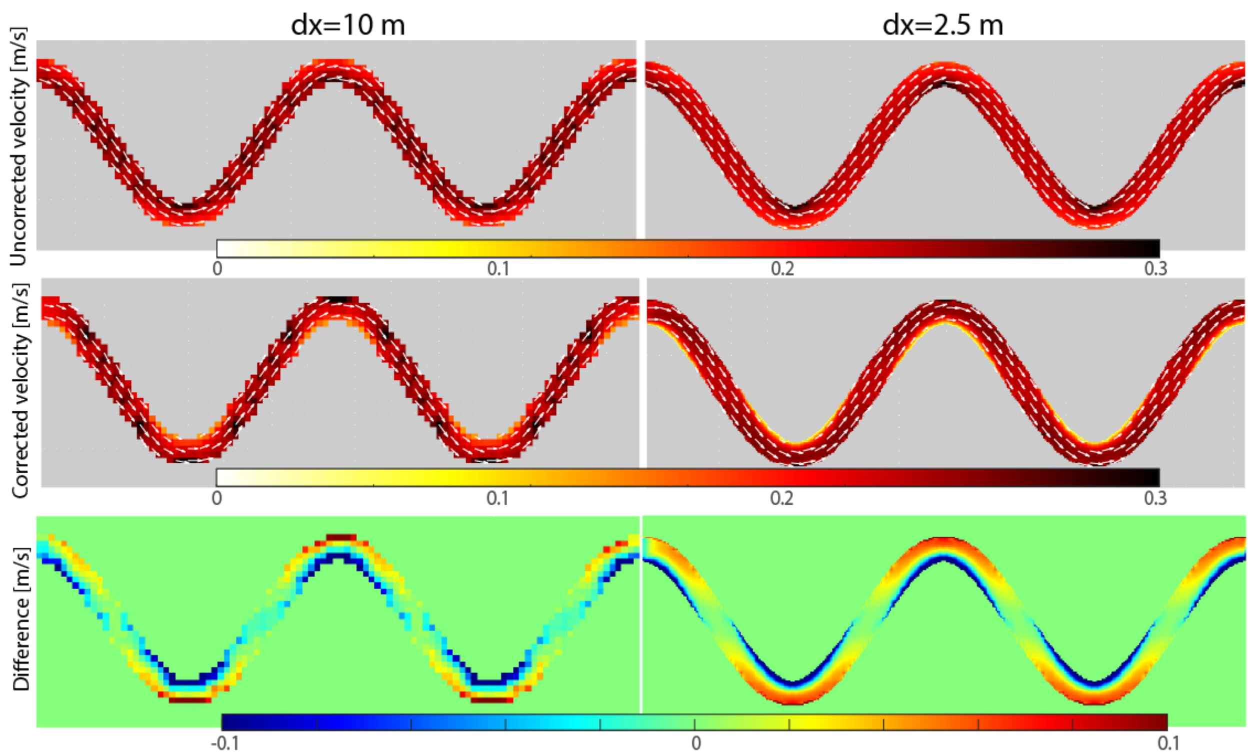


Figure 3: Comparison of the flow in a regular sinusoidal channel with and without curvature-driven flow modification. Results are reported for the same geometry and with two grid resolutions equal to *dx*=10 m and *dx*=2.5 m.

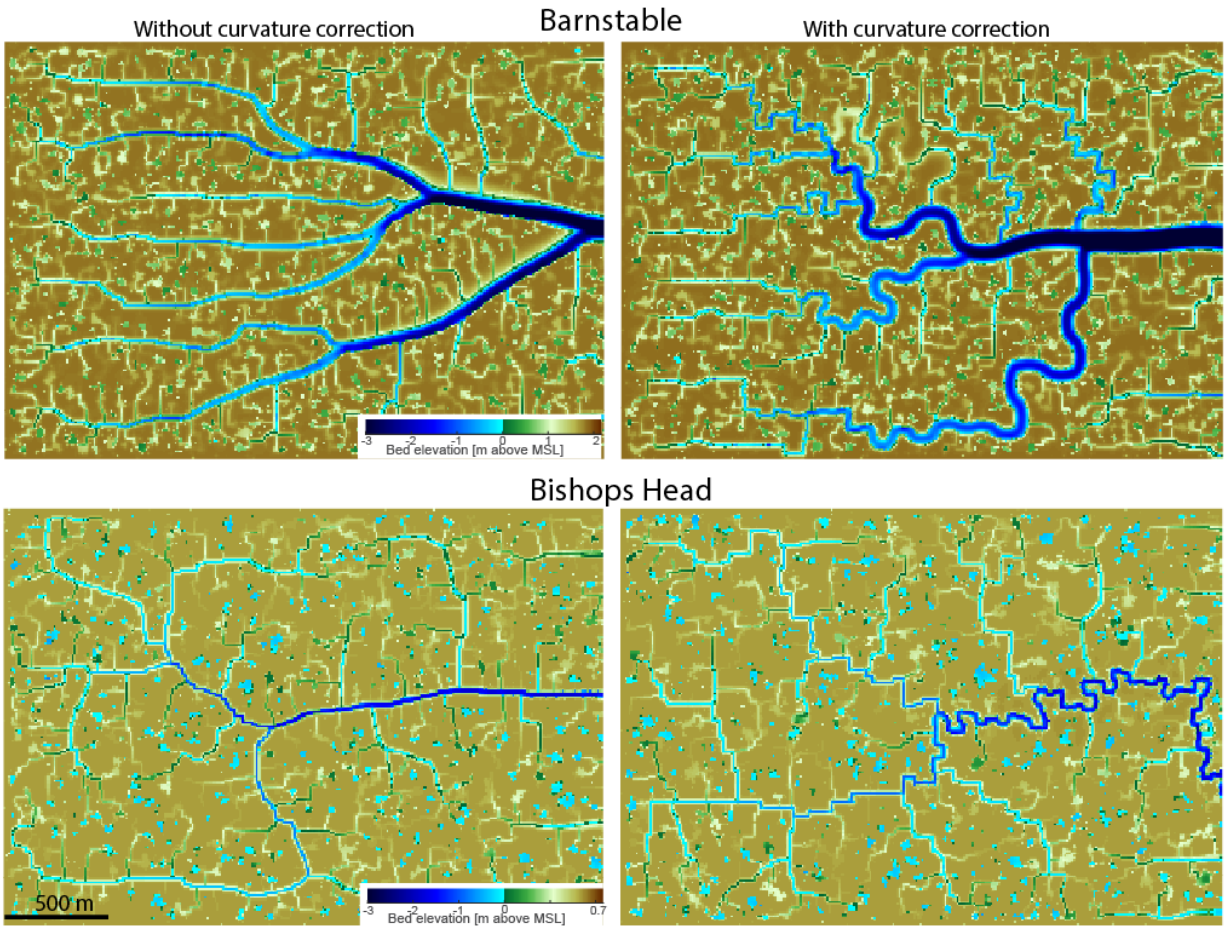
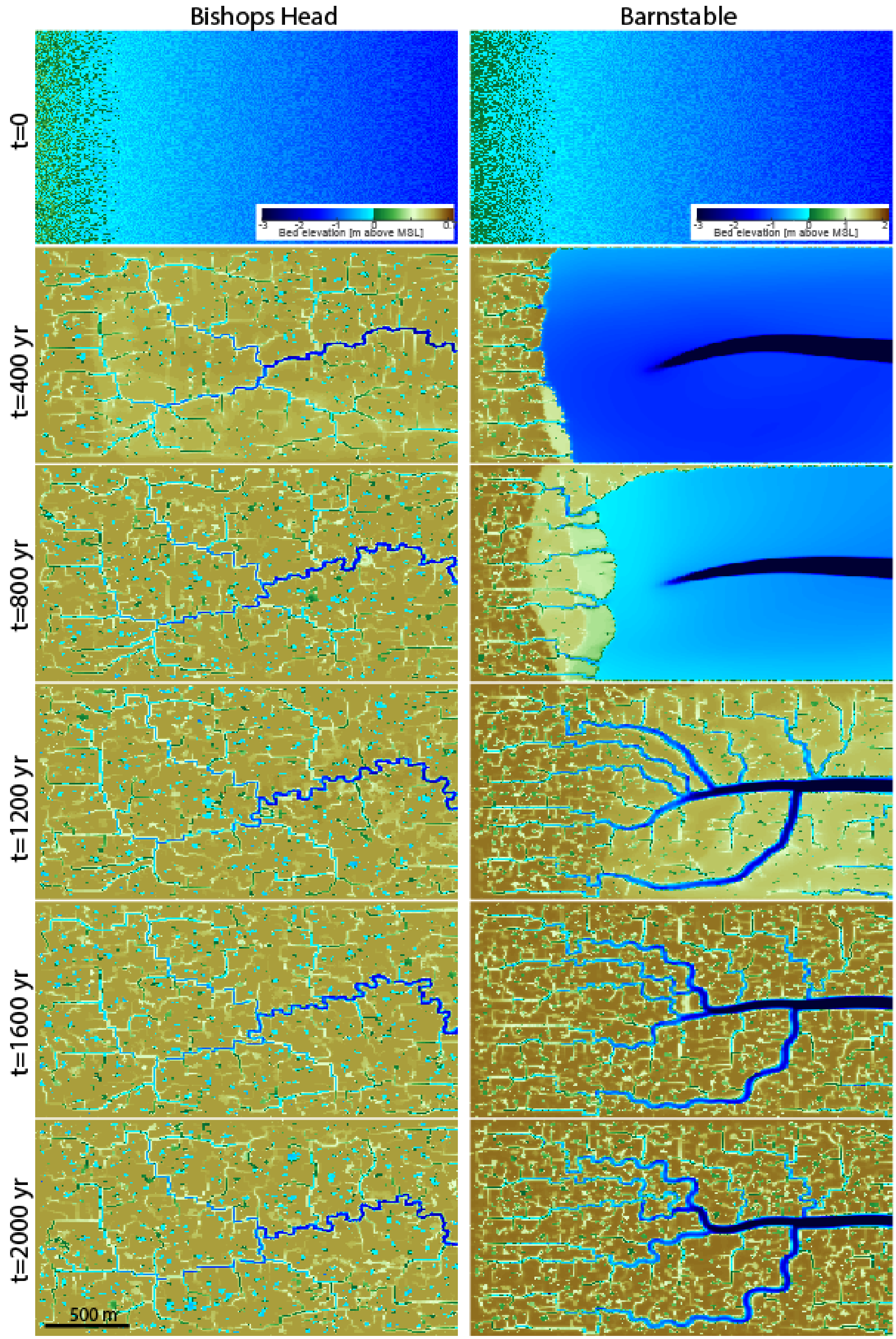


Figure 4. Comparison of numerically modeled topobathymetry with and without the curvature-driven flow modification, at t=3000 yr, for the reference simulation (k =0.2, dx=10 m). 1027 The domain is the same as the red rectangle in Figure 1.



Bishops Head
Barnstable
t=0
t=400 yr
t=800 yr
t=1200 yr
t=1600 yr
t=2000 yr
Bed elevation [m above MSL]
500 m

Figure 5. Numerically modeled topobathymetry at different times for the reference simulation (k=0.2, dx=10 m) in the case of Bishops Head and Barnstable. Note that channel sinuosity increases over time. The domain is the same as the red rectangle in Figure 1.

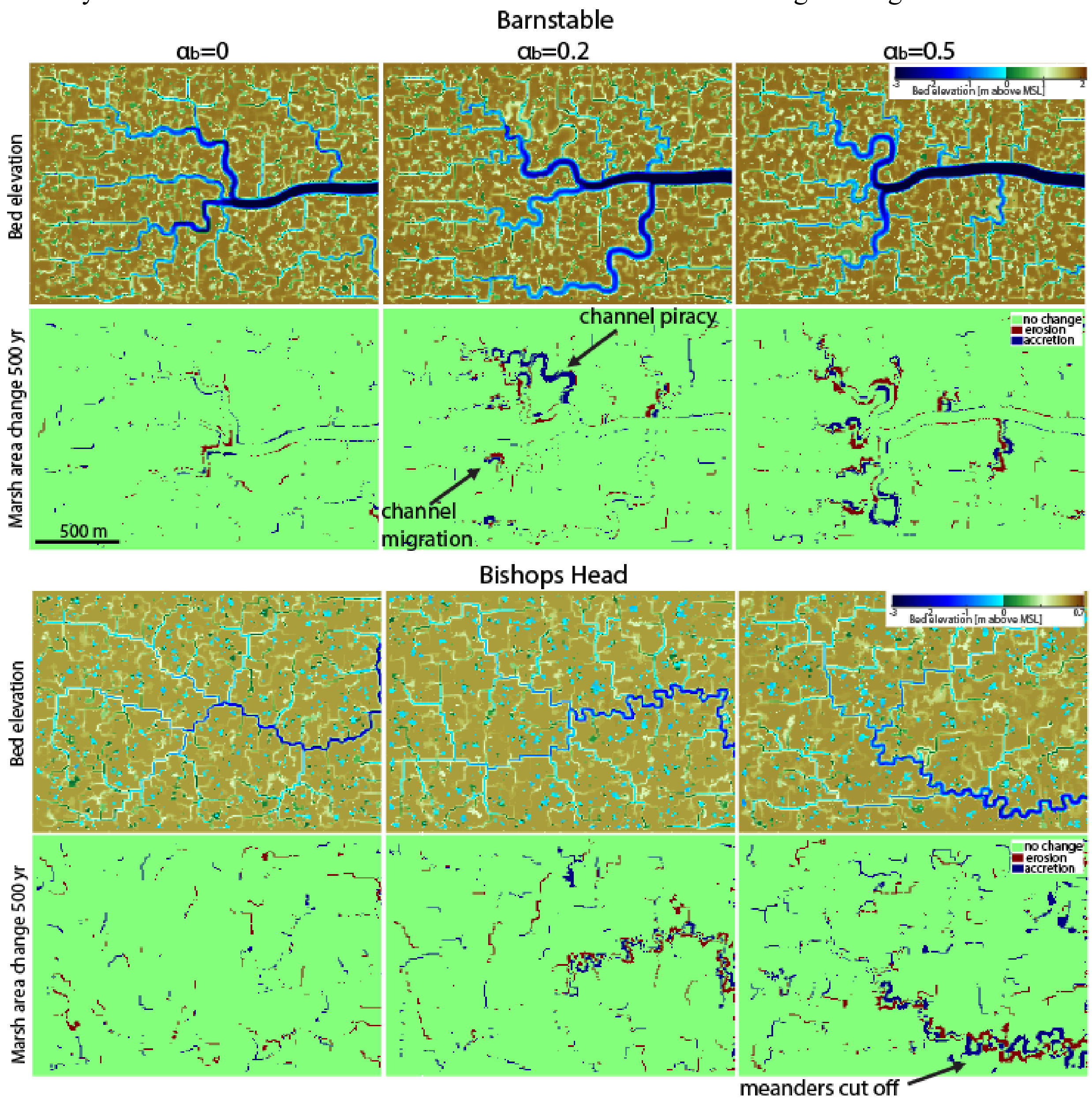


Figure 6: Comparison of simulated topobathymetry at t=3000 yr for different values of the bank erodibility coefficient k. The resolution of the computational grid dx=10 m is fixed for all the simulations. The Barnstable case with k=0.2 shows an instance of channel piracy, where the abandoned channel is quickly filling in (colored blue in the panel showing the marsh area change).

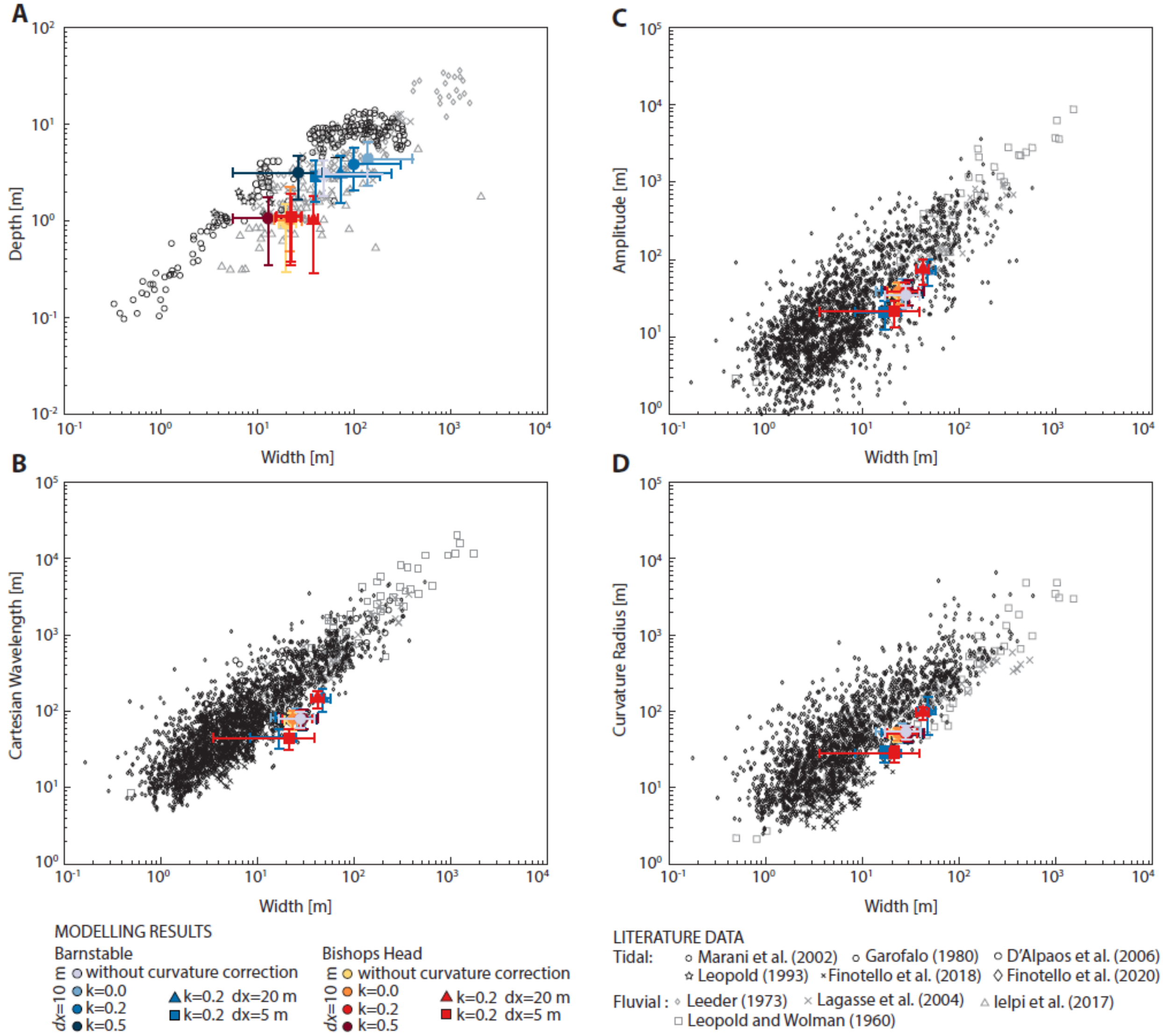


Figure 7: Comparison between planform features of modeled tidal meanders with field data retrieved from the literature. Modeling results are reported as average ensemble data for individual simulations performed using different model parameters at both the Barnstable and Bishops Head study sites, with error bars indicating standard deviation. In agreement with the literature data, panel A shows data referring to entire tidal meandering channels, whereas the other panels display morphometric data specific to individual meander bends.

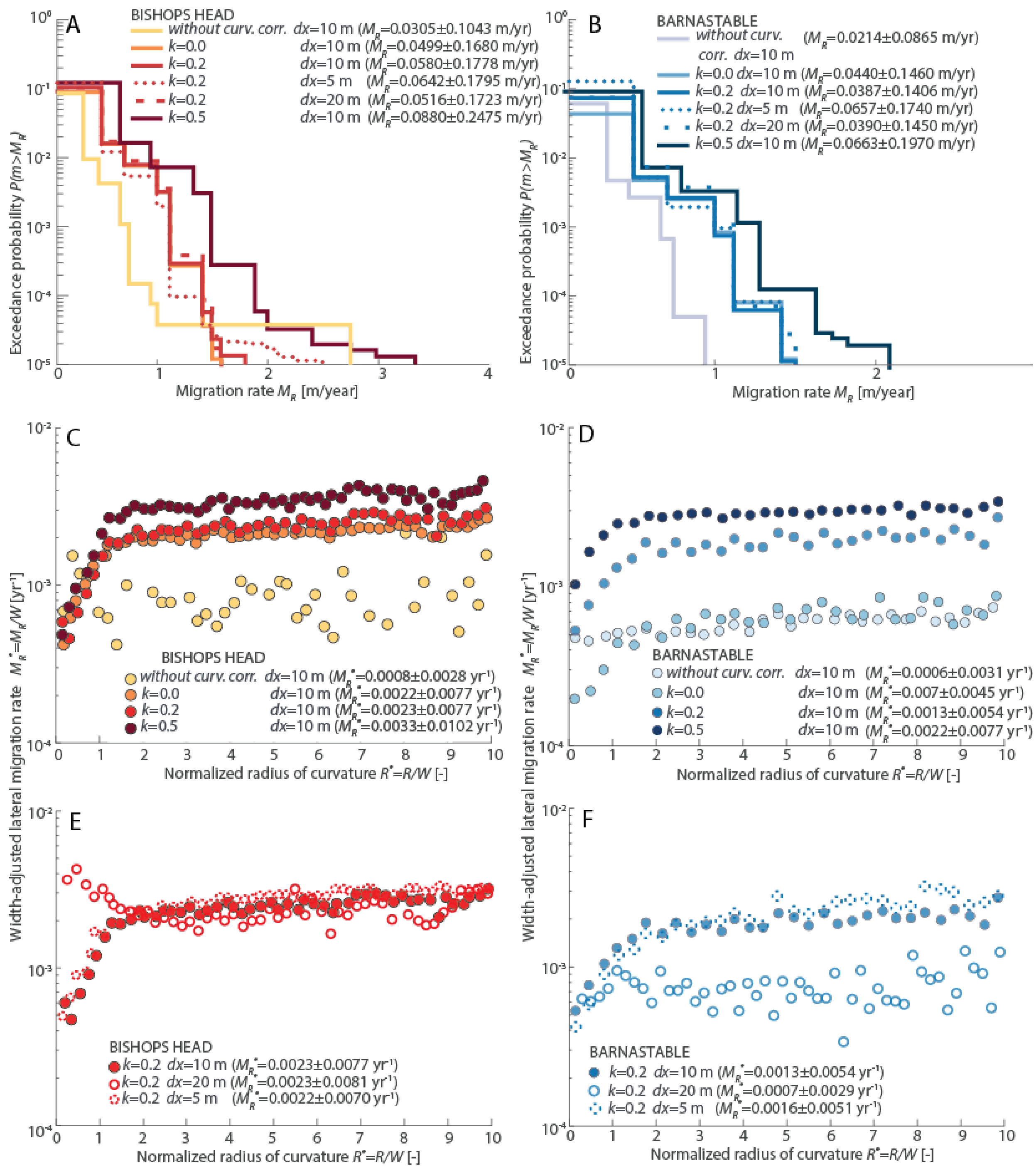


Figure 8: Characteristics of numerically modeled tidal meanders. A,B) Empirical probability distributions of numerically modelled meander migration rates M_R. C,D,E,F) Width-adjusted migration rates ($M_R^* = M_R/W$) plotted against the normalized meander radius of curvature ($R^* = R/W$), with dots representing the binned median values of the M_R^* distribution obtained by binning together a set of *n* data points equal to 100 and 50 for Barnstable and Bishops Head, respectively. In each panel, different marker/line color correspond to different simulations, with the parameters employed in each simulations reported in the legends.

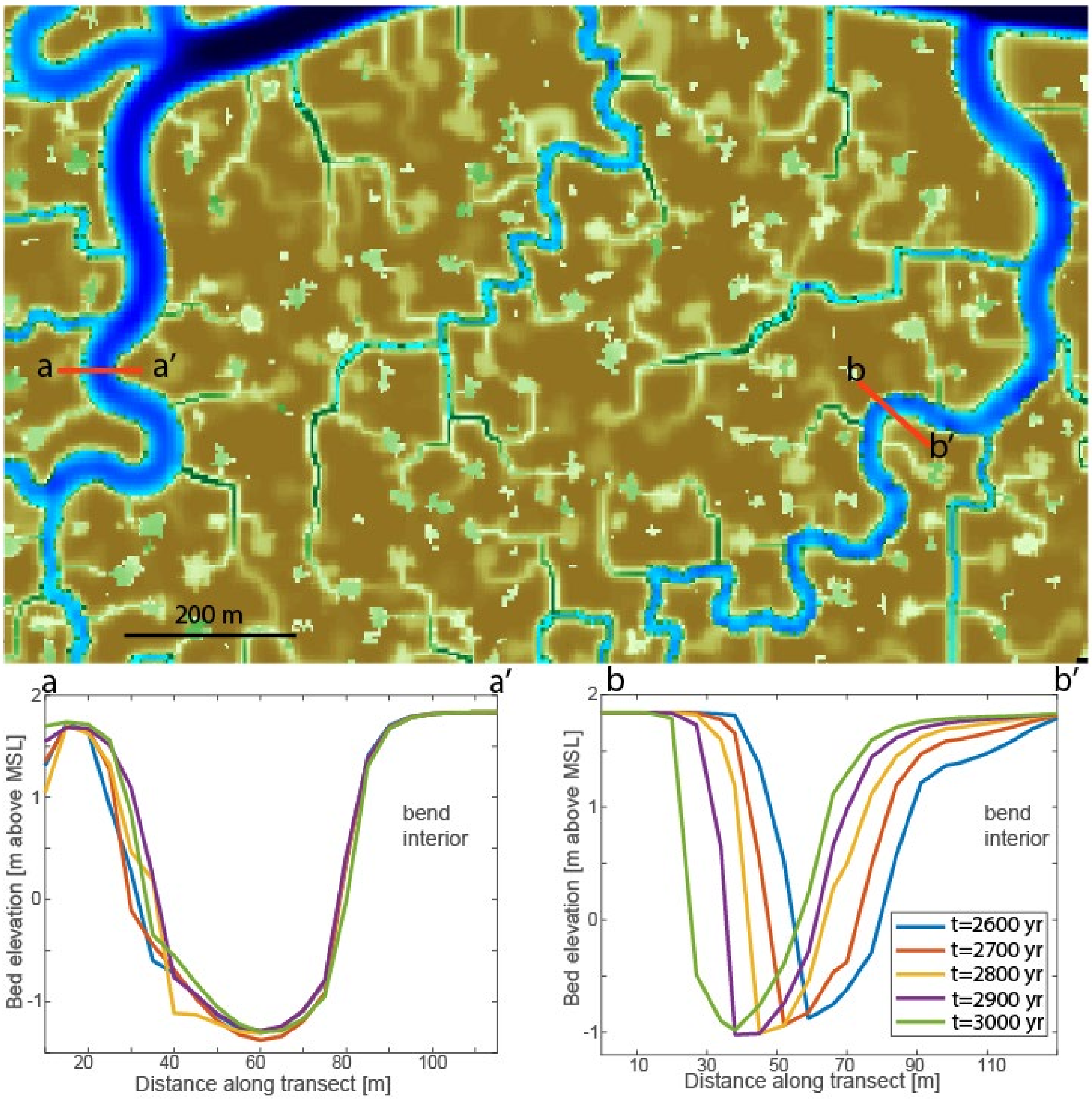


Figure 9: Details of channel migration at two bends, for the Barnstable case with dx=5 m, at t=3000 yr (as in Figure 11). Transect a-a' shows an example of a cuspate channel, where the thalweg is shifted toward the interior of the bend.

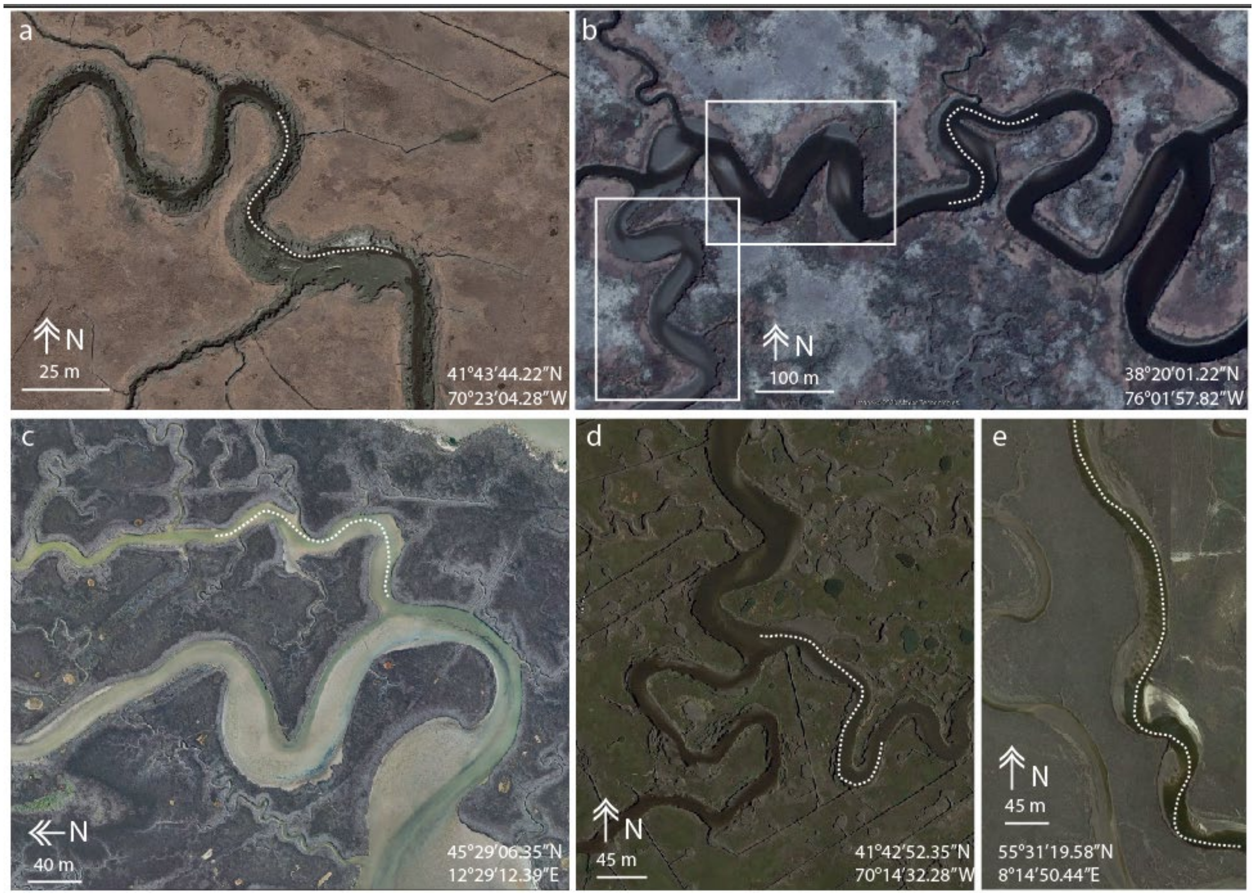


Figure 10: Examples of tidal meandering channels whose thalweg is not located at the outer bank. A) Barnstable Great Marshes, MA, USA (Image©Google, Landsat); B) Bishops Head, MD, USA (Image©Google, Maxar Technologies); C) Venice Lagoon, Italy (Image©Google, Landsat); D) Barnstable Great Marshes, MA, USA (Image©Google, Landsat); E) Skallingen Peninsula, Denmark (Image©Google, Landsat). The approximate thalweg position is denoted by dashed lines. Rectangles in panel B denote segregated ebb and flood channels.

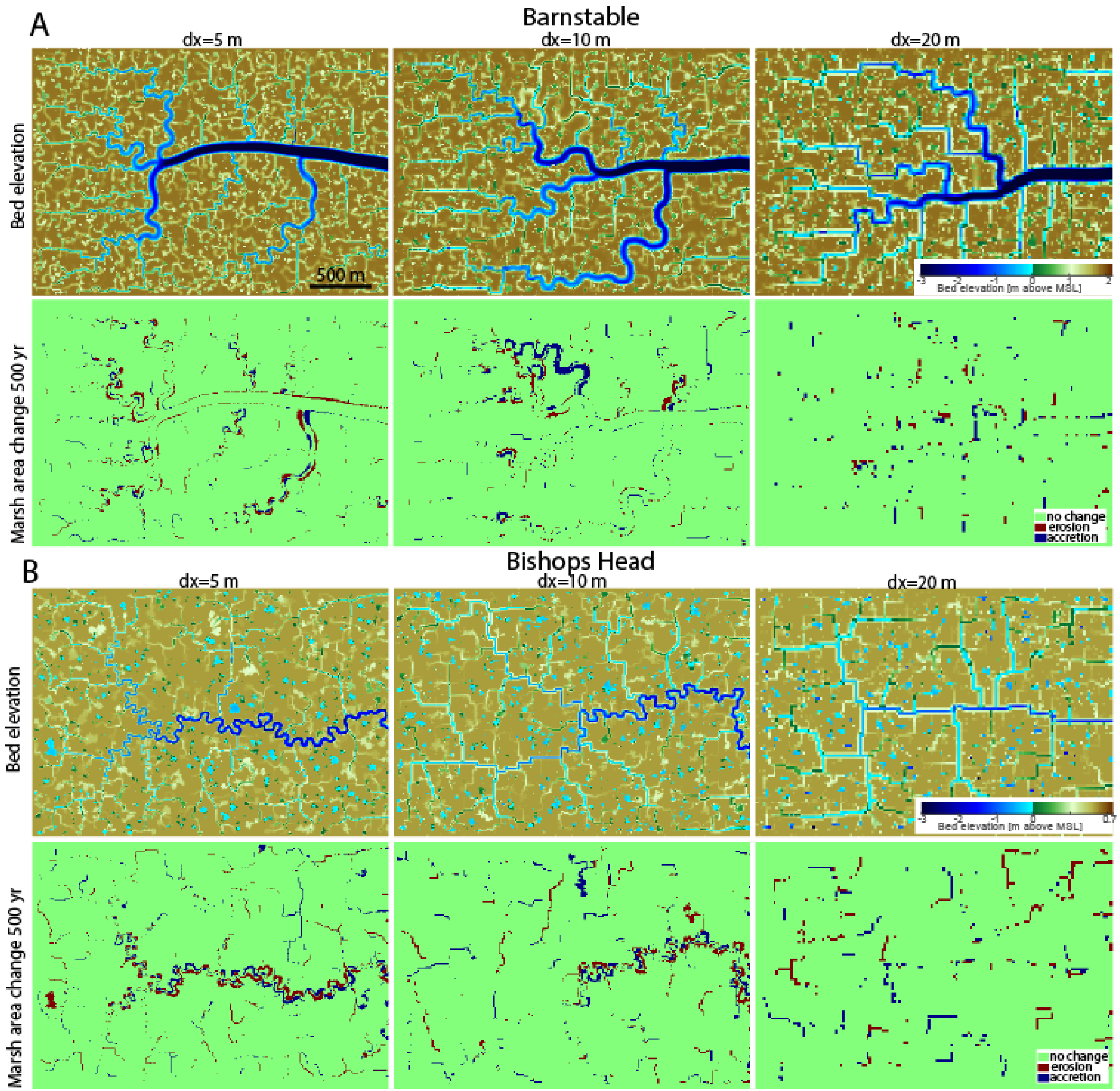


Figure 11: Comparison of simulated topobathymetry at t=3000 yr for different grid resolutions (*dx*) with the standard bank erodibility (*k* =0.2).

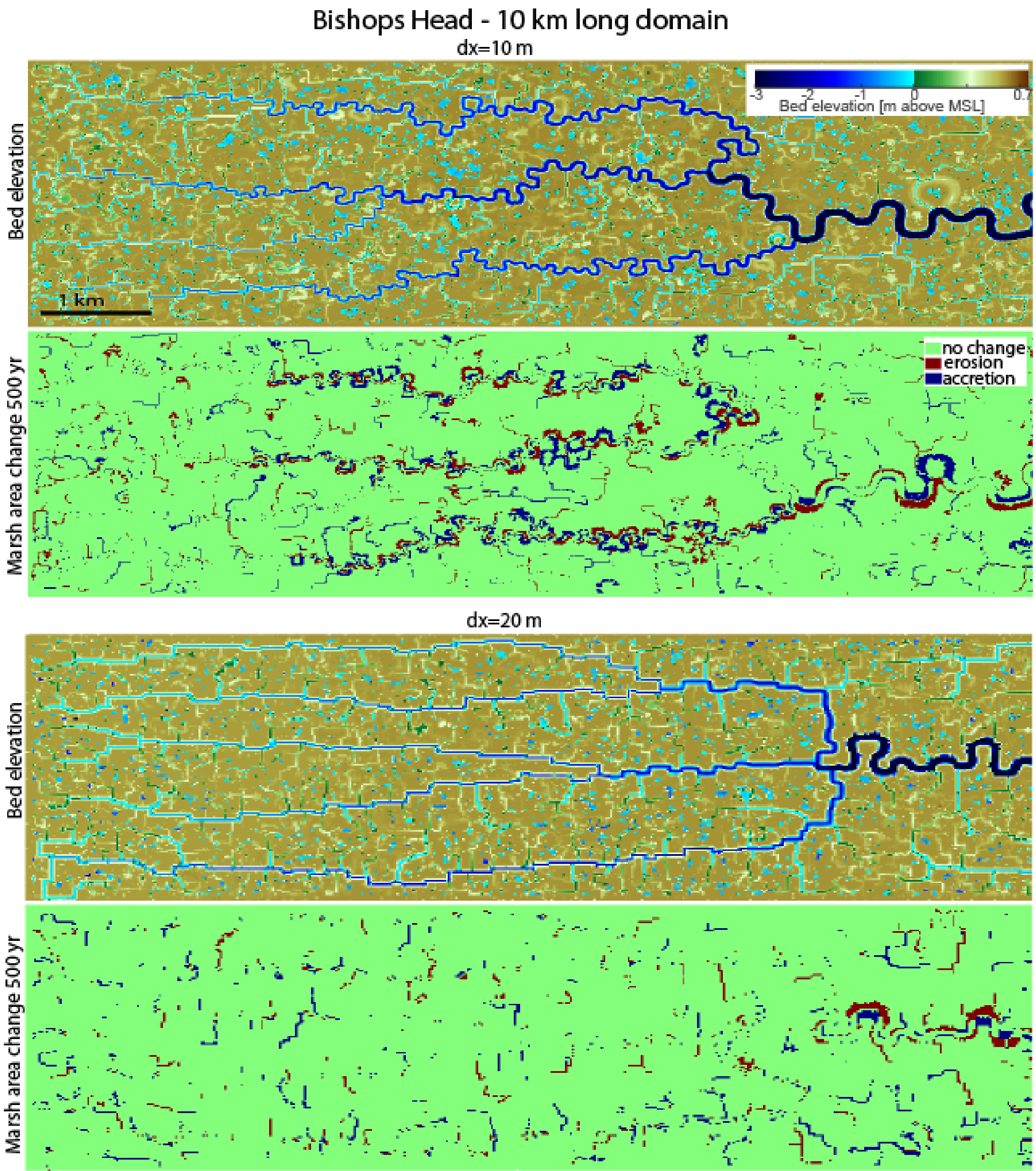


Figure 12: Comparison of simulated topobathymetry for Bishops Head at t=3000 yr for different grid resolutions (dx), with the standard bank erodibility (k=0.2). The domain is 10 km long instead of 3 km long (as in all other simulations).

Water Resources Research

Supporting Information for

A flow-curvature-based model for channel meandering in coastal marshes

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Text S1:

Analysis of meander development in a simplified, single-thread, straight tidal channel

We investigated the evolution of a simple, initially straight, single-thread channel with the goal of comparing our model to the one proposed by Solari et al. (2002). The latter authors developed a three-dimensional, linearized model to predict flow and bed topography in weakly meandering tidal channels forced by a perfectly symmetric, oscillatory tidal flow, providing the basis of a planimetric instability theory of the type developed for river meanders (Blondeaux & Seminara, 1985).

Despite being developed for similar purposes, our model and Solari et al.'s model exhibit several differences.

First and foremost, Solari et al.'s model is designed for single-thread, mildly sinuous, weakly meandering tidal channels with a fixed width and does not develop highly sinuous, high-amplitude bends or meander cutoffs. In addition to the strongly linearized nature of the model, the absence of high-curvature bends is attributed to the fact that oscillations associated with the basic (symmetric) tidal flow result in symmetric oscillations of the point bar–pool pattern around meander apexes, with no net migration during a tidal cycle (at least for periodic tides with zero mean, as used in Solari et al.). In contrast, our model also develops high-amplitude, highly sinuous meanders that migrate laterally (possibly forming cutoffs), features varying tidal channel width (as a function of varying tidal prism), and is suitable for use even in complex, dendritic networks of channels with widespread confluences.

Furthermore, Solari et al.'s model reproduces sediment transport (both bedload and suspended load) of non-cohesive sediment (i.e., fine sand) characterized by a settling velocity of 10 mm/s. In contrast, our model considers only the transport of cohesive sediment (mud) with a much smaller settling velocity of 0.2 mm/s, accounting for the creeping of bank soil, i.e., soil diffusion. As such, Solari et al.'s model primarily focuses on the development of sandbars in mildly sinuous tidal channels, despite their occasional presence not being universal. On the other hand, our model addresses the opposite end of the spectrum, where the sediment is exclusively composed of mud, and consequently, sandbars cannot exist.

Because of the highlighted structural differences between the two models, we do not necessarily anticipate a match in terms of numerical results. Nevertheless, to validate this hypothesis, we investigated how sinuosity (i.e., meandering) develops from an initially straight, single-thread channel in our model, allowing for a direct comparison with Solari et al.'s results. For this purpose, we considered a simplified case with i) initial bathymetry featuring a single-thread, straight, 3 km long tidal channel, with a rectangular cross section (1 m deep and 30 m wide) ; and ii) a constant tidal prism imposed as a constant value at the head of the channel, ensuring a constant flow rate in the along-channel direction (similar to the numerical simulation reported in Figure 2 in the main text). We note here that channel width and depth evolve naturally from the initially imposed values based on erosional and depositional processes. We neglected temporal variations in tidal range and sea-level anomalies and excluded wind waves, though we retained vegetation processes and sea-level rise to achieve realistic simulations over extended periods (i.e., millennia).

We carried out two different simulations using different computational cell sizes, equal to dx=5 m and dx=10 m, respectively. Both simulations utilize a fixed timestep dt=2 years and have a total duration of 2000 years (to allow the system to reach a dynamic equilibrium in terms of meander morphodynamics).

Commencing with the initially imposed single-thread straight morphology, sinuosity evolves naturally in our simulations driven by soil creeping from the channel banks, without the need to impose initial perturbations in the bed elevation (i.e., solely by the small errors associated with finite precision arithmetic). This is in contrast to Solari et al.'s model and many other meander centerline migration models, which necessitate small perturbations in the channel bed for meandering initiation. To directly compare the results of our model with those obtained by Solari et al., we conducted a quantitative analysis of the dimensionless meander wavenumber, $\lambda = 2\pi \cdot B/L_s$, where $B = W/2$ is the average meander half-width, and L_s is the intrinsic (i.e., along-channel) meander wavelength. The latter, like all other meander morphometrics, is computed using a reliable procedure for the automatic identification of meander inflection points as the locations of zeros of the curvatures after spectral approximations and appropriate windowing (Finotello et al., 2020; Marani et al., 2002). In our simulations, small perturbations in channel bed topography grow progressively larger, leading to the channel becoming increasingly sinuous (Figure S1 and S2). After approximately 400 years, high-amplitude meander bends form, with cutoff initiation occurring around 1000 years regardless of computational cell size, and then becoming progressively more frequent as more bends grow more sinuous (Figure 3). In accordance with this evolutionary trajectory, λ is initially large and decreases rapidly (with a larger dx leading to a faster decay; Figure 3). After approximately 1500 years, the average value of λ stabilizes, though still exhibiting fluctuations around the mean.

Once this stabilization occurs, the computed λ values are equal to 0.85±0.72 and 0.66±0.64 for dx=5 m and dx=10 m, respectively. These values broadly align with field data reported in the literature. As reported by Marani et al. (2002), the wavenumber in real study cases consistently ranges between 0.01 and 0.8, with average values in the order of 0.2-0.4 (i.e., L_s=10-15 B). Barnstable, in particular, features λ=0.6±0.2 (Fagherazzi et al., 1999; Rinaldo et al., 1999a, 1999b). In contrast, Solari et al. found wavenumbers on the order of λ=0.02-0.2, indicating that their model appears to overpredict meander wavelengths. Our model, in contrast, while somewhat underpredicting λ, provides results that better align with field data (although we note that considering a single tidal channel with constant width does not necessarily allow for a direct comparison with field data encompassing tidal meanders found in several distinct environmental contexts and with sizes spanning a wide range of widths and lengths). This suggests that a model like ours, albeit simplified, is capable of reproducing finite amplitude dynamics and might outperform models reproducing only small amplitude dynamics, such as those by Solari et al.

Finally, it is noteworthy that simulations with larger prisms (and hence larger channel width) produce longer meanders (Fig. S4), indicating that the wavelength scales with the channel width, which is also in agreement with field data.

Text S2:

Hydrodynamic comparison with Delt3D

To assess the ability of our simplified model (Eqs. 1-5) to reproduce hydrodynamic processes in meander bends, we also performed a comparison with Delf3D (Lesser et al., 2004), which can solve the shallow water equations in both 2D and 3D mode.

For the Delft3D simulations we considered a channel with identical geometry to that depicted in Figure 3 (depth of 3 m, with of 40 m, meander amplitude of 150 m, meander wavelength of 500 m). Additionally, we maintained a constant discharge of equal to 28 m^3/s, consistent with the simulation illustrated in Figure 3, thus reflecting a riverine scenario rather than a tidal one. Employing a curvilinear grid for enhanced precision, we initially conducted a depth-averaged simulation, where secondary circulation is absent. Subsequently, a 3D simulation was performed, comprising 30 vertical layers, to capture the development of secondary flow.

In the depth-averaged case, Delft3D predicts a higher velocity in the meander inner bank compared to the outer bank, akin to our simplified model without curvature correction (Fig. S5). Conversely, in the 3D simulation, the flow is shifted toward the outer bank, albeit asymmetrically, with a more pronounced shift toward the meander downstream side relative to the bend apex. This pattern would of course reverse if the flow were directed in the opposite direction.

As in our model simulation, the presence of a secondary flow in Delft3D increases the velocity on the outer bank on the order of 0.05-0.1 m/s. In both models (Delft3D in 3D mode and the simplified model with curvature correction), the maximum flow at the outer bank is in the order of 0.3 m/s (even though the location of the maximum velocity is slightly different). The presence of a secondary flow also reduces the Delft3D flow in the inner (convex) bank, but this reduction is less pronounced than that predicted by the simplified model.

The mismatch in velocities might indicate that the empirical flow correction in the simplified model is overestimated. However, it should be emphasized that our simplified hydrodynamic model was not meant to precisely replicate the flow in a curved channel. Rather, its purpose was to establish a reference flow that suffices in driving channel meandering. For example, the simplified flow field is designed to incorporate the impact of secondary circulations that lead to sediment deposition, and the concurrent formation of point bar, at the meander inner banks. Hence, an overestimation of the curvature correction might be considered as a way to compensate for the fact that the simplified model, being strictly 2D, does not reproduce cross-channel advection of sediment.

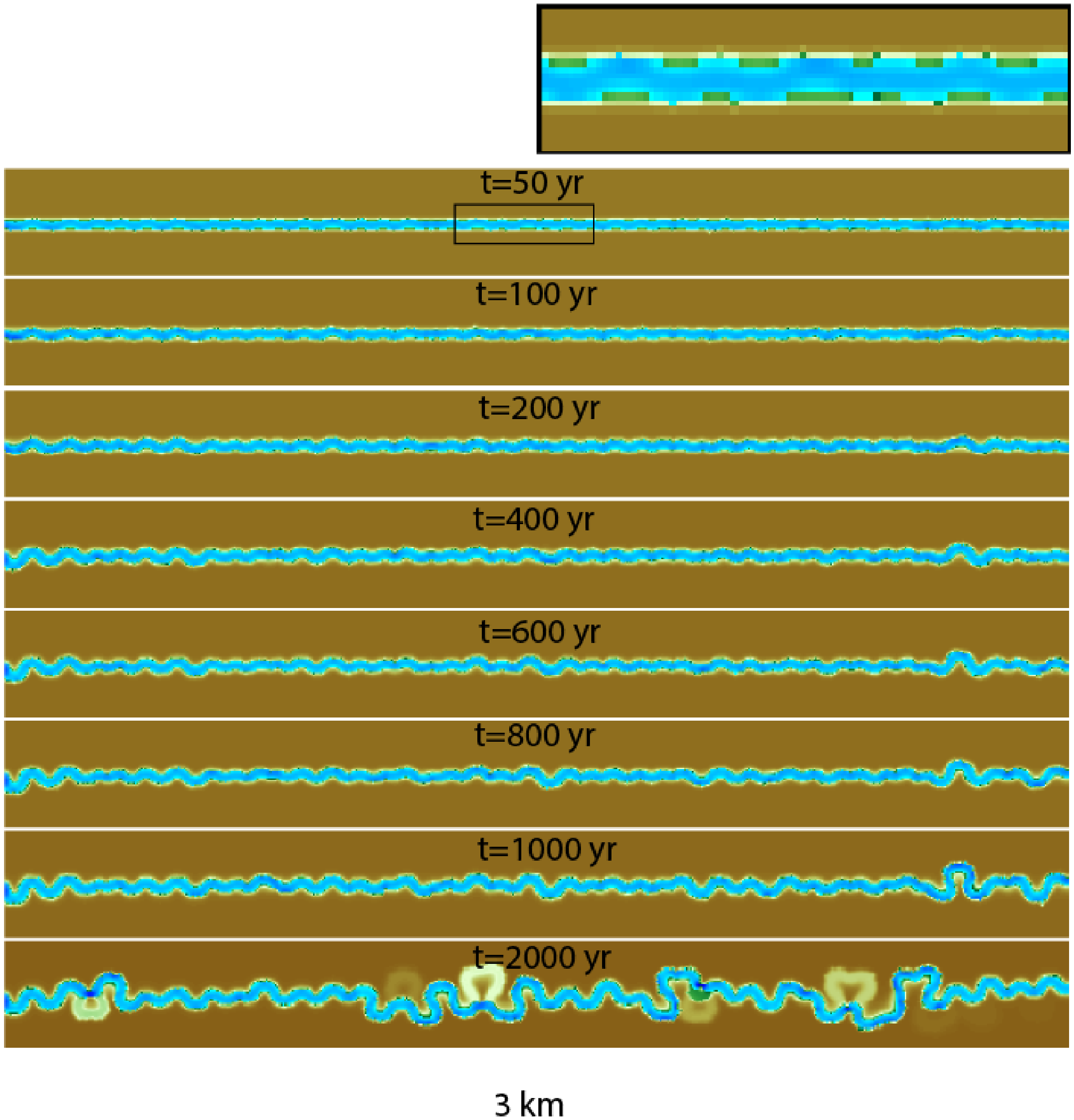

Figure S1: Development of meanders from an initially straight channel (dx=5 m). The tidal prism is imposed as a constant at the upstream (left) side.

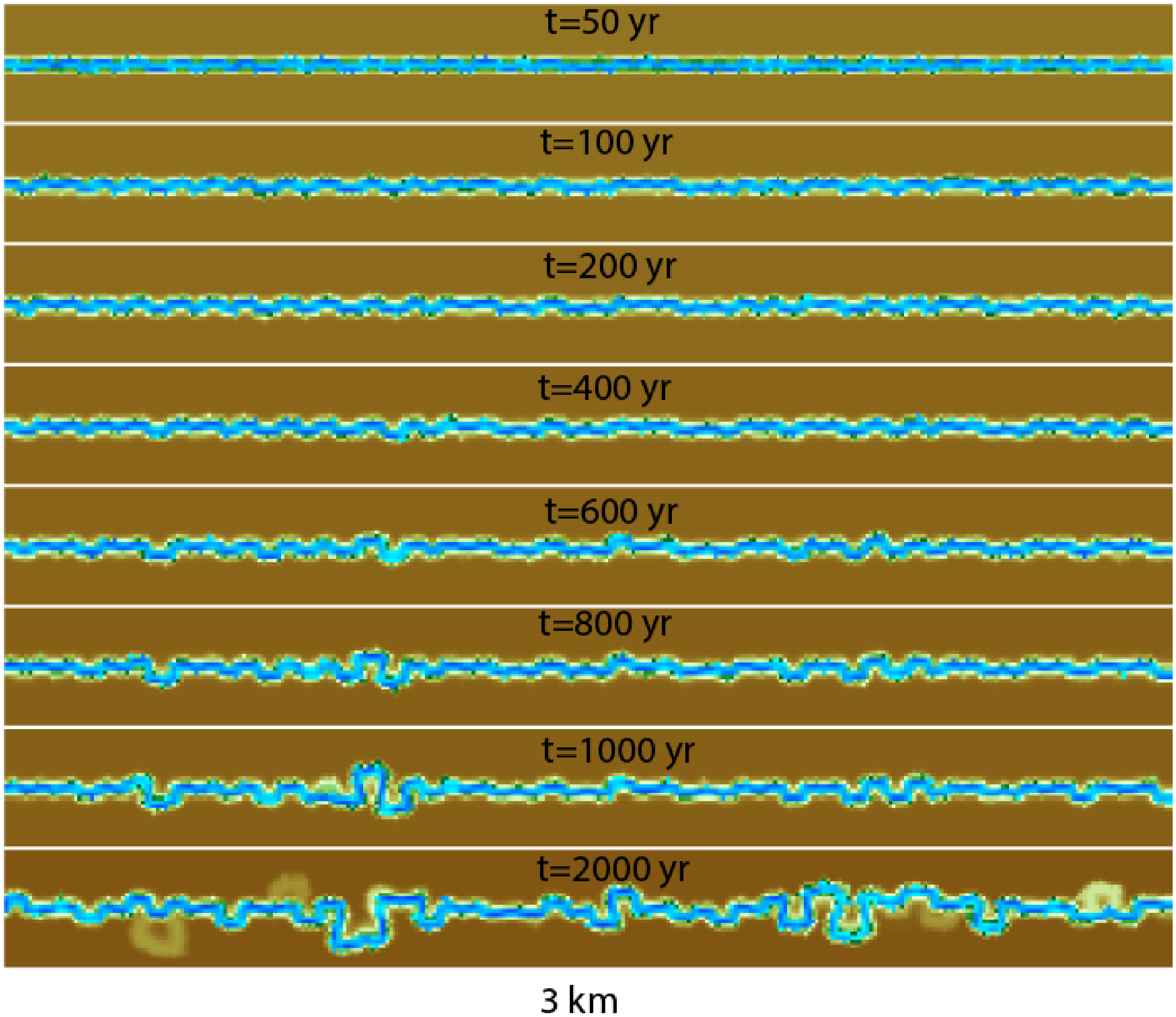

Figure S2: Development of meanders from an initially straight channel (dx=10 m). The tidal prism is imposed as a constant at the upstream (left) side.

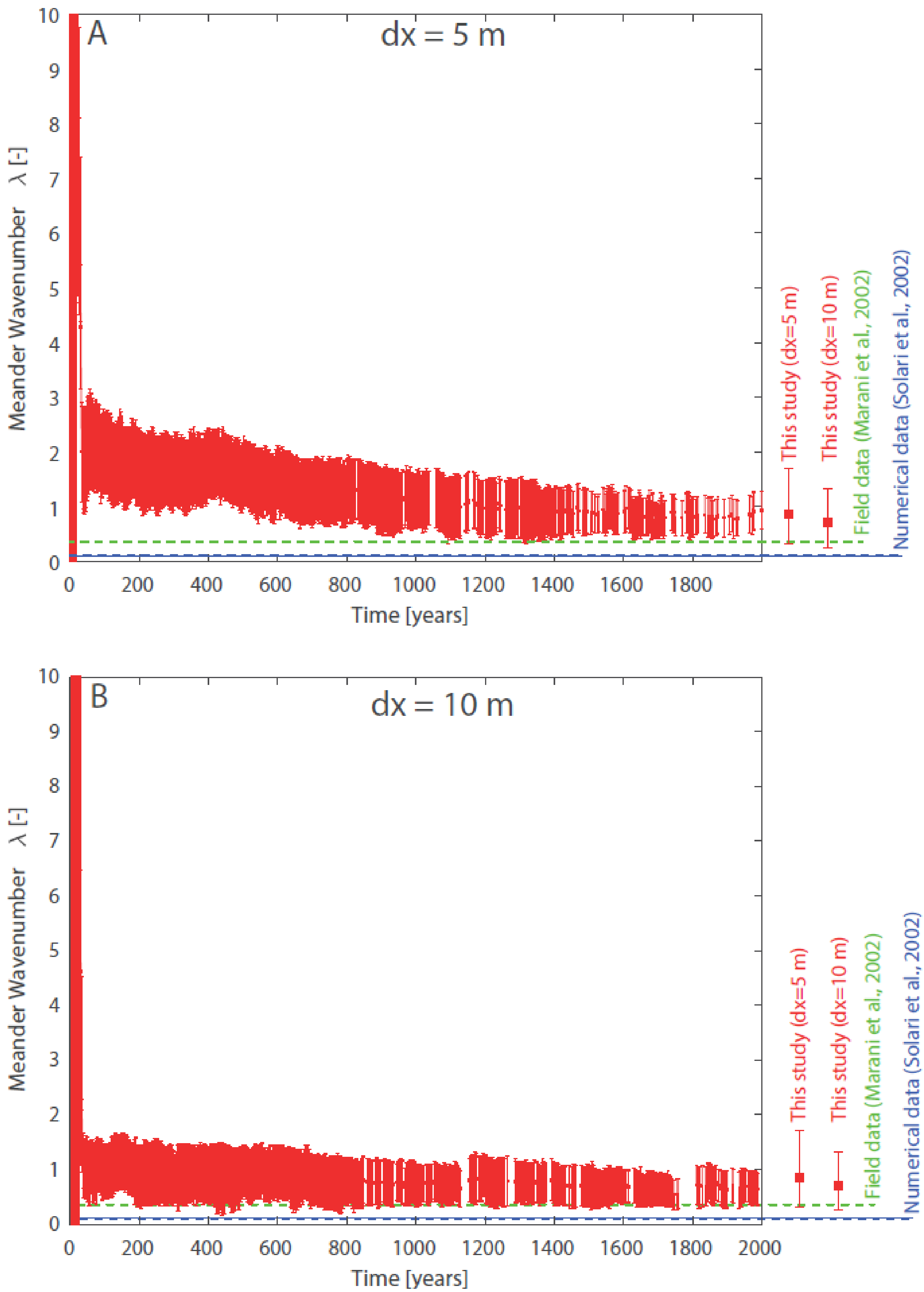

Figure S3: Evolution of the meander wavenumber value (λ, defined *sensu Solari et al., 2002)* as observed from the numerical simulations reported in Figure S1 (dx=5 m, panel A) and Figure S2 (dx=10 m, panel B). For each time step, the mean value of λ is reported together with its standard deviation, shown as error bars. Missing data after 800-1000 years denote timestep for which meander cutoff occurred, and which were not considered in the analysis. Insets in the right-hand side of the figure report field and numerical results derived from Marani et al. (2002) and Solari et al. (2002), respectively.

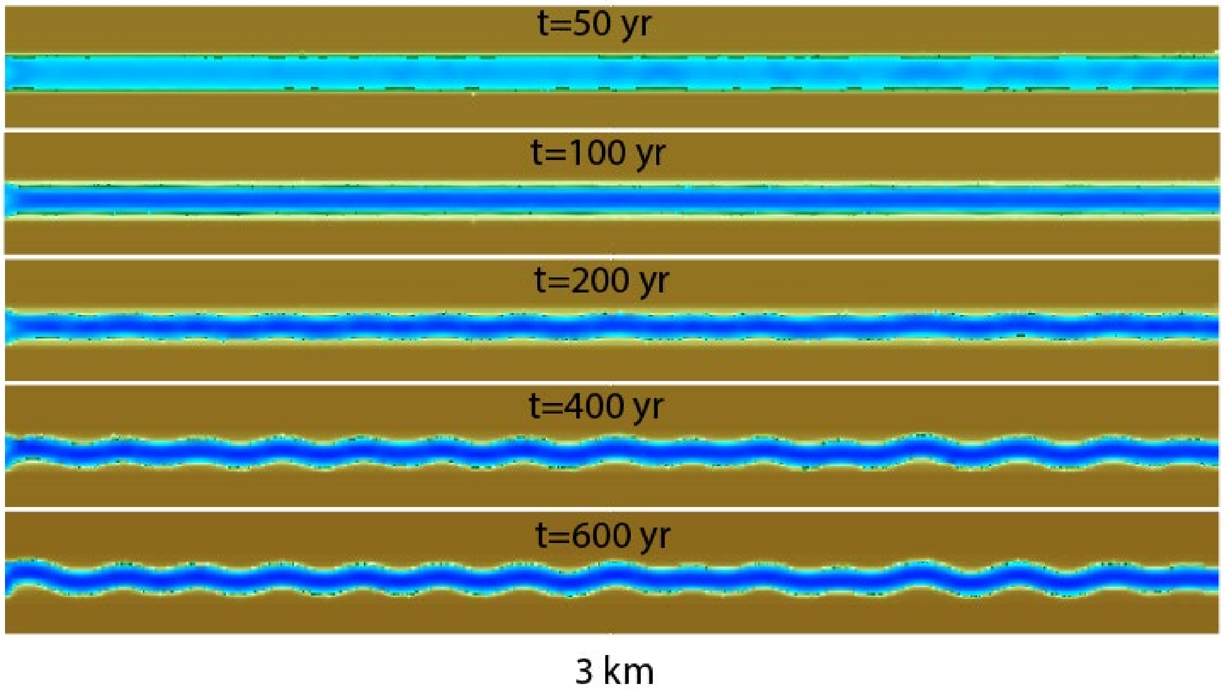

Figure S4: Results of the same numerical simulation reported in Figure S2 (*dx*=10 m), but considering the case with twice tidal prism. Note that the wavelength is now double that for the case with the smaller tidal prism, confirming that the wavelength scales with the channel width.

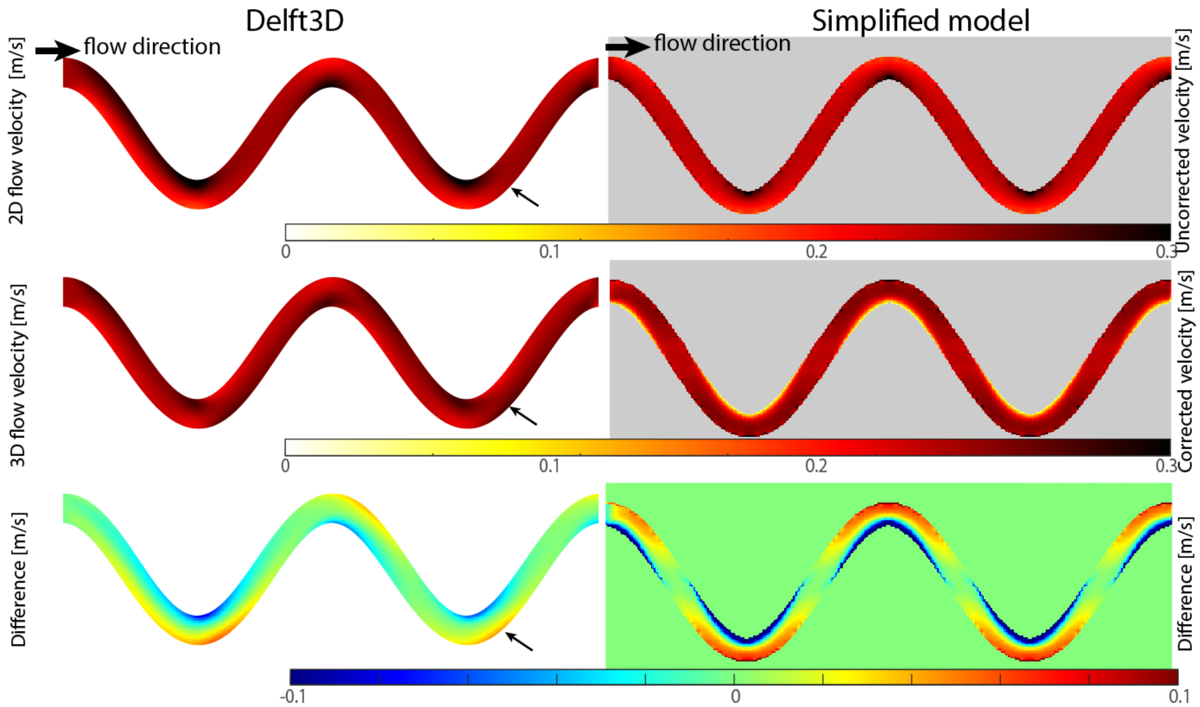

Figure S5. Hydrodynamic comparison between the simplified model (as in Figure 3) and Delft3D. In Delft3D we considered the 2D case (which does not have secondary circulation) and the 3D case, which fully resolves the secondary circulation and hence the shift in velocity toward the outer bank. The depth average velocity magnitude is reported for Delft3D.

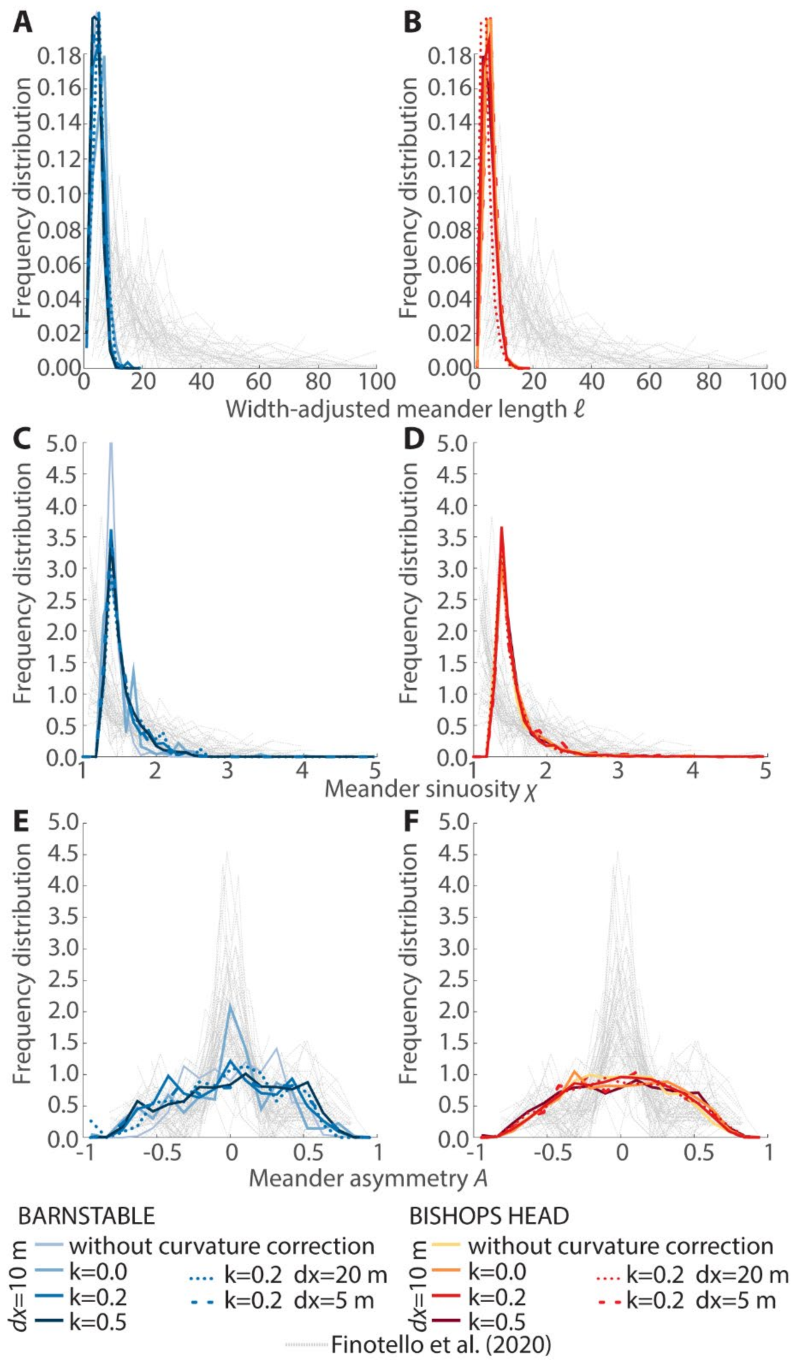

Figure S6. Frequency distributions of dimensionless meander morphometric variables. (a,b) width-adjusted meander intrinsic length; (c,d) meander sinuosity; (e,f) meander asymmetry index. Modeling results are presented as probability distributions obtained by considering the morphologies of the modeled meandering channels at the final time step of the numerical simulations at both the Barnstable (blueish colors) and Bishops Head (reddish colors) study sites. Different colors and line styles denote different numerical simulations. Gray lines in the background represent literature data on tidal meandering channels derived from Finotello et. al (2020).

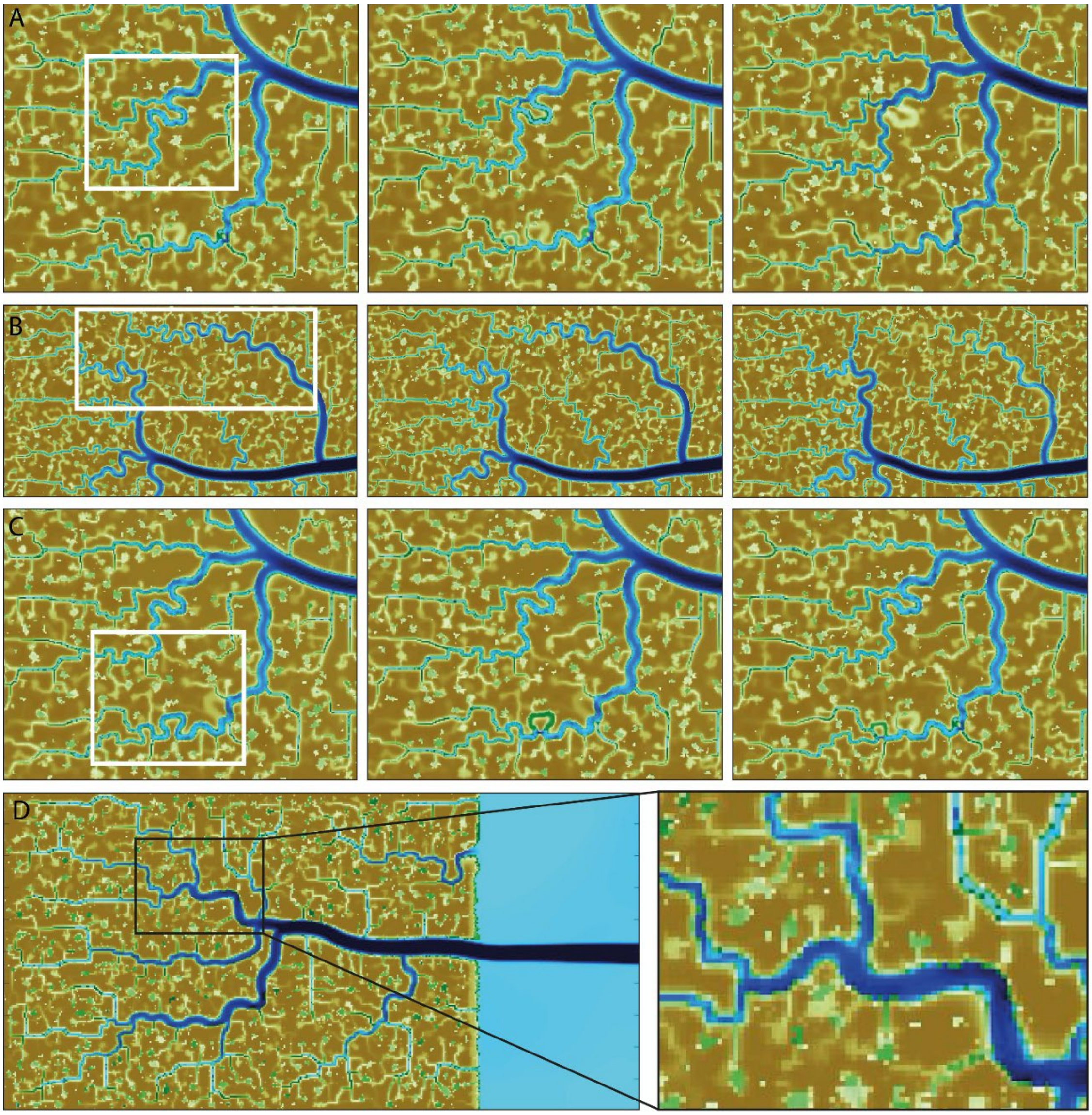

Figure S7. Examples of inter- and intra-channel dynamics and peculiar meander morphologies induced by stream meandering in the numerical simulations representing the Barnstable study case. A) Channel piracy forming a morphology akin to a meander cutoff. B) Repeated meander cutoffs and piracy resulting in a network reorganization. C) Meander Cutoffs. D) Meander bends with cuspate inner banks. Rectangles highlight the area of interest.

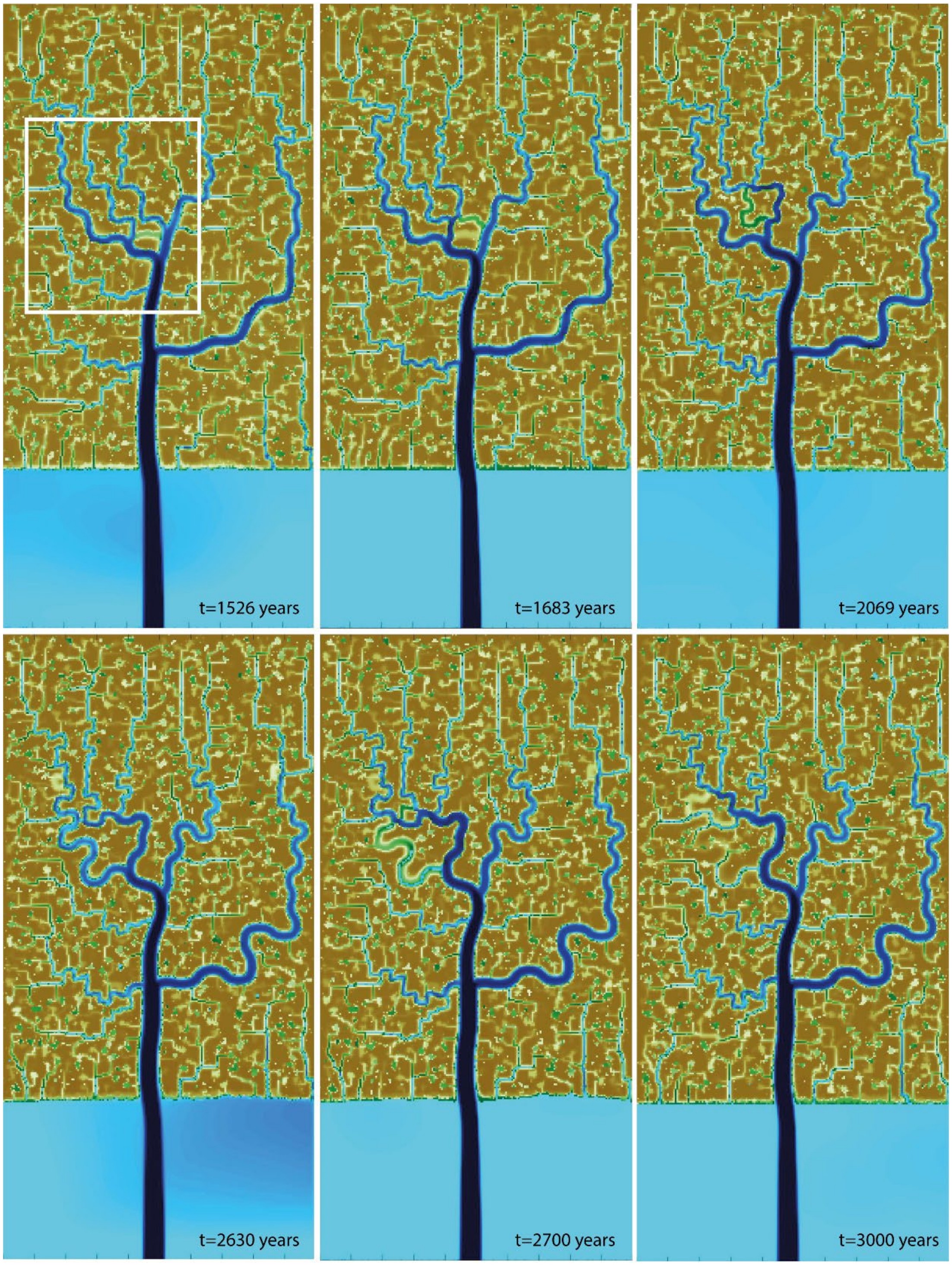

Figure S8. Examples of inter- and intra-channel dynamics (i.e., meander cutoffs and channel piracies) resulting in repeated network reorganizations in the numerical simulations representing the Barnstable study case.

Figure S9. Example of channel cutoff and piracies from the Seabrook-Hamptons Estuary (NH, USA) (Image ©Google, Landsat/Copernicus).

List of videos

Video S1. BARNSTABLE_k0_dx10. Simulation for Barnstable marsh, with dx=10 and k=0.
Video S2. BARNSTABLE_k02_dx10. Simulation for Barnstable marsh, with dx=10 and k=0.2.
Video S3. BARNSTABLE_k05_dx10. Simulation for Barnstable marsh, with dx=10 and k=0.5.
Video S4. BARNSTABLE_k02_dx5. Simulation for Barnstable marsh, with dx=5 and k=0.2.
Video S5. BARNSTABLE_k02_dx20. Simulation for Barnstable marsh, with dx=20 and k=0.2.

Video S6. BISHOPS _k0_dx10. Simulation for Bishops Head marsh, with dx=10 and k=0.
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Video S11. longBISHOPS_k0_dx10. Simulation for Bishops Head marsh with a long domain (as in Figure 12), with dx=10 and k=0.
Video S12. longBISHOPS_k02_dx20. Simulation for Bishops Head marsh with a long domain (as in Figure 12), with dx=20 and k=0.2

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