A method to estimate the longshore sediment transport at ebb-tidal deltas based on their volumetric growth: Application to the Guadiana (Spain–Portugal border)

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ABSTRACT: Current techniques assessing longshore sediment transport rates have large uncertainties, pleading for the development of alternative and complementary approaches. The present study proposes a method to estimate the decadal average rate of longshore transport at modern ebb-tidal deltas based on a sediment budget analysis of the outer shoal growth. This transport is obtained as the balance of the other contributions to the shoal with the total sediment input rate obtained from an inverse application of the inlet reservoir model. The method is applied to the Guadiana ebb-tidal delta, yielding an average longshore sediment transport rate (~85 000 m3 year−1) in good agreement with expectations for the region. It is exemplified that this decadal averaged rate can be used to improve longshore sediment transport expressions in order to study its variability over shorter time scales. At the Guadiana, the yearly longshore sediment transport from the improved formula ranges from ~25 000 m3 (westward) to ~245 000 m3 (eastward) and is related to the North Atlantic Oscillation index. Overall, the proposed method constitutes an alternative tool to constrain the average longshore sediment transport rate over decades in the vicinity of tidal inlets. It is applicable to ebb-tidal deltas where the outer shoal growth (from an early to a mature stage) is well documented by bathymetric maps, and where the main transport pathways towards the outer shoal can be specified. © 2019 John Wiley & Sons, Ltd.

KEYWORDS: Longshore transport; Ebb-tidal delta; Inlet Reservoir Model; CERC equation; NAO index; substance use disorders

Introduction

Longshore sediment transport (LST) is the main driver of the shoreline planform evolution along many coastal environments and must be reliably quantified for sustainable management and planning (CERC, 1984; Komar, 1998; Dean and Dalrymple, 2004). The most common LST estimations are obtained from the application of empirical bulk equations like the ones proposed by Komar and Inman (1970), Bijker (1971), CERC (1984), Kamphuis (1991), or Bayram et al. (2007) and derivations/improvements of those (e.g. del Valle et al., 1993; Mil-Homens et al., 2013; López-Ruiz et al., 2014). These equations are based on waves and physical properties in the surf zone, since longshore transport results mainly from currents produced by oblique incident breaking waves (Longuet-Higgins, 1970) and are calibrated against – or rather adjusted to (see Cooper and Pilkey, 2004) – data measured in the field or in laboratory experiments. The resulting expressions are explicit and easily implemented with limited information from the site but are known to have large uncertainties (Schoonees and Theron, 1993; Wang et al., 1998; Bayram et al., 2001; Esteves et al., 2009; Güner et al., 2011). In particular, LST predictions can vary widely depending on the bulk formula used (Fowler et al., 1995; Wang et al., 1998; Van Welten et al., 2000; Schoonees and Theron, 2001; Tonk and Masselink, 2005; Güner et al., 2011; Mil-Homens et al., 2013). Direct quantifications at the site are required to calibrate (or improve, at least) bulk equations or validate their results, but are rarely available (Cooper and Pilkey, 2004; Smith et al., 2009).

Tracing techniques (e.g. Inman et al., 1980; Ciavola et al., 1997; Tonk and Masselink, 2005), sediment traps (e.g. Dean et al., 1982; Kraus and Dean, 1987; Wang et al., 1998; Tonk and Masselink, 2005) and OBS measurements (e.g. Tonk and Masselink, 2005) are commonly used to quantify the LST over the short term. Quantitative information at longer time scales (e.g. yearly, decadal) can be obtained from the analysis of shoreline or topo-bathymetric changes like at barrier spits (Lee and Birkemeier, 1993; Aagaard et al., 2004). Similarly, direct observations of the volume of the littoral drift accumulated against engineering structures (jetties, groins and breakwaters) can provide solid estimates of the longshore transport considering the entire beach system (rather than the upper beach only) if the crossshore transport is negligible or well-quantified (e.g. Bruno et al., 1981; Dean et al., 1982; Bodge and Dean, 1987; Wang...
Some authors have also assessed LST rates based on the infilling of frequently dredged pits at harbours (e.g. Patsch and Griggs, 2008). Besides, sediment budget analyses – establishing the balance of volumes for sources and sinks along a portion of the coast during a given time (Bowen and Inman, 1966; Rosati, 2005) – are also frequently implemented to constrain the LST within regional sediment cells (e.g. Patsch and Griggs, 2008; Kaminsky et al., 2010; Santos et al., 2014). However, all these techniques have uncertainties linked to the measurement methods, to the complexity of the transport process and to the large temporal (from hours to centuries) and spatial (from tens of metres to hundreds of kilometres) scales at which the longshore transport impacts nearshore environments (Larson and Kraus, 1995; Hanson et al., 2003; Cowell et al., 2003a, 2003b). Assessing the longshore transport is especially difficult over long time periods when the scale range is wide. There is, therefore, a need to develop alternative or complementary assessment methods to constrain the medium to long-term (years to decades) LST based on field evidence (Stive et al., 2002).

In the present study, an innovative approach is described to constrain the medium-term (decadal) LST rate at the updrift beach of ebb-tidal deltas based on a sediment budget analysis of the outer shoal development. The method is applicable to deltas where the outer shoal growth from an early to a mature stage is well documented by bathymetric maps and requires specification of the main sediment transport pathways across the inlet. Through the reverse application of an aggregate model, the decadal average LST rate is constrained as best estimate along with upper prediction bounds at confidence intervals (useful to define coastal management strategies). The results obtained from a case study are compared with predictions from various bulk equations and further used, as an example, to adjust the CERC (1984) equation in order to assess the yearly LST rate variability.

**Regional Setting**

The study area encompasses the ebb-tidal delta (ebb delta, hereafter) of the Guadiana estuary, at the southern border between Spain and Portugal (Figure 1). Referring to the terminology of Hayes (1980), the Guadiana is a mixed-energy, tide-dominated inlet (Morales, 1997). The tidal regime in the area is semi-diurnal with a mean range of 2 m. Waves of moderate energy dominate the offshore wave climate, with yearly average significant height and peak period of 1 m and 8.2 s, respectively (Costa et al., 2001). The wave field (see Figure 1) comprises both swell and sea waves from the W–SW directions (71% occurrence) and mostly sea waves from the SE (23% occurrence).

Before jetty installation in 1972–1974, the historical ebb delta was broad, asymmetric eastwards and characterized by the presence of a large sandy shoal (the O’Bril bank) in front of the estuary mouth (Gonzalez et al., 2001). In 1969, for instance, the bank was ~1 km wide and extended ~4 km eastward from the western margin (see dashed black contour in Figure 1). In response to tidal flow stabilization and constriction by jetties, the eastern part of the historical delta collapsed, while a modern ebb delta formed off the mouth (Garel et al., 2014). The modern delta developed into an inlet channel scoured into the O’Bril bank, bounded by sand storage areas. The latter areas include a relatively straight updrift lateral bar and an outer shoal – or ebb shoal ‘proper’ following Kraus’s (2000) terminology – with a typical horseshoe shape in plan view marked by a series of sub-parallel lobate swash bars. In the east, the outer shoal connects to a broad downdrift complex, which corresponds to the swash...
platform of the historical delta (Figure 1) and is affected by widespread erosion (Garel et al., 2014). Dredging of the outer shoal was performed in 1987 and 2015 to improve the navigability of the entrance channel (see Garel, 2017).

The grain size in the area shows a large variability (between fine and coarse sand), typical of deltaic environments (see the Iberian Atlantic Margin Sediments Database described in Costas et al., 2018). The updrift beach is constituted of fine sand with mean grain size ($D_{50}$) about 0.3 mm on average. Sand with a similar $D_{50}$ is only observed on the downdrift swash platform, where part of the bypassed material is expected to deposit. $D_{50}$ at the lateral updrift bar is much coarser (2 mm), indicating a very energetic area exposed to SW storms constituted mainly of lagged material.

The LST direction in the region is always reported to be from west to east (CEEPYC, 1979; Granja et al., 1984; Andrade, 1990; Cuenca, 1991; Bettencourt, 1994; Gonzalez et al., 2001; Vicente and Pereira, 2001; Santos et al., 2014). The proposed regional rate estimates vary widely, from a minimum of 6000 m$^3$ year$^{-1}$ obtained by Andrade (1990) to a maximum of 300 000 m$^3$ year$^{-1}$ (Consulmar, 1989 cited in Bettencourt, 1994; Vicente and Pereira, 2001). Results for the east flank of Ria Formosa (immediately updrift of the study area; see Figure 1) range between 100 000 m$^3$ year$^{-1}$ (Andrade, 1990) and 150 000 m$^3$ year$^{-1}$ (Bettencourt, 1994). At the study site, a local rate of 180 000 m$^3$ year$^{-1}$ was established based on sand accumulation against the western jetty (Gonzalez et al., 2001). This value is considered a high estimate as a large part of this accumulation results from cross-shore transport due to the welding of a sandbank to the beach following jetty installation (Garel et al., 2015). Based on rough qualitative assessments about sediment sources and losses in the sediment cell, Santos et al. (2014) consider that the LST rate must be less than that of the adjacent sediment cell in the west (110 000 m$^3$ year$^{-1}$).

### Methods

#### Description of the approach

The proposed approach uses the inlet reservoir model (IRM), an analytical aggregate model based on mass conservation that describes the long-term volumetric evolution of ebb deltas and evaluates the associated bypassing rate (Kraus, 2000, 2002). The model requires the partitioning of the delta in distinct morphological elements that correspond to different deposition areas (so-called ‘reservoirs’) embracing the main transport pathways across the inlet. Such elements typically include the external bars of the ebb delta (FitzGerald et al., 2000; Carr and Kraus, 2001), and possibly other features depending on the complexity of the system (e.g. Dabee and Kraus, 2004, 2005, 2008; Cox and Howe, 2012). The volumetric growth of the outer shoal only is considered for the present model application. This is typically the first element of the modern delta to develop after barrier breaching or jetty installation.

The IRM assumes that the outer shoal (like the other reservoirs) has an equilibrium volume ($V_o$) corresponding to a maximum sand-retention capacity that is limited by wave action and cannot be exceeded. Its volume $V(t)$ increases through time as sediment is brought to it. Meanwhile, the shoal is continuously ‘leaking’ (i.e. bypassing) an increasing fraction of sand downdrift. Borderline cases are at the initial development of the shoal (most sediment arriving at the shoal deposits on it) and at the final development stage when its volume is close to equilibrium (most sediment arriving is bypassed). The model further assumes that the sand input rate to the outer shoal ($Q_{in}$) is proportionally linear to the sand output rate downdrift ($Q_{out}$).

In this case, the volumetric evolution of the outer shoal is described as:

$$V(t) = V_0 \times \left(1 - e^{-\frac{t}{T_{ed}}}\right)$$  \hspace{1cm} (1)

Note that through application of the same procedure to the other reservoirs, the model implicitly accounts for the temporal storage of material, for example within migrating sand waves and swash bars. For further details about the model, the reader is referred to Kraus (2002).

The sediment input rate $Q_{in}$ in Equation (1) represents all the sediment that feeds the outer shoal from various sources. It generally corresponds to the longshore transport plus other significant contributions related to back-barrier or river export and to the erosion of local features such as the historical delta. For typical IRM implementation, $Q_{in}$ is estimated from numerical models or coastal processes information from other studies (e.g. Kraus, 2000; Dabee and Kraus, 2004, 2005, 2008; Cox and Howe, 2012). By contrast, in the proposed approach $Q_{in}$ is not known beforehand but derived (along with $V_o$) from the best fit between predictions with Equation (1) and observations of the outer shoal volume. This is an inverse application of the model that relies on the availability of a relatively large bathymetric dataset describing the modern delta development.

To estimate the longshore transport, a sediment budget analysis is conducted at the outer shoal. The analysis requires the establishment of the main sources of sand that feed the outer shoal and of the associated sediment transport pathways. The input rates (other than longshore transport) are estimated from previous or complementary studies (see the Guadiana case study example) and averaged over the study period. The balance of these contributions with $Q_{in}$ (obtained from the reverse IRM application) allows constraining the longshore sediment transport rate, denoted $LST_{EA}$ hereafter. This rate represents the decadal averaged yearly volume of sediment that reached the delta from the updrift beach through longshore transport. It is important to note that by using the largest possible number of bathymetries and by considering averaged rates over the study period, the approach aims to smooth out any shorter-term LST variability.

### Implementation at the Guadiana ebb delta

The volumetric evolution of the Guadiana ebb delta was evaluated based on a series of 13 bathymetric maps ranging from 1969 to 2017 (Table I). The time interval between the maps is irregular, being at maximum 10 years (1995–2005) and at minimum 1 year (from 2014 to 2017). The original material (raw data, grids or maps) was converted to the ED50 UTM29N projection system and standardized to a 25 m grid size. Water depths in this study are referred to hydrographic zero, which is 2 m below mean sea level. The bathymetry of 1969 represents the surface before the completion of jetties and is used as a reference to measure the volume of the morphological elements that constitute the modern ebb delta (Dean and Walton, 1975). It is noted that this is 3 years before the beginning of jetty construction (1972, which is also the initial time of IRM simulations). Yet, between 1969 and 1972, the area was mainly constituted of the O’Bril bank (see Figure 1, black dashed contour), which was a rather stable feature at a yearly time scale (see Gonzalez et al., 2001). Therefore, it is reasonable to consider that the bathymetry did not change significantly during these 3 years and that the reference grid of 1969 represents adequately the pre-jetty surface.
Three morphological elements were distinguished for the establishment of their volumetric evolution: the updrift lateral bar, the inlet channel and the outer shoal (Figure 2). Typical at jetty-tied inlets (Bruun and Gerritsen, 1959; FitzGerald et al., 2000; Carr and Kraus, 2001; Gaudiano and Kana, 2001), these features correspond to the main transport pathways from the updrift beach of the delta towards the downdrift complex (Figure 2). The updrift lateral bar receives mainly sediment from the longshore transport \( LST_{\text{BA}} \) in Figure 2). From there, sediment is transported towards the outer shoal \( Q_{\text{BO}} \) and within the inlet channel \( Q_{\text{BI}} \) by wave and current actions. The inlet channel also receives sand exports from the Guadiana River estuary \( RE \). The material from both the updrift lateral bar and the inlet channel feeds the outer shoal \( Q_{\text{BO}} \) and \( Q_{\text{IO}} \), respectively, from where sand is transported towards the downdrift complex \( Q_{\text{out}} \), considered as the final sediment sink (since the focus is on the outer shoal evolution only). In the Results section, the input rates from the local sources to the outer shoal \( Q_{\text{IO}} \) and \( Q_{\text{BO}} \) are derived from observed volume variations of the inlet channel and lateral bar. Likewise, the average river export \( RE \) is established based on the refinement of previous estimates from a numerical model, updated with the results from more recent observational surveys. \( Q_{\text{out}} \) is not reported as it is not required to estimate \( LST_{\text{BA}} \).

The above described sediment transport pathways are considered as the main ones across ebb-tidal inlets. Former morphodynamic analyses of the historical delta collapse (after jetty installation) based on aerial photographs revealed some updrift transport from the downdrift complex towards the inlet channel, with the westward extension of sand banks that started to overtop the eastern jetty in the early 2000s (Garel et al., 2014). This transport cannot be quantified in this study due to map coverage limitations, but is probably relatively weak due to sheltering from wave actions by the jetties and by the extended shoals of the historical delta (see Figure 1). Inputs from the downdrift complex back into the channel are therefore considered minor.

### Table I.

| Year | Source    | Data                        | \( \Delta z = z(t) - z(t-1) \) |
|------|-----------|-----------------------------|---------------------------------|
| 1969 | MPW       | Digitalized topo-bathymetric map 1/5000 | —                              |
| 1977 | IPTM      | Digitalized topo-bathymetric map 1/5000 | +0.05                          |
| 1982 | IPTM      | Digitalized topo-bathymetric map 1/5000 | +0.25                          |
| 1986 | MPW       | Digitalized topo-bathymetric map 1/5000 | −0.10                          |
| 1988 | MPW       | Digitalized topo-bathymetric map 1/5000 | +0.20                          |
| 1992 | IPTM      | Digitalized topo-bathymetric map 1/5000 | 0                              |
| 1995 | IH        | 50 m grid size              | +0.5                           |
| 2005 | IH        | 25 m grid size              | 0                              |
| 2010 | IH        | 25 m grid size              | 0                              |
| 2014 | UAlg      | SB, 50 m transect interval  | 0                              |
| 2015 | UAlg      | SB, 50 m transect interval  | 0                              |
| 2016 | UAlg      | SB, 50 m transect interval  | 0                              |
| 2017 | UAlg      | SB, 50 m transect interval  | 0                              |

**Notes:** MPW = Ministry of Public Works, Hydrography Section; IPTM = Port and Maritime Transport Institute; IH = Hydrographic Institute; UAlg = University of Algarve; SB = Single-Beam echo-sounder. \( \Delta z \) (m) indicates the difference from the previous bathymetry at deep areas.

![Figure 2. Map of the Guadiana ebb-tidal delta in 1986 (a) and 2016 (b) showing the location of the outer shoal, updrift lateral bar and inlet channel areas (red, green and blue boundaries, respectively). The contour interval is 0.10 m. The grey arrows indicate the main transport pathways over the delta (see text for the definitions of \( LST_{\text{BA}} \), \( RE \), \( Q_{\text{BI}} \), \( Q_{\text{BO}} \) and \( Q_{\text{out}} \)). The offshore distance of the outer shoal is established along the yellow dashed line extending from the tip of the western jetty. The location of the cross-shore profile used to compute the LST with the bulk expressions is indicated with a black line (Tr). For general localization of the study area, see Figure 1. [Colour figure can be viewed at wileyonlinelibrary.com]"
in comparison with the main contributors: the eastward LST$_{BA}$ and river exports RE (Figure 2). Likewise, the rather constant morphology and isobaths seaward of the outer shoal suggest that sand transport to deeper water is negligible.

The three elements defined above were already clearly identified in 1977 (i.e. the first map available after jetty installation; see Supporting Information). Their boundaries have variable positions through time, identified based on the morphostructural patterns of the maps with 0.1 m contour lines (see Figure 2). The external limits of both the updrift lateral bar and the outer shoal were considered as the −3 m isobaths, which delineate well the seaward border of the delta on each map. The outer shoal corresponds to the area that includes the subparallel lobate swash bars off the mouth. It is bounded landward by the inlet channel, which rises sharply seaward in the vicinity of the outer shoal. The precise limit between both areas is marked by the gentler channel thalweg slope that generally corresponds to the foot of the most landward (lobate) bars of the outer shoal. The outer shoal is limited laterally by the updrift (in the west) and downdrift (in the east) areas. These boundaries correspond to the transition from the lobate bars of the outer shoal to the broader, shallower and straighter bars of lateral areas. The eastern limit (between the outer shoal and the downdrift areas) is also marked from the 2000s by a sharp change in the bars’ orientation (e.g. Figure 2b). The landward limit of the updrift lateral bar is parallel to the beach, and corresponds to the −1 m isobaths along the beach face not affected by the presence of the delta (e.g. west of Tr in Figure 2); this is the location closest to the beach, which is common to all the maps. Due to cross-shore transport, the beach accreted substantially during the initial growth of the modern delta, along with jetty impoundment, and has been relatively stable since (see Figure 5a in Carel et al., 2015); all the littoral drift from the updrift beach is therefore considered to reach the ebb delta. In the east, the updrift lateral bar is limited by the inlet channel area. The border between these areas is the line drawn between the jetty tip and the northwest corner of the outer shoal area (see Figure 2). In this way, all the bars observed on the updrift side of the inlet channel are included in the updrift area. The inlet channel area extends seaward until the outer shoal, and is limited laterally by the jetty, updrift lateral bar and downdrift complex (defined by the line drawn between the eastern jetty tip and the outer shoal). The northern limit of the inlet channel area is common to all maps, except that of 2017, which has a limited extension landward between the jetties (see Supporting Information) and for which the inlet channel volume was thus not computed. This limit was set to include in the area the portion of the O’Brill bank located in 1969 between the future jetties (see black dashed line in Figure 1), and scoured by jet currents. Finally, following Stable (1998), the offshore migration of the ebb delta is measured along the line extending seaward from the western jetty until the external boundary of the outer shoal (Figure 2, yellow dashed line). All the bathymetric maps with indication of the selected areas are reported in the Supporting Information.

The exact boundaries of the above-defined morphological elements are intricate and diffuse, inducing some uncertainties in the (volume and area) estimates (Rosati, 2005). Various tests have shown, however, that slight modifications of the boundaries’ positions do not significantly affect the results. Furthermore, uncertainties associated with volume estimates depend on the data acquisition (accuracy, extent of coverage and density of points) and post-processing (Kraus and Rosati, 1999), for which information was generally not available. For example, significant vertical shifts (up to 0.5 m) were observed for some consecutive maps before 2005, based on water depth comparisons at the deepest and furthest areas from the delta, supposed to be the most stable (Table 1); it is not known if these shifts are due to vertical errors or (at least partly) to genuine changes in bed elevation. Overall, these potential vertical errors should be smoothed and have minor effects on the results, since the morphological analysis focuses on the long-term average (45 years) evolution of the ebb delta.

**LST from bulk equations**

For comparison with the results from the sediment budget analysis (LST$_{BA}$), the longshore transport rate was computed through a cross-shore beach profile located in the west of the updrift lateral bar (see Tr in Figure 2), using various bulk equations: (1) the commonly used Kamphuis (1991) formula; (2) the well-known CERC (1984) equation using the recommended value of $K = 0.39$, where $K$ is a sediment transport coefficient that accounts for all the parameters that are ignored by the equation, such as sediment and beach properties; (3) the CERC (1984) equation where the coefficient $K$ was related to $D_50$ based on the relation derived empirically by del Valle et al. (1993), considering the average $D_50$ (0.35 mm) near Tr (obtained from the Iberian Atlantic Margin Sediments Database; see Costas et al., 2018); and (4) the enhanced CERC formula proposed by Mil-Homens et al. (2013) through optimization algorithms applied to an extensive dataset. Hourly LST rates were obtained for the period 1972–2017 using the equations mentioned in (1)–(4) above and referred to as LST$_{Kam}$, LST$_{CERC1}$, LST$_{CERC2}$ and LST$_{CERC3}$, respectively.

To obtain the above LST values, a downsampling technique satisfactorily applied in other coastal areas was implemented (for details, see Bergillos et al., 2016; López-Ruiz et al., 2018a, 2018b). A database of 500 representative deep-water wave conditions was defined using the hindcast data from SIMAR point 5026021 (see Figure 1). These sea states were propagated to the nearshore to obtain the wave-breaking conditions using the SWAN model (Booij et al., 1999) over a computational domain (see red area in Figure 1 inset), defined by a single curvilinear grid with variable resolution ranging between ~60 × 60 m at the open ocean to 15 × 15 m at the coastal area. All the cases of the database were propagated over the 2016 bathymetry only, due to limited data availability at deep-water depths and in the vicinity of Tr; the implications of such simplification are discussed in the Comparison with Predictions from Bulk Equations section. To compute the LST, the complete dataset of breaking wave conditions between 1972 and 2017 was reconstructed by means of interpolation based on radial basis functions. The yearly LST estimates obtained from hourly data range from the beginning of July of the preceding year to the end of June of the actual year, in order to match the acquisition of the bathymetric data (generally in summer, although not always known) and to integrate the entire maritime winter (from October to March) in the same yearly record. For example, the LST rate for year 2010 was estimated based on wave conditions from 1 July 2009 to 30 June 2010.

**Results**

**Morphometric evolution of the ebb delta**

The morphometric evolution of the delta is detailed to correctly define the sediment pathways and volume exchanges between the outer shoal, inlet channel and lateral bar areas. The initial modern delta development was marked by fast (~80 m year$^{-1}$) offshore migration between 1977 and 1982 under ebb jet action, slowing down to ~7.5 m year$^{-1}$ afterwards (Figure 3a). All three elements also featured a rapid evolution during the first decade after jetty construction. In particular, ~2 Mm$^3$ of
sand had been scoured in 1986 by ebb jets to form the inlet channel (Figure 3b, blue). The scoured material has contributed to the rapid growth of the outer shoal, which was more than 1.5 Mm³ in 1982 (Figure 3b, red). The updrift lateral bar development was less significant and slower, with the accumulation of ~1 Mm³ from 1969 to 1988 (Figure 3b, green). From the 1990s onward, each of the three elements displayed distinctive volume evolution trends: the outer shoal has been growing continuously, the updrift lateral bar has been slightly eroding, and the inlet channel has remained relatively stable (Figure 3b).

For example, the volume compared to the mean for the period 1995–2017 increased from −25 to +20% at the outer shoal, decreased from +50 to −30% at the updrift lateral bar, and varied by less than 10% at the inlet channel area. Some points out of the trends, such as the volume decrease of the lateral bar in 2005, are attributed to short-lived natural shifts of the delta from its average evolution over longer time scales.

Similar to volume variations, the areas of the three elements show a rapid initial development until ~1990, before stabilizing (Figure 3c). For example, variations were generally less than 10% of the mean for the period 1995–2017. This indicates that the volume evolutions of the outer shoal (growth) and updrift lateral bar (decrease) after the 1990s were mainly achieved through vertical changes (accretion and erosion, respectively) rather than area variations, despite the ongoing delta offshore migration. Concordantly, the average vertical changes and volumetric evolution trends are similar for both elements (Figures 3b and d). The only significant divergence is observed during the initial development of the outer shoal (until 1988), when the average sand deposit decreased (Figure 3d, red) due to a redistribution of the sand from the O’Bril bank over an area expanding rapidly in deeper water (Figure 3c, red). In 2017, more than 3 m of sand, on average, was deposited on top of the pre-jetty surface at the outer shoal (Figure 3d). For the same period, the updrift area eroded by ~0.7 m, decreasing in volume from ~1 Mm³ (1988) to ~0.5 Mm³ (2017). This erosion corresponds to the reworking of a relatively broad and shallow shoal (relict of the O’Bril bank), resulting in a better defined (straighter and narrower) updrift lateral bar linking the beach to the outer shoal (compare 1982 and 1986 in the Supporting Information). Finally, it is also noted that the volumes of dredged material in 1987 and 2015 were too low (e.g. 0.063 Mm³ in 2015, corresponding to a vertical average shift of ~0.02 m over the outer shoal) to have a notable impact on the volume evolution of the delta morphological elements (see Garel, 2017).

Sediment budget analysis

Transport pathways
The above morphometric results indicate that the inlet channel area did not store sediment after jetty installation (scouring of ~2 Mm³ in 1972–1986, stable volume afterwards). Thus, the material that reached this element (see RE and Q_{io} in Figure 4) was transported directly to the outer shoal. Along with the sediment scoured locally (Q_{re} in Figure 4), these contributions represent the total sediment inputs from the inlet channel to the outer shoal (Q_{io} in Figure 4).

Regarding the updrift lateral bar area, morphometric observations (Figure 3) indicate an initial period of growth (until 1988), followed by erosion. Downdrift and updrift lateral bars are usually fed with sand transported from the outer shoal by wave action, and thus develop after the outer shoal has reached a sufficient size (Kraus, 2002). In addition, at settings with a predominant LST direction, updrift wave-induced transport is generally too weak to produce an updrift lateral bar (Kraus, 2002, 2009). The observed early development of the updrift area at the study site is atypical and derives from the reworking of the O’Bril bank (providing a large local sand supply) rather than longshore transport (Garel et al., 2015). Since this growing phase is followed by erosion, it can be considered that all the littoral drift arriving at this area from the updrift beach) was transported to the inlet channel and outer shoal (see LST_{up} in Figure 4). The contribution of the updrift lateral bar to the outer shoal (Q_{up} in Figure 4) consists of a fraction of the LST_{up} plus a fraction of the material eroded locally (Q_{re} in Figure 4). The other fractions are transported to the inlet channel and correspond to Q_{io} (Figure 4). Since the inlet
application of the IRM, providing an average rate for the study period (1972–2017). Thus, the other assessed rates (RE, Qin, and Qout) are also averaged for the same period, even though some temporal variability has been observed (e.g. inlet channel scouring Qsc occurred during the first decade after jetty installation).

Sediment transport rates

The average sediment input to the outer shoal (Qin) over the period 1972–2017 is estimated as the best fit between observations and volume predictions using Equation (1). The best fit \( r^2 = 0.92 \) corresponds to an equilibrium volume (\( V_e \)) of 3.16 Mm\(^3\) with 75% prediction bounds of \(-2.8\) and 3.5 Mm\(^3\). The best estimate of Qin is \( 160 \) 000 m\(^3\) year\(^{-1}\) with 75% prediction bounds of \(\pm 103\) 000 and 223 000 m\(^3\) year\(^{-1}\) (Figure 5).

According to the results depicted in Figure 3, a volume of \(\sim 2\) Mm\(^3\) was scoured by ebb jets during the first 10 years of the inlet channel development. Over the study period (45 years), this scouring represents an average sediment transport rate Qsc of \(-44\) 000 m\(^3\) year\(^{-1}\). Furthermore, the volume of the updrift lateral bar has been reduced by \(-0.5\) Mm\(^3\) since the 1990s, corresponding to an average transport rate Qin,sc of \(-11\) 000 m\(^3\) year\(^{-1}\) from 1972 to 2017.

The volume of sand delivered from the estuary to the ebb delta (RE) is poorly constrained and depends largely on the river discharge and vertical stratification of estuarine waters. Observations suggest that sand export is significant for discharges \(>1500\) m\(^3\) s\(^{-1}\) (i.e. when the entire estuary is filled up with freshwater) but weak for lower rates. For example, during a discharge event of \(1500\) m\(^3\) s\(^{-1}\), a salt wedge developed at the mouth of the estuary, promoting near-bed flows oriented upstream such that the sediment export was of the same order of magnitude as for low-flow conditions (Garel and Ferreira, 2011). Based on an uncalibrated 2D model, an upper limit for RE of \(100\) 000 m\(^3\) year\(^{-1}\) was proposed by weighting the simulated yearly sand transport for various river discharges to the frequency of occurrence of such events (Portela, 2006). This value seems large compared with other similar long and narrow systems. For example, at the Saco estuary the yearly sand export is \(10\) 000–16 000 m\(^3\), mainly during spring freshets that can last for weeks (Kelley et al., 2005). Part of the overestimation might be due to stratification effects, not reproduced with the 2D model. Furthermore, the predicted export rate for low discharge conditions (\(30\) 000 m\(^3\) year\(^{-1}\)) is questionably high, being for example one order of magnitude larger than more recent estimates (\(\sim 5000\) m\(^3\) year\(^{-1}\)) based on cross-channel velocity measurements (Garel and Ferreira, 2011). Finally, since 2002 the river inflow has been strongly regulated by the large Alqueva dam (Garel and D’Alimonte, 2017), reducing the sand export by at least one order of magnitude (Garel and Ferreira, 2011). Following Portela’s weighting method (see Table 1 in Portela, 2006) but with the latter (low) rate for either any river discharge below \(1500\) m\(^3\) s\(^{-1}\) or after 2002, the yearly export is \(\sim 20\) 000 m\(^3\) year\(^{-1}\). This value is clearly a rough estimate, but considered more reasonable than previous assessments.

Based on the above quantifications, the average longshore transport reaching the delta between 1972 and 2017 (LSTBA) is derived from Equation (2). Accounting for the local contributions Qin and Qout, and for the river export RE, the LSTBA best estimate is on the order of \(85\) 000 m\(^3\) year\(^{-1}\). This rate is in good agreement with expectations for the region of \(\sim 110\) 000 m\(^3\) year\(^{-1}\) at maximum (Santos et al., 2014). Considering the 75% uncertainty level for Qin, the predicted rate is constrained to an upper limit of \(-145\) 000 m\(^3\) year\(^{-1}\).
Comparison with Predictions from Bulk Equations

The yearly longshore transport rates predicted by the bulk equations share similar trends, characterized by a predominant eastward transport, as expected, with large-magnitude fluctuations (Figure 6). An episode of westward transport is noted in 2016–2017. The average (1972–2017) \( LST_{\text{CERC}} \) is \(-110,000 \text{ m}^3\text{ year}^{-1}\) (Table II), relatively similar to the best estimate from the budget analysis. By contrast, the average \( LST_{\text{CERC}} \) rate is approximately two times larger, \(-190,000 \text{ m}^3\text{ year}^{-1}\) (Table II). It is also noted that the maximum \( LST_{\text{CERC}} \) rate (\(-500,000 \text{ m}^3\text{ year}^{-1}\)) generally corresponds to coasts exposed to energetic wave conditions, such as the monsoon-affected east coast of India (Komar, 1983) or the southeast Queensland coast, Australia (Splinter et al., 2012). \( LST_{\text{CERC}} \) (using the Mil-Homens expression) is approximately half of \( LST_{\text{CERC}} \), with a mean of \(-40,000 \text{ m}^3\text{ year}^{-1}\) (Table II). With the Kamphuis expression, the average transport rate \( \langle LST_{\text{Kamp}} \rangle \) is one order of magnitude greater than \( LST_{\text{CERC}} \), with a mean above 2 \( \text{Mm}^3\text{ year}^{-1}\) (Table II; not shown in Figure 6).

The above results exemplify the large variations in bulk equation predictions, as previously reported in other studies (e.g. Güner et al., 2011; Mil-Homens et al., 2013). Amongst the four equations considered, \( LST_{\text{CERC}} \) gives the results that are best in line with expected values in the region and predictions from the budget analysis. It is noted that this equation attempts to represent basic physical processes governing longshore transport without considering the morphodynamic characteristics of the study site. In the formula, these characteristics are embedded in the sediment transport coefficient \( K \), which is a meaningless number devised to achieve agreement between observed and computed LST volumes (Cooper and Pilkey, 2004). Since \( K \) accounts for parameters which are both temporally and spatially variable, the concept entailed in the CERC1 formula of a single \( K \) value universally applicable to all beaches is not reasonable.

By contrast, in the other equations, morphodynamic effects are accounted for through adjustment of the coefficient \( K \) based on the mean grain size in the expression of del Valle et al. (1993) and on wave properties in the expression of Mil-Homens et al. (2013) and through input of both the grain size and bed slope in the expression of Kamphuis (1991). Besides, Komar (1988) found that the \( K \) parameter of the original CERC expression was only slightly dependent on the grain size and interpreted this as an indication of the bad quality of the available data used to develop the equation. Many other authors have pointed out that the CERC formula provides only order-of-magnitude accuracy without calibration data (Fowler et al., 1995; Wang et al., 1998). It is therefore probable that the similarity between the \( LST_{\text{BA}} \) and average \( LST_{\text{CERC}} \) prediction obtained in this study is coincidental.

A source of inaccuracy in LST rate predictions from bulk equations might come from the input wave parameters. The accuracy of these parameters depends strongly on the precision of the topo-bathymetry used for wave propagation. In the present study, only one bathymetry (2016) was implemented in the SWAN model. However, the wave propagation processes were altered by a submerged nearshore bar welding to the beach during the initial development of the modern delta (until ~1995; see Supporting Information and Garel et al., 2015). Moreover, the simplified bulk formulations fail to account for morphological features on the beach that affect the breaking type of waves and accuracy of LST rate predictions (Kamphuis and Readshaw, 1978; Smith et al., 2009). As an example, if plunging breaking were more frequent than spilling for the more energetic conditions which in turn generate the larger transport rates, then one of the main hypotheses of the bulk expressions (spilling breaking) would not be fulfilled (Smith et al., 2009). Yet, the results from bulk equations are still highly variable when averaged for the period 1995–2017, characterized

Table II. LST rates \((\text{m}^3\text{ year}^{-1})\) obtained from bulk equations (1)–(4) for the period 1972–2017 and 1972–1980

| Period     | \( LST_{\text{CERC1}} \) | \( LST_{\text{CERC2}} \) | \( LST_{\text{CERC3}} \) | \( LST_{\text{Kamp}} \) |
|------------|--------------------------|--------------------------|--------------------------|--------------------------|
| 1972–2017  | 109 763                  | 187 717                  | 37 393                   | 2 147 196                |
| 1995–2017  | 125 432                  | 214 513                  | 41 826                   | 2 530 466                |

Figure 6. Yearly longshore transport \((10^3 \text{ m}^3\), positive eastward) from 1972 to 2017 predicted with the bulk equations \( LST_{\text{CERC1}} \) (green), \( LST_{\text{CERC2}} \) (blue) and \( LST_{\text{CERC3}} \) (red) at location Tr (see Figure 2). The dashed lines represent average values. \( LST_{\text{Kamp}} \) is one order of magnitude larger and not represented. [Colour figure can be viewed at wileyonlinelibrary.com]
by a more typical beach morphological setting (Table II). It is also noted that although the model was not calibrated, the uncertainties obtained for the bulk expressions are much larger than the possible model corrections (e.g. by fitting the friction coefficient). The large variability of LST predictions highlights the difficulty of selecting the most adequate bulk equation to be used at a site without additional information to constrain the results.

By contrast, the decadal LST\(_{BA}\) rate is obtained from a ‘response approach’ integrating all potential variables such as morphology, grain size, wave climate, etc. It should therefore be more representative of actual conditions than bulk equations. Application of the method at the Guadiana ebb delta provided a realistic rate of ~85 000 m\(^3\) year\(^{-1}\), despite the simple delineation of the sediment transport pathways over the delta (Figure 4). More detailed studies might consider gross longshore transport rates to represent more accurately the sand inputs to the inlet (Bodge, 1993), along with more complex morphological features and sediment pathways for IRM implementation (e.g. Dabees and Kraus, 2008). In any case, sediment budget analyses typically require estimation of quantities that are not very well known (Kraus and Rosati, 1999) and uncertainties should be evaluated (Rosati, 2005). With the proposed approach, the accuracy of sediment transport rates depends largely on the amount and quality of the bathymetries used to quantify the volumes of each morphological element. In particular, water depth accuracy is dependent on many potential errors present in the measurement process (Byrnes et al., 2002). Because of the great differences in survey types and resolution, it is generally not possible to assign a single uncertainty to volume changes (Hicks and Hume, 1997). However, it can be assumed that these errors are random and tend to cancel out when considering a large number of surveys. Alternatively, the present reverse application of the IRM allows us to specify prediction bounds that define a level of uncertainty for the fitted parameters (in particular \(Q_{BA}\)). The choice of a reasonable confidence interval is arbitrary and depends on the dataset (in particular, small datasets will have very wide intervals). For example, a level of 75% is considered here because this interval width includes most observations over the study period (Figure 5, thin dashed lines). To reduce at maximum the uncertainty of the results (through the fit of Equation (1) with observations), the method should therefore include the largest number of bathymetries from the initial development of the delta until it reaches equilibrium in volume. As such, the approach is designed to estimate the transport rate averaged over decades, where shorter time variability is smoothed out. Nevertheless, the results can be used to adjust the \(K\) parameter of the CERC (1984) equation for a specific site in order to address variability over short (e.g. yearly) time scales, as exemplified in the next section.

**LST Yearly Variability**

Since the coefficient \(K\) of the CERC (1984) expression depends on various physical parameters (Rosati et al., 2002), improved LST predictions at a particular site can be obtained if the coefficient is calibrated with data from observations (Bodge and Kraus, 1991; Wang et al., 1998; 2002; Smith et al., 2009). The usage of a constant \(K\) value for a site is questionable because this parameter implicitly accounts for factors that vary through time (see Cooper and Pilkey, 2004). However, matching the CERC estimations (LST\(_{CERC}\)) with the average rate derived from the sediment budget analysis (LST\(_{BA}\)) yields a \(K\) coefficient that is representative of the mean conditions at a decadal time scale. This ‘decadal’ \(K\) coefficient might be distinct from those representative of the annual conditions that drive the yearly LST. Yet, such tuning of \(K\) based on decadal observations provides a mean to quantify more precisely the yearly LST variability in the vicinity of ebb-tidal deltas, especially when compared to the results of generic (non-adjusted) bulk equations. This approach is exemplified based on the results of the present study (although the CERC1 equation and the sediment budget analysis yield similar rates, i.e. the tuned \(K\) should be close to 0.39).

The average LST was computed from 1972 to 2017 at location Tr using the classical CERC (1984) equation, with \(K\) values ranging from 0.1 to 0.6. For an LST\(_{BA}\) rate of 85 000 m\(^3\) year\(^{-1}\), the corresponding (tuned) \(K\) is 0.3. The yearly rate predicted by the adjusted equation (LST\(_{CAL}\)) varies between ~245 000 and ~500 000 m\(^3\) year\(^{-1}\) eastward, and equals ~25 000 m\(^3\) westward in 2016–2017 (Figure 7). Such high yearly variability may explain the large range of LST rates proposed previously in this coastal area.

It is noted that single years with strong transport (>200 000 m\(^3\) year\(^{-1}\)) alternate with calmer periods (<100 000 m\(^3\) year\(^{-1}\)) lasting several years. A similar pattern of peak variability has been reported in previous studies of wave activity in the region, and related to the North Atlantic Oscillation (NAO; see Almeida et al., 2011; Plomaritis et al., 2015). The NAO has long been known to affect climate variability in the Northern Hemisphere (Hurrell, 1995) and, consequently, the wave climate arriving at the west coast of Europe (e.g. Bacon and Carter, 1993; Dodet et al., 2010; Martínez-Asensio et al., 2016), in particular in winter (e.g. Bromirski and Cayan, 2015). Likewise, previous studies have reported relations between large-scale climate indices and yearly LST estimates, for example in East Australia (Splinter et al., 2012). At the study site, Figure 7 features a good correspondence between periods with high eastward LST\(_{CAL}\) and negative peaks of the NAO index (DJFM average; Hurrell and NCAR Staff, 2017) below an approximate threshold value of −1. For example, the five largest transport rates (>200 000 m\(^3\) year\(^{-1}\)) of the time series are associated with the lowest NAO index values, between −1.9 and −4.6 (only the low index of 1977 is not associated with a large LST rate). Notably, the largest LST\(_{CAL}\) rate (in 2010) corresponds to the lowest NAO index value. Such an NAO index threshold has been reported previously in relation to large wave activity in the Gulf of Cadiz (Plomaritis et al., 2015), confirming that the magnitude of longshore transport is mainly controlled...
by storms. Along the western coast of Europe, modulation at multi-decadal time scales of the longshore sediment transport by the atmospheric circulation in the north Atlantic has been evidenced at wave-built barriers (Costas et al., 2016; Poirier et al., 2017). The present analysis verifies that climate variability may also affect significantly the longshore transport rates at a yearly scale. Future studies should evaluate how this temporal variability affects the coastal evolution in the region.

Conclusions

The present study proposes a method to estimate average (decadal) LST rates in the vicinity of modern ebb-tidal deltas. A sediment budget analysis is conducted to specify the sediment transport pathways towards the outer shoal and the associated rates (other than LST). These contributions are balanced with the total sediment input rate to the shoal, estimated based on an inverse application of the IRM, yielding a best estimate of the average LST rate during the study period along with upper bounds (at confidence levels) that are often desirable for the design of management and planning strategies.

The method has been applied to the 45-year (1972–2017) development of the modern ebb-tidal delta that followed jetty installation at the mouth of the Guadiana estuary. The results indicate an average LST rate of ~85 000 m$^3$ year$^{-1}$ (best estimate), in good agreement with expectations for the region. Considering a 75% uncertainty level for $Q_{br}$, the rate is ~145 000 m$^3$ year$^{-1}$ at maximum. For comparison, the average LST predictions from bulk equations for the same period vary by more than one order of magnitude, illustrating their well-known large variability. Nevertheless, the average (decadal) rate from the sediment budget analysis can be used to adjust the K coefficient of the CERC (1984) equation to local conditions for the quantification of LST rate variations at shorter time scales. As an example, it is shown that yearly absolute rates at the study site vary by a factor of ~50 (5000–245 000 m$^3$ year$^{-1}$) in relation to NAO index variations.

The method is applicable to ebb-tidal deltas where the main transport pathways towards the outer shoal can be specified. It is also required that the outer shoal development from an early to a relatively mature state is well documented by bathymetric maps. Hence the approach applies to ebb-tidal deltas that developed – or were largely disrupted – relatively recently. Many of such systems can be found worldwide. Examples include ebb deltas subject to extensive dredging of the outer shoal, affected by jetty or breakwater installation or created by (artificial or natural) barrier breaching (e.g. Hansen and Knowles, 1988; Dabees and Kraus, 2008; Poirier et al., 2017; Castelle et al., 2007; Buonaiuto et al., 2008; Dabees and Kraus, 2008; Patsch and Griggs, 2008; Kaminsky et al., 2010; Cox and Howe, 2012; Beck and Wang, 2019). Overall, the approach provides an alternative means to constrain the average LST rate over decades that smooths out any shorter-term variability. Where the modern delta development is particularly well documented (e.g. at yearly frequency), the approach could be implemented to estimate the average LST at shorter time scales. Finally, with the ongoing development of satellite-derived bathymetry, the method could be applied to an increasing number of sites where echo-sounding surveys are not conducted frequently.

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**Supporting Information**

Additional supporting information may be found online in the Supporting Information section at the end of the article.

**Data S1.** Supporting information