The morphology of fluvial-tidal dunes

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ABSTRACT

This paper quantifies changes in primary dune morphology of the mesotidal Lower Columbia River (LCR), USA, through ~ 90 river kilometres of its fluvial-tidal transition. Measurements were derived from a Multibeam Echo Sounder dataset that captured low-river stage bedform dimensions within the thalweg (≥ 9m depth) of the LCR main channel with respect to fluvial-tidal current interactions. Measurements revealed two categories of dunes: i) fine to medium sand ‘fluvial-tidal to tidal’ (upstream-oriented, simple, and 2D) low-angle dunes (heights ≈ 0.3-0.8m; wavelengths ≈ 10-25m; mean lee-angles ≈ 7-11°), and ii) medium to coarse sand ‘fluvial’ (downstream-oriented, compound, and 2.5-3D) low-angle dunes (heights ≈ 1.5-3m; wavelengths ≈ 60-110m; mean lee-angles ≈ 11-18°). Approximately 86% of the fluvial-tidal transition is populated by ‘fluvial’ dunes, whilst only ~ 14% possesses ‘fluvial-tidal to tidal’ dunes that form in the downstream-most reaches. River currents are thus the first-order control governing dune morphology, with tidal-currentsexerting a second-order influence (especially in the downstream part of the transition zone). Two mechanisms are reasoned to dictate their low-angle character: (1) high-suspended sediment transport near peak tidal-currents that lowers the leeside-angles of ‘fluvial-tidal to tidal’ dunes, and (2) superimposed bedforms that erode the crests, leesides, and stoss-sides, of ‘fluvial’ dunes, which results in the shallowing of leeside-angles. Fluctuations in river discharge creates a ‘dynamic morphology reach’, spanning river kilometres 12-40, which displays the greatest variation in dune morphology. Similar channel reaches likely exist in fluvial-tidal transitions with similar physical characteristics as the LCR and may provide a distinct signature of their fluvial-tidal transition.

Keywords: Lower Columbia River, fluvial-tidal bedforms, low-angle dunes, superimposed bedforms

1 INTRODUCTION

Bedforms are ubiquitous within subaqueous environments and are generated by unidirectional, short to long period oscillatory, and combined-flows (currents with unidirectional and oscillatory components), which deform a mobile bed through erosion and deposition of sediment. For centuries, laboratory, field and theoretical investigations have focused on bedform genesis, morphological equilibrium, and their depositional structures (e.g., Du Buat, 1786; Blasius, 1910; Kennedy, 1969; Harms et al., 1975; Allen, 1983; Southard, 1991; Baas, 1994; Kleinhans, 2001; Venditti, Church, & Bennett, 2005a,b; Doucette & O’Donoghue, 2006; Reesink & Bridge, 2009; Perillo et al., 2014a,b,c; Bradley & Venditti, 2019a,b). One of the most common bedforms are dunes (Best, 2005; Venditti, 2013), whose strata represent a fundamental building block of the rock record (Ashley, 1990; Myrow and Southard, 1991; Myrow, Fischer, & Goodge, 2002; Bridge, 2003; Martinius & Van den Berg, 2011; Reynaud & Dalrymple, 2012). Dunes possess heights of 0.075 to > 5m, wavelengths from 0.6 to > 100m, and can be compound (possessing crests, stoss-sides, or leesides, populated with smaller-scale superimposed bedforms) or simple (lacking superimposed bedforms) in form (Dalrymple, Knight, & Lambiase, 1978; Dalrymple, 1984; Ashley, 1990). Primary (largest) and secondary (superimposed) dune morphology is a function of their growth, migration, and decay, as controlled by varying current magnitudes and orientations (cf. Dalrymple, Knight, & Lambiase, 1978; Sherwood & Creager, 1990; Dalrymple and Rhodes, 1995; Hendershot et al., 2016) in conjunction with changes in the ratio of bedload, q_{bed}, to suspended-load, q_{sus}, transport rates (q_{sus}/q_{bed}; Amsler & Schreider, 1999; Best, 2005; Hendershot et al., 2016; Bradley & Venditti, 2017; Ma et al., 2017; Naqshband & Hoitink, 2020). Dunes are therefore spatially and temporally dynamic and follow coupled flow and sediment transport hysteresis loops (Allen, 1974, 1976; Martin & Jerolmack, 2013; Parsons & Best, 2013), which result in transient morphologic properties (Dalrymple, Knight, & Lambiase, 1978; Bradley & Venditti, 2019a,b). These properties include height (η), wavelength (λ), aspect ratio (AR), lee- and stoss- side angles (θ_{lee} and θ_{stoss}, respectively), dimensionality (2 to 3D), roundness, symmetry, and scaling of η and λ to a characteristic flow depth, H, which typically is taken as local (mean) depth, H_{mean}, or maximum depth, H_{max}. Thus, capturing how changes in dune morphology induce variations in flow-fields via form drag (Smith & McLean, 1977;

Parsons et al., 2005; Sukhodolov et al., 2006; Guerrero & Lamberti, 2011; Lefebvre et al., 2011), which affects their migration rates and thus bedload transport rates (cf. Nittrouer, Allison, & Campanella, 2008; Gómez et al., 2010; Knox & Latrubesse, 2016; Schippa et al., 2016), is vital towards building more robust hydraulic and morphodynamic models (cf. Sandbach et al., 2018; van de Lageweg & Feldman, 2018; van de Lageweg et al., 2018; Unsworth et al., 2020).

Based upon longitudinal profiles (e.g., θ_{lee}, symmetry, and roundness), past work has divided subaqueous dunes into two categories (Kostaschuk & Villard, 1996, 1999; Venditti, 2013): high-angle dunes (HADs) and low-angle dunes (LADs). High-angle dunes tend to be asymmetric and often are only slightly rounded, with long, gentle θ_{stoss} and short, steep θ_{lee} near, or at, the angle-of-repose (~ 25-30°)They have been speculated to commonly be observed in bedload dominated laboratory flumes and shallow (H < 2.5m) rivers (cf. Venditti & Bauer, 2005; Bradley & Venditti, 2017), but have also been sporadically observed in the tidally-dominated Cobequid Bay, Bay of Fundy, Canada (Dalrymple, 1984). In contrast, low-angle dunes possess mean θ_{lee} below the angle-of-repose (typically ≤ 15° in deep rivers where H ≥ 2.5m, Best & Kostaschuk, 2002; Best, 2005; Kostaschuk et al., 2009; Bradley & Venditti, 2017; and ≤ 10° in estuarine settings, Dalrymple & Rhodes, 1995), and especially in rivers can be more rounded and symmetric. Furthermore, experimental (Best & Kostaschuk, 2002), numerical (Lefebvre, 2019; Lefebvre & Winter, 2016; Lefebvre et al., 2014a,b), and field studies (Smith & McLean, 1977; Kostaschuk & Villard, 1996; Williams et al., 2003; Holmes & Garcia, 2008; Kostaschuk et al., 2009; Bradley et al., 2013; Cisneros et al., 2020), suggest that flow separation over LAD leesides is intermittent (θ_{lee} ~ 10-24°) to non-existent (θ_{lee} < 10°), whilst HADs possess continuous leeside flow separation.

Three modern environments have received the most attention regarding dune morphology: i) relatively deep (H ≥ 2.5 m; see Bradley & Venditti, 2017) rivers (e.g., Harbor, 1998; Nittrouer et al., 2008, 2011; Sambrook Smith et al., 2013; de Almeida et al., 2016; Knox & Latrubesse, 2016; Bradley & Venditti, 2017,; Galeazzi et al., 2018; Kostaschuk & Venditti, 2019; Cisneros et al., 2020), ii) shallow marine combined oscillatory wave-tidal settings (Carle & Hill, 2009; Ernstsen et al., 2010; Barnard, Erikson, & Kvitek, 2011; Lefebvre, Ernstsen, & Winter, 2011; Fraccascia et al., 2016; Wu et al., 2016), and iii) meso- to macro- tidal estuaries or deltas (e.g., Langhorne, 1973; Wright et al., 1973; Dalrymple, Knight, & Lambiase, 1978; Elliot & Gardiner, 1981; Dalrymple, 1984; Aliotta & Perillo, 1987; Harris, 1988; Davis & Flemming, 1991; Sherwood & Creager, 1990; Dalrymple & Rhodes, 1995; Gómez, Cuadrado, & Pierini, 2010; Hendershot et al., 2016). Two recent advances in quantifying dune morphology are measurements of θ_{lee} that focus on the prevalence and physical causes of LADs *vs* HADs (e.g., Dalrymple & Rhodes, 1995; Hendershot et al., 2016; Kostaschuk & Venditti, 2019; Cisneros et al., 2020), and the re-evaluation of existing η and λ to H scaling relations (Bradley & Venditti, 2017, 2019a; Cisneros et al., 2020). Ever since the proposed original empirical scaling relations ($\eta = 0.17H$; $\lambda = 5H$) by Yalin (1964) and Allen (1982), it has been widely adopted (although not universally accepted; see Dalrymple & Rhodes, 1995; Bradley & Venditti, 2017) that dune η and λ scale to boundary layer thickness (Allen, 1968; Ashley, 1990; Southard & Boguchwal, 1990a,b; Best, 2005), which can be no greater than H_{max}. For instance, Bradley & Venditti (2017) included how formative hydraulics and sediment transport processes change as dunes grow with increasing H and modified the original relations to $\eta = 0.13H$ and $\lambda = 5.9H$, whereas dune analyses from multiple rivers by Cisneros et al. (2020) found that the scaling for η should be closer to $\eta = 0.10H$. However, regardless of the scaling relation , modern river and tidal setting field data places the normalised dune η (η/H) and λ (λ/H) within the ranges of ~ 0.025-0.4 and 1-16, respectively (Allen, 1982; Dalrymple & Rhodes, 1995; Venditti, 2013; Bradley & Venditti, 2017; Cisneros et al., 2020).

As the number of mean θ_{lee} (average of leeside angles measured from crest to the bottom of downstream trough) and maximum θ_{lee} (slipface angle) measurements of dunes from modern environments increases, it is apparent that silt- to gravel- bed rivers, estuaries, and deltas are dominated by simple and compound low-angle dunes (e.g., Dalrymple & Rhodes, 1995; Hendershot et al., 2016; Bradley & Venditti, 2017; Ma et al., 2017; Kostaschuk & Venditti, 2019; Cisneros et al., 2020). The physical processes driving their formation, however, remain debated (Best et al., 2020; Best & Fielding, 2019), with mechanisms for compound low-angle dunes being: i) erosion of primary dune crests and leesides via heightened localised bed shear stresses generated by superimposed bedforms (Allen & Collinson, 1974; Allen, 1978; Reesink & Bridge, 2009), and/or restriction of bedload supply to the primary dune crest by superimposed bedforms, which starves them of sediment needed to maintain steep avalanching slipfaces (Carling et al., 2000; Sukhodolov et al., 2006); ii) superimposed bedforms may suppress flow separation over the crests of primary dunes (Dalrymple & Rhodes, 1995); and iii) development of compound dunes at oblique orientations relative to the local flow direction can suppress flow separation due to the apparent leeside angle being smaller than the true leeside angle (Sweet & Kocurek, 1990; Dalrymple & Rhodes, 1995). Whereas the mechanisms for simple low-angle dunes include: i) shift of sediment deposition from dune crests to dune leesides and/or lee-troughs caused by higher rates of suspended-sediment transport , which reduces the sediment available to maintain a steep avalanching slip-face (Smith & McLean, 1977; Bridge & Best, 1988; Kostaschuk & Villard, 1996; Kostaschuk et al., 2009; Hendershot et al., 2016; Bradley & Venditti, 2017; Ma et al., 2017; Naqshband & Hoitink, 2020); ii) downslope lee-face currents resulting from intermittent to non-existent flow separation over dune leesides (Kostaschuk & Venditti, 2019), and iii) liquefied grainflows produced by high excess pore pressure causing failure of dune brinkpoint wedge deposits (Hendershot et al., 2016; Kostaschuk & Venditti, 2019), although this mechanism seems unlikely based on abundant morphological data (Best et al., 2020). There is insufficient evidence to suggest that any one mechanism (for simple or compound low-angle dunes) unequivocally explains their prevalence (see Kostaschuk & Venditti, 2019; Cisneros et al., 2020) or that the proposed mechanisms for simple dunes do not also play some role in compound dune formation. For example, in sand-bed (200-900μm) rivers, simple and compound low-angle dunes coincide with estimated q_{sus}/q_{bed} ratios of > 2.3 (Kostaschuk & Venditti, 2019), whilst in the low Froude number (0.17 < Fr < 0.30) dune experiments of Naqshband & Hoitink (2020) designed to mimic deep rivers, lee-angles progressively lowered via the onset of increasingly higher rates of suspended-sediment transport. Overall, this evidence suggests that the q_{sus}/q_{bed} ratio likely plays some role in both simple and compound low-angle dune formation.

Four principal deficiencies persist in our understanding of the morphology of fluvio-tidal dunes. Firstly, although studies tend to report η, λ, aspect ratios, and θ_{lee}, they rarely quantify dimensionality, roundness, and symmetry, via standardised parameters (cf. Harms, 1969; Arnott & Southard, 1990; Oost & Baas, 1994; Dumas, Arnott, & Southard, 2005; Venditti, Church, & Bennett, 2005a,b; Sekiguchi & Yokokawa, 2008; Perillo et al., 2014a,b,c), but see Dalrymple, Knight, & Lambiase (1978), Dalrymple (1984), Dalrymple & Rhodes (1995), and Hendershot et al. (2016) for exceptions. Secondly, many river and delta studies are spatially restricted to either unsteady and non-uniform 'backwater' channel reaches, or to steady and uniform 'normal flow' reaches (but see Harbor, 1998, for an exception). Thus, few studies capture any of the broader pattern changes driven by upstream to downstream variations in channel bed grain size in conjunction with changes in mean flow hydraulics (see Mei et al., 2021 for exception), and many only report findings acquired over short temporal windows that are fixed to a narrow range of river stages, which makes it impossible to evaluate sediment transport regime and hysteresis effects caused by discharge variations (cf. Bradley & Venditti, 2019a,b; Naqshband & Hoitink, 2020). Thirdly, estuarine studies tend to be spatially confined to the lowermost backwater reaches, which possess relatively low fluvial input and are thus dominated by bidirectional currents (Dalrymple & Rhodes, 1995; Gómez et al., 2010). Furthermore, only a few studies have investigated temporal fluctuations over tidal-cycles (Dalrymple, Knight, & Lambiase, 1978; Sherwood & Creager, 1990; Hendershot et al., 2016), or captured upstream to downstream variations in dune morphology that are linked to changes in grain size and spatio-temporal changes in tidal- and fluvial- energy.

To address several of these limitations, the present study analyses the primary dunes within the deepest (≥ 9m depth from Mean Sea-Level, MSL, or local H/H_{max} ≥ 0.7) portions of the main channel of the mesotidal Lower Columbia River (LCR), OR/WA, USA, using a National Oceanic and Atmospheric Administration (NOAA) multibeam echo sounder (MBES) dataset collected through ~ 90 river kilometres (rkm) of its fluvial-tidal transition (FTT; Dalrymple & Choi, 2007, or fluvial-tidal zone; Van den Berg, Boersma, & Van Gelder, 2007; Figure 1). Since FTTs are sensitive to sea-level rise and are host to a large proportion of the world's population, spatial characterisation of primary dune morphology can inform hydraulic and morphodynamic models used for assessing flooding risk and ecological sustainability and remediation/dredging strategies. Three questions are addressed herein:

1) How does the primary dune morphology in the deepest (H/H_{max} ≥ 0.7) channel vary through the LCR fluvial-tidal transition?

2) Do any variations in characteristics relate to transitions in hydraulic and sediment transport processes, and/or bed grain size?, and

3) How does their morphology differ between tidally-dominated estuaries and unidirectional current-dominated rivers and flume experiments?

2 STUDY SITE: LOWER COLUMBIA RIVER

The Lower Columbia River mean annual discharge, Q_{Wavg}, is ~ 7,000 m^{3}s^{-1} at the Port Westward (Beaver) gauge station (~ rkm 85) near the upstream margin of the study reach (Figures 1 and 2A; Naik & Jay, 2011). Thus, low-river stages possess discharges that are ≤ 7,000 m^3s^{-1}, whilst high-river stages are > 7,000 m^3s^{-1}. Peak modern flows, Q_{Wpeak}, are controlled by dam releases to between ~ 15,000-17,000m^3s^{-1} (Gelfenbaum, 1983; Naik & Jay, 2011; Simenstad et al., 2011), which is lower than pre-1900s 'natural' spring freshets (≥ 18,000m^3s^{-1}; Naik & Jay, 2011; Figure 2A). The LCR channel bed slope, S_c, is ~ 1.15 x 10^{-5} (Hickson, 1912; Figure 2A), and is influenced by mesotidal mixed diurnal and semidiurnal tides. At its mouth, the mean diurnal tidal prism, Q_{tide}, is ~ 11.0 x 10^8 m^3 (Walton & Adams, 1976) and the mean tidal range and highest astronomical tide are 1.7 and 3.6m, respectively, which rise marginally to 2.0 and 4.0m near Astoria, OR, as the result of tidal funnelling (Figure 2A; Fain et al., 2001; Simenstad et al., 2011). Its hydrographic ratio, H_g, defined as $\left(Q_{tide}/Q_{Wavg}\right)\, x\ 6hrs$, is equal to 7 (where H_g < 10 reflects fluvialdominance; Peterson et al., 1984), where this fluvial dominance is also supported by hydrodynamic modelling (Hamilton, 1990; Sandbach et al., 2018). Yet, in contrast to this finding, many oceanographic and geologic studies describe the LCR as an estuary (e.g., Hughes and Rattray, 1980; Gelfenbaum, 1983; Fox et al., 1984; Jay, 1984; Hamilton, 1990; Jay & Smith, 1990; Jay et al., 1990; Sherwood & Creager, 1990; Simenstad et al., 2011; Peterson et al., 2013, 2014). Its geologic designation as an estuary, however, was recently re-evaluated by Prokocki et al. (2015, 2020), who suggest that its mid to late Holocene geomorphology is that of an 'entrenched' fluvio-deltaic environment possessing a subaqueous deltaic top-set extending to ~ rkm 50, thus supporting fluvial dominance, whilst providing a physical explanation for its initial geomorphic designation as an estuary.

With respect to the LCR fluvial-tidal transition, Jay, Giese, & Sherwood (1990) computed the temporally averaged mean flux divergence of fluvial vs tidal potential-energy through modern channel-sections, and used this to divide its fluvial-tidal transition into three hydraulic zones (Figs 2B and 3A): 1) The tidally-dominated regime (TDR; rkm 0-21), or lower delta, which experiences bidirectional tidal-currents, saltwater intrusion, and development of a turbidity maximum (TM; Fox et al., 1984; Sherwood & Creager, 1990); 2) The mixed tidal-fluvial regime (MTFR; ~ rkm 21-56; or upper brackish water delta to lower freshwater tidal river reach; Hoitink & Jay, 2016), where fluvial-currents and bidirectional tidal-currents compete for dominance; and 3) The fluvially-dominated, tidally-influenced regime (FDTIR; ~ rkm 56-235; or mid to upper freshwater tidal river reach; Hoitink & Jay, 2016), which is governed by downstream-oriented currents and terminates at the landward most point of tidally-forced variations in water surface elevation, or tidal limit. However, like all FTTs (Dalrymple & Choi, 2007; Dalrymple et al., 2015), the longitudinal boundaries of LCR regimes are dynamic and expand or contract in accordance with spring to neap tidal-cycles and varying river-stages. For example, when transitioning from high- to low- river stage, the upstream boundaries of the TDR, MTFR, and FDTIR, therefore expand to ~ rkm 35, 109, and 235 (Bonneville Dam; i.e., tidal limit), respectively, which causes (Figure 3A): i) brackish water intrusion to ~ rkm 50 (Fox et al., 1984; Chawla et al., 2008), ii) fluvial current reversals between rkm 50 to ~ 109 (Clark & Snyder, 1969), and iii) tidally-induced cyclic water surface height variations beyond rkm 172 (Vancouver, WA; Kukulka & Jay, 2003). Thus, during low-river stages, ~ rkm 0 to 35 (zones 1-5; hereafter z1-z5) is mainly dominated by tidal flows (Figure 3B), whilst ~ rkm 35-90 (zones 6-11; hereafter z6-z11) experiences a downstream to upstream increase in fluvial-energy as tidal influence diminishes (Figure 3B).

The mean annual sediment-load, Q_s, of the Lower Columbia River where $Q_s = Q_{sand} + Q_{wash}$ (Q_{sand} represents bedload + suspended-load particles (> 63 to ≤ 2000μm), and Q_{wash} represents suspended-load particles ≤ 62 μm), is sourced from tributaries upstream of ~ rkm 172 (Jay, Giese, & Sherwood, 1990; Sherwood et al., 1990; Naik & Jay, 2011), whilst negligible Q_s is derived from continental shelf sources (Gelfenbaum et al., 1999; Templeton & Jay, 2013). Prior to the 1900s, its Q_s is estimated to have been ~ 10 Mtyr^{-1} (Sherwood et al., 1990; Gelefenbaum et al., 1999; Naik & Jay, 2011), but more recent estimates suggest that after 1970 (post-dam era) Q_s has decreased by ~ 70% to ~ 3.2 Mtyr^{-1} (Naik & Jay, 2011). Grain size sampling of deep (H/H_{max} ≥ 0.7) channel-beds (Fox et al., 1984; Sherwood & Creager, 1990), shows that its D_{50} is fine sand (> 125 to 250μm) from ~ rkm 0 to 32, but increases to medium to coarse sands (> 250 to < 1000μm) upstream of rkm 32 (Figure 3B). This D_{50} breakpoint tends to remain spatially fixed during both low- and high- river stages, but during significant river floods, minimal coarsening of the channel-bed from fine to fine-medium sand (> 125 to ~ 275μm) may occur between ~ rkm 16-32 (Sherwood & Creager, 1990).

Previous analyses of primary dunes (≥ 9m depth; H/H_{max} ≥ 0.7) within seaward channel reaches of its FTT (Sherwood & Creager, 1990) recognized three types, whose morphology is summarized in Table 1. The Type-A dunes occurred between ~ rkm 0-9, were simple in form, , developed under significant suspended-sediment transport conditions, and remained tidally-reversing during both low- and high- river stages (Figure 4A, B). Conversely, Type-B were also simple in form and existed between ~ rkm 9-35 during low-river stages, but were restricted to between ~ rkm 9-16 during high-river stages (Figure 4A, B). These dunes also formed within ahigh suspended-sediment transport context and during high-river stages possessed lee-faces that remained downstream oriented over tidal-cycles, but during low-river stages, they maintained upstream orientations over tidal-cycles (Figure 4A, B). The Type-C compound dunes were the coarsest grained and were confined to upstream of ~ rkm 30 during low-river flow but extended downstream to ~ rkm 24 at high-river flow (Figure 4A, B). Throughout all tidal-cycles and river-stages, Type-C remained downstream-oriented and formed within lower suspended-sediment transport conditions relative to Type-A and B. These descriptions (Table 1), however, are limited in several ways: i) the number (N) of dunes analysed is not provided, and observations are restricted to ~ rkm 0-35, ii) η, Λ, and aspect ratios, are reported as general ranges without mean values, and mix both low- and high- river stage observations, iii) dune symmetry, roundness, and dimensionality were not quantified via standardised methods, and iv) fundamental properties such as lee-angles and scaling of η and Λ to flow depth were not measured.

3 METHODS OF ANALYSIS

The low-river stage (< 7,000 m^3s^{-1}; Table 2) bathymetry dataset utilised (~ rkm 0-90, z1-z11; Figures 1 and 3B) represents an integration of NOAA acquired multibeam echo soundings from multiple survey systems, whose integrated maps are gridded at 0.5 or 1m spacing and possess vertical resolutions of ~ 0.05m. For clarity, soundings at sites z1-8 (Figure 1) are normalised to the Mean Lower-Low Water (MLLW) level at the NOAA Astoria tide gauge (ID: 9439040), whilst z9-11 soundings are normalised to the Columbia River Datum, which approximates the extension of the MLLW at Astoria, OR, above ~ rkm 48 (Stolz, Martin, & Wong, 2005). Surveys either occurred (Table 2): i) simultaneously with 2005-2010 dredging of the navigation channel (z1-7, and z9-11), or ii) immediately after 2010 dredging (~ rkm 40-48; z8). Therefore, recently dredged channel reaches outside or within the zones analysed were identified and avoided (see Supporting Information: Type 1 and 2 dredged channel beds) to only evaluate naturally occurring bedforms (Figures 5-8). Next, in each zone, 2D bed transects digitally sub-sampled at 1cm horizontal spacing were taken perpendicular to bedform crestlines. The location of these transects are shown in Figures 5-8, whilst example partial sectional-profiles (A-A' to K-K' in Figures 5-8) are displayed in Figure 9. Note that all transects, and thus bedforms evaluated, occur at depths ≥ 9m (localised H/H_{max} ≥ 0.7; Figures 5-8) to remain consistent with previous Lower Columbia River bedform analyses.

Following Hendershot et al. (2016), individual bed features from all zones were detected automatically within 2D transects using a Matlab script (Perillo et al., 2014a) that identifies consecutive local minimum-maximum-minimum bed elevation points along sectional-profiles (Figure 10A). The η values of all identified bed features were then computed . At this stage, the computed η values of bed features includes both primary and secondary (superimposed) bedforms (Figure 10A). However, to directly compare and integrate the findings of this study with that of the primary dunes examined by Sherwood & Creager (1990), all secondary bedforms were filtered from the zone data. Therefore, a primary dune η cut-off value was established by finding the maximum secondary bedform η within zone transects via hand measurements. For each zone, these values thus represent the η breakpoint between primary dunes and secondary bedforms (Figure 10B and Table 3), where the primary dunes of a given zone must possess an η > the maximum η of secondary bedforms. However, since the η of primary dunes, and thus also the maximum η of secondary bedforms, differs between zones (Figures 9 and 10A), a unique primary dune η cut-off value was utilised for each zone (Table 3). After applying the primary dune η cut-offs, the wavelengths, ʎ, of remaining primary dunes were calculated as the absolute lateral distance between upstream and downstream minimum trough elevations (Figure 10B). The primary dunes of each zone were then further characterised by quantifying the following geometric properties.

Dune aspect ratio was calculated as:

$$\mathrm{AR} = \lambda/\eta \qquad (1)$$

whereas dimensionality (crestline sinuosity) was measured following the approach of Venditti, Church, & Bennett (2005a), who defined a non-dimensional span index, NDS, as:

$$\mathrm{NDS} = \mathrm{L_c}/\mathrm{L_y} \qquad (2)$$

where $\mathrm{L_c}$ represents the actual length of a dune crestline, and $\mathrm{L_y}$ is the straight-line distance between the ends of this same crestline (Figure 11A). Using natural breakpoints within the data, the following values were used to classify the bedform sinuosity:

- NDS < 1.08 = 2D
- NDS ≥ 1.08 < 1.16 = 2.5D
- NDS ≥ 1.16 = 3D

Bedform symmetry and roundness indices (BSI and BRI, respectively) were computed (Tanner, 1967; Perillo et al., 2014a) using:

$$\mathrm{BSI} = \lambda_{\mathrm{stoss}}\lambda_{\mathrm{lee}}^{-1} \quad (3)$$

$$\mathrm{BRI} = \lambda_{0.5_{\mathrm{stoss}}}\lambda_{\mathrm{stoss}}^{-1} \quad (4)$$

where λ_{stoss} is stoss-side length, λ_{lee} is leeside length, and $\lambda_{0.5_{\mathrm{stoss}}}$ represents the length from dune crest to stoss-side at 0.5η (Figure 11B), where the sectional form boundaries were classified as (Perillo et al., 2014a):

- BSI > 1.5 = Asymmetric
- BSI ~ 1.3-1.5 = Quasi-asymmetric
- BSI < 1.5 = Symmetric
- BRI ≥ 0.6 = Rounded
- BRI < 0.6 = Not rounded

Next, maximum and mean θ_{lee} values were determined utilising the procedures and relations given in Figure 11C. Individual measurements were conducted by visualizing singular dune profiles within the Global Mapper software, and then calculating all leeside angles via the slope measurement tool. Recent research (Cisneros et al., 2020; Lefebrve & Winter, 2016) has illustrated the complexity of dune leeside shape, which are often segmented with lower angle slopes both near the crest and trough of the dune. To account for this complexity, the maximum, as well as the mean, leeside angles were quantified as defined in Figure 11c. Lastly, scaling of mean η and λ to mean flow depth, H_{mean}, was assessed, where H_{mean} in fluvial-tidal settings is taken from the Mean Sea-Level (MSL) water elevation.

Tests conducted on data from each zone show that primary dune η, λ, AR, BSI, BRI, and maximum and mean θ_{lee} can be approximated via Gamma distributions (cf. Paola & Borgman, 1991; van der Mark, Blom, & Hulscher, 2008; Cisneros et al., 2020), whilst NDS indices cannot due to the lower number (N) of dunes analysed. Therefore, the zone mean, median, standard deviation, and 25th and 75th quartile values reported herein, for the metrics of η, λ, AR, BSI, BRI, and maximum and mean θ_{lee}, were acquired by applying a Gamma probability density function (PDF) fit to raw measurements, whilst mean zone NDS values and standard deviations were obtained arithmetically from raw measurements.

4 RESULTS: PRIMARY DUNES

Two categories of deep (local H/H_{max} ≥ 0.7) primary dunes were recognised: i) smaller-scale upstream oriented (~ rkm 1-27; z1-5), and ii) further upstream, larger-scale downstream oriented (~ rkm 34-85; z6-11). In total, ~ 1,400 dunes were analysed, with the number of dunes examined per zone and their morphologic attributes being reported in Table 4 and as shown graphically in Figure 12A-F. Herein, the average morphology of these two categories is described within the context of past research as well as: i) varying longitudinal position along the fluvial-tidal transition of the LCR, ii) mean channel depths, and iii) channel bed grain size.

4.1 Upstream oriented dunes: zones 1-5

The z1-5 dunes are positioned within the tidally-dominated regime (~ rkm 1-32) and are composed of fine sand (125-250μm; Figure 3B). The z1 and z2 dunes are potentially a mixture of simple and compound forms, whilst those of z3-5 tend to have simple forms

(Figure 9; A-A' to E-E'). Topographic profiles in isolation, however, cannot distinguish simple from compound dunes (especially those that are compound and possess superimposed bedforms with η ≥ their host; Dalrymple & Rhodes, 1995). Detailed stratigraphic evidence (i.e., presence of simple *vs* compound cross-bedding) is thus needed to unequivocally confirm whether some of the z1-2 dunes are indeed compound in form (Dalrymple, 1984). Without this information, the z1-5 dunes are considered herein as simple in form following the previous observations of Sherwood & Creager (1990; Table 1). Overall, z1-5 dunes are relatively small-scale with mean η and λ ranging from ~ 0.3-0.8m and ~ 13-24m (Figure 13), and thus possess mean aspect ratios of ~ 24-68 (Figure 12A, B). They are predominantly 2D (NDS ≤ 1.08), asymmetric (BSIs > 1.5), not rounded (BRIs < 0.6), and possess near equal mean and maximum θ_{lee} (~ 7-11° and ~ 10-15°, respectively; Figure 12C-F). These dunes are therefore low-angle dunes possessing leesides with relatively mild dipping slip-faces (i.e., lee form 1; Figure 11C), where their mean θ_{lee} is comparable to both sand-bed river and estuarine LADs (≤ 10°, Figure 12F). Overall, they resemble the Type-A and B dunes (Table 1) of Sherwood & Creager (1990), where those of z1 and z2 are interpreted to be tidally-reversing (i.e., Type-A) and those of z3-5 reverse migration in response to the fluvial-hydrograph (i.e., Type-B).

Here, scaling of dune η and λ to H_{mean} shows that η/H_{mean} and λ/H_{mean} ratios span ~ 0.02-0.04 and ~ 0.7-1.2, which fall well below published best fit values for deep rivers and estuaries (Figure 14A, B), and thus plot near, or below, the lower boundary of both tidal and fluvial dunes where η/H_{mean} = 0.025 and λ/H_{mean} = 1.0

. Compared to the H_{mean} and channel bed grain size of other rivers, their aspect ratios are relatively small, but plot within, or above, the range of those in estuaries (Figure 14C and D). Lastly, their relatively low maximum θ_{lee} likely causes intermittent flow separation over crests , which is consistent with other sand-bed river LADs (i.e., grain size > 125 to ≤ 2000 μm; Figure 14E and F).

4.2 Downstream oriented dunes: zones 6-11

At low-river stage, the mixed tidal-fluvial regime (z6-11; ~ rkm 34-84) begins immediately upstream of the maximum extent of saltwater intrusion during high-river flows, where bed sediment increases in size to medium to coarse sands (Figure 3B). The first consequence of this shift in hydraulics and grain size is that all primary dunes become seaward-oriented (Figure 9) and larger in size, where average η, λ, and aspect ratios range from ~ 2-3m, ~ 60-110m, and ~ 30-55, respectively (Figure 12A). These values are equivalent to fully-fluvial dunes in the middle Columbia River reported by Smith & McLean (1977), whilst their aspect ratios remain comparable to those of the tidally-dominated regime (Figure 12B). In contrast to the tidally-dominated regime, the larger and coarser-grained dunes of the mixed tidal-fluvial regime (z6-11) are compound in form (Figure 9 F-F' to K-K', and Figure 15), and thus host superimposed bedforms on their crests, stoss-sides, and occasionally their lee-faces. However, similar to the tidally-dominated regime, they tend to be asymmetric (BSIs > 1.5) and not rounded (BRIs < 0.6; Figure 12C and D), but unlike their 2D downstream counterparts, they are mainly 2.5-3D (NDS ≥ 1.08; Figure 12E). According to their mean θ_{lee} (~ 11-18°), they are low-angle, but their mean θ_{lee} is marginally steeper than those of the tidally-dominated regime and other river and estuarine settings (Figure 12F). A greater mean θ_{lee} is a consequence of their slip-faces becoming: i) more distinct (i.e., transition in dune lee-side morphology to that of lee form 2A and 2B; Figure 11C) and ii) steeper (maximum θ_{lee} ≈ 18-29°) relative to those of the tidally-dominated regime (Figure 12F). Furthermore, from downstream to upstream, they display a slight increase in both maximum and mean θ_{lee} (z6 to z11; ~ 23-29° and 14-18°; Figure 12F). Overall, the mixed tidal-fluvial regime low-angle dunes closely resemble the Type-C dunes of Table 1, and those found in other sand-bed rivers (cf. Harbor, 1998; Nittrouer et al., 2008, 2011; Sambrook Smith et al., 2013; de Almeida et al., 2016; Knox & Latrubesse, 2016; Bradley & Venditti, 2017,; Galeazzi et al., 2018; Kostaschuk & Venditti, 2019; Cisneros et al., 2020).

In contrast to the tidally-dominated regime, scaling of mean η and λ to H_{mean} shows that mixed tidal-fluvial regime dunes plot along, or very close to, the best fit values for deep rivers where their η/H_{mean} and λ/H_{mean} ratios span 0.12-0.15 and 3.7-6.2 (Figure 14A and B). However, like their finer-grained downstream counterparts, they possess relatively small aspect ratios for a large river given H_{mean} and grain size, but their aspect ratios tend to be greater than, or equal, to that of estuarine dunes (Figure 14C and D). Comparison of maximum θ_{lee} *vs* H_{mean} and grain size shows that z6-9 dunes possess maximum θ_{lee} that are steeper than the mean of those from other sand-bed rivers, and thus likely possess intermittent flow separation over their crests (Figure 14E and F). However, these same relations for z10 and z11 dunes show that their maximum θ_{lee} are notably steeper than most river dunes (regardless of H_{mean}) with similar, or coarser, bed sediment (Figure 14E and F). It is therefore probable that some z10 and z11 dunes can maintain continuous flow separation over their crests, as has been shown in shallow river and laboratory flume high-angle dunes (Figure 14E and F).

5 DISCUSSION

5.1 Controls on primary dune morphology

Dune morphologic change is a function of how sediment is redistributed by suspended-load and bedload transport, but more specifically it is the duration of time a particular transport rate and mode (or q_{sus}/q_{bed} ratio) persists that determines their adjustment (Baas, 1999; Perillo et al., 2014c). Thus, dunes possess a turnover time (i.e., lag time), T_t, which is equal to A/q_s where A is the cross-sectional area of the dune and q_s is the sediment transport rate per unit width (Myrow et al., 2018). This means that for a constant sediment transport rate and grain size: i) larger dunes possess longer turnover, or developmental times, relative to smaller dunes, ii) below the transition to upper flow regime, higher transport rates tend to produce larger dunes over a given time period, and iii) unsteady repetitive flow cycles (i.e., tidal-cycles), and therefore unsteady sediment transport rate and q_{sus}/q_{bed} ratio cycles, produce faster morphologic changes in smaller dunes, whilst the constant changes in sediment transport rates and directionality throughout tidal-cycles limits their size.

Within this context, at low-river stage, the deepest (H/H_{max} ≥ 0.7) low-angle dunes of the tidally-dominated regime (z1-5) are interpreted to be controlled by the twice daily bidirectional tidal-current hysteresis loop that produces cyclic variations in the q_{sus}/q_{bed} ratio of fine to medium sand (125-250µm) due to acceleration and deceleration of ebb- and flood- tidal currents. This loop is characterised by a decrease in the magnitude of dune heights, wavelengths, aspect ratios, and lee-angles, when the q_{sus}/q_{bed} ratio is likely at its highest near peak tidal-velocities, but these characteristics recover as the q_{sus}/q_{bed} ratio, and therefore suspended-sediment concentrations, lowers during the phase shift between ebb- and flood- tidal flows (see figure 6, Hendershot et al., 2016). Thus, the spatio-temporal unsteadiness of flow, sediment transport rates, and the q_{sus}/q_{bed} ratio, restricts the traction-load renourishment of these dunes to a short time window during the waning stages of ebb- and flood- tidal flows when currents are decelerating and formerly intermittently suspended sand is reincorporated into bedload transport (e.g., Sherwood & Creager, 1990; Oost & Baas, 1994). This short renourishment window prevents their heights and wavelengths from reaching values that scale to mean flow depths (cf. Baas, 1999; Perillo et al., 2014c), impedes lee-angles from steepening towards the angle-of-repose, and promotes the creation of simple forms that are 2D (e.g., Baas, 1994, 1999; Dalrymple & Rhodes, 1995; Venditti, Church, & Bennett, 2005a; Rubin, 2012). Secondarily, the suppression of their heights and wavelengths are also likely augmented by internal flow stratification, and thus restriction of boundary layer thickness, caused by either saltwater intrusion or high near-bed suspended-sediment concentrations during peak tidal-current velocities (Jay & Smith, 1990; Dalrymple & Rhodes, 1995; Kay & Jay, 2003).

During low-river stages, however, the tidal hysteresis loop of cyclic q_{sus}/q_{bed} ratios is not spatially equal from z1 to z5, due to the upstream development of tidal-current asymmetry at depths where H/H_{max} > 0.5 (i.e., flood-tidal currents become stronger than ebb-tidal currents towards z5 as a product of saltwedge density-stratification and tidal-funnelling; Jay, 1984; Jay & Smith, 1990), and a slight coarsening of the channel bed. Thus, within z1 and z2, the ebb- and flood- tidal cycle q_{sus}/q_{bed} ratio hysteresis loops may be more symmetric and higher in magnitude, which leads to higher q_{bed} of fine to medium sand during the waning stages of tidal-phases and thus larger dune heights. In comparison, the ebb- and flood- current and q_{sus}/q_{bed} ratio loops will become increasing asymmetric (the flood-tidal phase becomes more dominant) from z3 to z5, although their overall magnitude will be weaker. This causes lower q_{bed} of fine to medium sand during the waning stages of tidal-phases, and consequently leads to smaller heights. Together, these factors are reasoned to cause the downstream to upstream trend in decreasing dune height from z1 to z5 and the switch from tidal-cycle reversing migration (z1 and z2) to upstream-only migration (z3-5; Figure 4A). Furthermore, it is thought that the overall greater rate of tidally-induced intermittently suspended sands (higher q_{sus}/q_{bed} ratios) throughout z1-5 is the principal mechanism driving these simple dunes to possess low (and nearly equivalent) maximum and mean lee-angles (i.e., lee form 1 morphology in Figure 11C), or gentle sloping slip-faces (e.g., Smith & McLean, 1977; Bridge & Best, 1988; Kostaschuk & Villard, 1996; Kostaschuk et al., 2009; Hendershot et al., 2016; Bradley & Venditti, 2017; Ma et al., 2017; Naqshband & Hoitink, 2020).

At low-river stage, the morphology of the larger and coarser-grained (250-750μm) low-angle dunes of the mixed tidal-fluvial regime (z6-11) is interpreted to be a function of a seasonally varying hysteresis loop involving a first stage dominated by unidirectional currents at high-river flows , and a second governed by bidirectional tidal-currents during subsequent low-river flows. Through time, the first stage is responsible for establishing and maintaining their larger size and downstream-orientations. This is because peak fluvial-discharges commonly last up to 3-4 months and are typically ~ 2.5 to 4 times larger than low-flow discharges (Table 2; Fig. 2A). These greater flows drive considerably greater suspended and bedload transport rates since the LCR discharge-sediment rating curve takes the form of a power-law (Naik & Jay, 2011). These higher transport rates (especially bedload transport rates) will enhance primarydune migration and growth rates (Baas, 1999; Perillo et al., 2014c), which in turn promotes higher order dimensionality (2.5-3D; e.g., Baas, 1994, 1999; Dalrymple & Rhodes, 1995; Venditti, Church, & Bennett, 2005a; Rubin, 2012). Enhanced sediment transport rates, along with the fact that their grain size is near or above the commonly observed threshold of ≥ 274μm for compound dune formation (Jackson 1976; Dalrymple 1984; Dalrymple & Rhodes, 1995; Bartholdy et al., 2002), will promote the development of superimposed bedforms on their stoss-sides and crests thus giving them their compound form. Additionally, higher q_{sus} conditions in conjunction with superimposed bedform migration and growth will tend to drive shallower leeside angles (e.g., Dalrymple & Rhodes, 1995; Carling et al., 2000; Sukhodolov et al., 2006; Reesink & Bridge, 2009; Naqshband & Hoitink, 2020). Furthermore, according to the LCR numerical simulations of Sandbach et al. (2018), such high fluvial-flows eliminate flood-tide induced current reversals throughout z6-11, thus leaving seaward-oriented currents as the singular flow constituent affecting dune morphodynamics. Together, the above factors provide a rationale as to why many of the geometric characteristics of z6-11 dunes closely resemble those of the middle Columbia River, or in general, those in fully-fluvial settings of similar grain size.

However, the 9-8 month long second stage (i.e., low-river period) of the hysteresis loop is interpreted to impact these relict compound dunes in three ways. First, bidirectional tidal-currents, and perhaps especially ebb-currents, likely promote the development of new superimposed bedforms, or sustain previously existing ones, thus helping to maintain their compound forms. Secondly, superimposed bedforms likely suppress flow separation over their host dunes and rework their crests, leesides, and stoss-sides, which helps to maintain their lower mean lee-angles (e.g., Allen & Collinson, 1974; Allen, 1978; Dalrymple & Rhodes, 1995; Carling et al., 2000; Sukhodolov et al., 2006; Reesink & Bridge, 2009; Best et al., 2020; Cisneros et al., 2020). Thirdly, the slight increase in their heights from downstream to upstream (z6 to z11; Figure 12A) is reasoned to be the product of decreasing reworking potential of superimposed bedforms as a function of decreasing tidal-current energy and the coarsening of bed alluvium (Figure 3B).

Several findings from this study diverge from previous research concerning dune morphodynamics. Firstly, the compound low-angle dunes of the mixed tidal-fluvial regime (i.e., dominated by downstream oriented currents) are asymmetric and not rounded, whilst a large percentage of the upstream most dunes (z10 and z11) likely display continuous flow separation (i.e., maximum lee-angles > 25°). Their asymmetry and lack of roundness thus makes them distinct from those in rivers but similar to those in estuaries (Dalrymple & Rhodes, 1995; Bradley & Venditti, 2017), whereas the continuous flow separation over the z10 and z11 dunes runs contrary to those in both river and estuarine settings. This suggests that: i) asymmetry and lack of roundness may be an indicator of tidal-forcing within the mixed fluvial-tidal regimes of fluvial-tidal transitions, and ii) some low-angle dunes may sustain continuous flow separation over their crests (Cisneros et al., 2020). However, further research is necessary to confirm whether these observations are universal or whether they are unique to the Lower Columbia River. Secondly, at low-river stage, there exists a downstream to upstream (i.e., tidally-dominated to mixed tidal-fluvial regime) trend in the steepening of dune maximum and mean lee-angles as well as a shift in leeside morphology (from lee form 1 to lee form 2A and B; Figure 11C).

This trend is reasoned to be the product of the downstream to upstream: i) coarsening of channel bed grain size thus a decrease in q_{sus}/q_{bed} ratios (i.e., bedload transport becomes more dominant in landward direction), and ii) reduction of superimposed bedform reworking of compound dunes throughout the mixed tidal-fluvial regime due to the decrease in tidal-current energy. This condition is likely enhanced in the Lower Columbia River since ≥ 50% of 'sand' grains are heavy minerals (Specific Gravity ≥ 2.8; Whetten, Kelley, & Hanson, 1969; Scheidegger & Phipps, 1976). Therefore, for a given bed shear stress, the LCR possesses a lower q_{sus}/q_{bed} ratio (i.e., more bedload transport dominated) than systems whose 'sand' consists of lighter grains of quartz and feldspar. Since higher q_{sus}/q_{bed} ratios reduce lee-angles (cf. Hendershot et al., 2016; Kostaschuk & Venditti, 2019; Naqshband & Hoitink, 2020), it is unsurprising that the steepest lee-angles (thus steepest slip-faces: lee form 2A or B) occur in the coarsest grained regions of the upper mixed tidal-fluvial regime (z10 and z11) where q_{sus}/q_{bed} ratios are presumably the lowest at low-river stages. Lower overall q_{sus}/q_{bed} ratios may also help to explain why the low-angle dunes of its fluvial-tidal transition tend to be more asymmetric and not rounded like the high-angle dunes of sand-bed shallow rivers and flumes. Although progress has been made in understanding the effects of grain density on sediment sorting (for example, Viparelli et al., 2015), more research is needed to fully understand the effects of heavy minerals on bedform morphology. Additionally, during low-river stages, it is unlikely that the upstream trend in decreasing q_{sus}/q_{bed} ratios is exclusive to the LCR. this suggests that all sand-bed fluvial-tidal transitions displaying downstream fining trends in grain size may possess a landward steepening in primary dune lee-angles, and thus also their cross-bed dip-angles.

5.3 Implications of trends in dune morphology

During low-river stages, ~ 86% of the deepest (H/H_{max} ≥ 0.7) channel bed of the Lower Columbia River fluvial-tidal transition is populated by larger-scale primary dunes that are compound and possess seaward orientations, and therefore are more ‘fluvial’ in their character, whilst ~ 90% is populated by them during high-river stages (see Sherwood & Creager, 1990; Figure 16A, B). Thus, only the most-seaward main channel of the fluvial-tidal transition (maximum of ~ 14% of total) displays smaller-scale, simple dunes with spatio-temporally fluctuating orientations indicative of a ‘fluvial-tidal to tidal’ signature (Figure 16A, B). Overall, this large-scale morphology pattern is more comparable to a tidally-influenced fluvio-deltaic environment, rather than a tidally-dominated estuary, which further supports the findings of Prokocki et al. (2015, 2020). Furthermore, the low-river stage downstream to upstream trend in increasing dune size through its fluvial-tidal transition also means that there is an accompanying upstream-directed increase in their form drag. Seasonal variations in the fluvial hydrograph , however, complicate this pattern by generating a localised ‘dynamic morphology reach’ extending from ~ rkm 12-35, where dunes here experience the greatest spatio-temporal variation in morphology, due to the down-river expansion and contraction of hydraulic regimes, associated changes in saltwater intrusion lengths, and channel bed grain size variations (Figure 16A, B).

For instance, at high-river stage, there is a seaward expansion of coarser grained ‘fluvial’ dunes from ~ rkm 35 to 27, and a 180° orientation reversal (from upstream to downstream directed) of the smaller-scale ‘fluvial-tidal’ dunes between ~ rkm 12-21. Whereas during the following low-river stage, tidal-energy (especially flood-tidal energy and saltwater intrusion) penetrates further upstream causing the ‘fluvial-tidal’ dunes between ~ rkm 12-21 to reverse their orientation by 180° (downstream to upstream) and to extend their development upstream to ~ rkm 32 via the (Figure 16A,B): i) cannibalisation of previously existing high-river stage ‘fluvial’ dunes between ~ rkm 27-32, and/or ii) fining of channel bed alluvium below the grain-size limit necessary to form the coarser ‘fluvial’ dunes. These morphology fluctuations will inevitably generate variations in the magnitude of form drag within the ‘dynamic morphology reach’, which should be parameterised for incorporation into hydraulic and morphodynamic simulations (Unsworth et al., 2020). Moreover, relative to those positioned farther downstream or upstream, the style, thickness, and orientation of primary dune deposits within the ‘dynamic morphology reach’ likely show the greatest variance , switching between upstream and downstream oriented simple to compound cross-bedding whose cross-sets will vary from thinner to thicker. These unique cross-bed sets may represent a distinct sedimentological signature to use when evaluating evidence of deltaic fluvial-tidal transitions or river dominated reaches of estuarine fluvial-tidal transitions in ancient fluvio-tidal environments that possess a similar downstream fining trend (coarse/medium to fine sands) as the LCR.

6 CONCLUSIONS

The deepest (H/H_{max} ≥ 0.7) channel reaches of the Lower Columbia River, and likely other deltaic fluvial-tidal transitions and river-dominated reaches of estuarine fluvial-tidal transitions with similar downstream fining trends (i.e., coarse/medium to fine sand) and hydraulic characteristics, are dominated by low-angle primary dunes, whose heights (and thus form drag) may abruptly increase landward of the high-river stage extent of salinity intrusion or within the mixed tidal-fluvial hydraulic regime. In these fluvial-tidal transitions, seaward-directed currents tend to control dune morphology, whilst the effects of bidirectional tidal-energy are subordinate. Thus, up to 90% of their longitudinal extents will consist of coarse to medium sand, 'fluvial' (large-scale, 2.5-3D, downstream-oriented, and compound) low-angle dunes whose morphology is very similar to those in sand-bed rivers. Only the most seaward reaches (< *c*. 20% of total longitudinal extent) will display 'fluvial-tidal or tidal' (smaller-scale, 2D, and simple or compound?) low-angle dunes composed of fine sand with reversing orientations caused by tidal-cycles and/or fluvial-discharge fluctuations. The low-angle character of the seaward-most 'fluvial-tidal or tidal' dunes are likely the product of higher suspended-sediment transport, q_{sus}, relative to bedload sediment transport, q_{bed}, (i.e., high q_{sus}/q_{bed} ratios) , whilst the upstream 'fluvial' dunes, where q_{sus}/q_{bed} ratios are lower, mainly owe their low-angle character to the reworking of their crests, stoss-sides, and lee-sides via the migration and development of superimposed bedforms and higher q_{sus} during high-discharge periods.

Through the fluvial-tidal transition of the LCR, primary low-angle dunes display a downstream (maximum lee-angle ~ 11-15°; mean lee-angle ~ 7-11°) to upstream (maximum lee-angle ~ 18-29°; mean lee-angle ~ 11-18°) steepening trend. This trend suggests that: i) their cross-bed dip-angles may also steepen, and ii) a greater proportion will transition from those with intermittent flow separation (max lee-angles ≥ 10 to ≤ 25°) to those displaying continuous flow separation (max lee-angles > 25°). this trend may exist at low-river stages in all upstream-coarsening (i.e., fine to coarse/medium sand) deltaic fluvial-tidal transitions and river-dominated portions of estuarine fluvial-tidal transitions and is potentially a signature of their fluvial-tidal transition. Lastly, Lower Columbia River discharge fluctuations create a 'dynamic morphology reach' within the deepest channel of its fluvial-tidal transition, which displays the greatest variability in primary dunes, and thus associated form drag. If the Lower Columbia River provides a case study that can be applied to other similar environments, the form drag variations in these reaches should be incorporated in hydraulic or morphodynamic simulations of their fluvial-tidal transitions. Also, such variance in dune morphology likely causes the style, thickness, and orientation of stacked cross-sets to possess the greatest deviations relative to both seaward and landward channel extents.

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