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The morphology of fluvial-tidal dunes: Lower Columbia River, OR/WA, USA

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 $\approx 0.3-0.8$ m; wavelengths $\approx 10-25$ m; mean lee-angles \approx 7-11°), and ii) medium to coarse sand 'fluvial' (downstream-oriented, compound, and 2.5-3D) low-angle dunes (heights \approx 1.5-3m; wavelengths \approx 60-110m; mean lee-angles \approx 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 35 the downstream-most reaches. River currents are thus the first-order control governing 36 dune morphology, with tidal-currentsexerting a second-order influence (especially in the 37 downstream part of the transition zone). Two mechanisms are reasoned to dictate their 38 low-angle character: (1) high-suspended sediment transport near peak tidal-currents 39 that lowers the leeside-angles of 'fluvial-tidal to tidal' dunes, and (2) superimposed 40 bedforms that erode the crests, leesides, and stoss-sides, of 'fluvial' dunes, which 41 results in the shallowing of leeside-angles. Fluctuations in river discharge creates a 42 'dynamic morphology reach', spanning river kilometres 12-40, which displays the 43 greatest variation in dune morphology. Similar channel reaches likely exist in fluvial-44 tidal transitions with similar physical characteristics as the LCR and may provide a 45 46 distinct signature of their fluvial-tidal transition.

47

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

50

51 **1 INTRODUCTION**

Bedforms are ubiquitous within subaqueous environments and are generated by 52 unidirectional, short to long period oscillatory, and combined-flows (currents with 53 unidirectional and oscillatory components), which deform a mobile bed through erosion 54 and deposition of sediment. For centuries, laboratory, field and theoretical investigations 55 56 have focused on bedform genesis, morphological equilibrium, and their depositional structures (e.g., Du Buat, 1786; Blasius, 1910; Kennedy, 1969; Harms et al., 1975; 57 Allen, 1983; Southard, 1991; Baas, 1994; Kleinhans, 2001; Venditti, Church, & Bennett, 58 2005a,b; Doucette & O'Donoghue, 2006; Reesink & Bridge, 2009; Perillo et al., 59

2014a,b,c; Bradley & Venditti, 2019a,b). One of the most common bedforms are dunes 60 (Best, 2005; Venditti, 2013), whose strata represent a fundamental building block of the 61 rock record (Ashley, 1990; Myrow and Southard, 1991; Myrow, Fischer, & Goodge, 62 2002; Bridge, 2003; Martinius & Van den Berg, 2011; Reynaud & Dalrymple, 2012). 63 Dunes possess heights of 0.075 to > 5m, wavelengths from 0.6 to > 100m, and can be 64 compound (possessing crests, stoss-sides, or leesides, populated with smaller-scale 65 superimposed bedforms) or simple (lacking superimposed bedforms) in form 66 (Dalrymple, Knight, & Lambiase, 1978; Dalrymple, 1984; Ashley, 1990). Primary 67 (largest) and secondary (superimposed) dune morphology is a function of their growth, 68 migration, and decay, as controlled by varying current magnitudes and orientations (cf. 69 Dalrymple, Knight, & Lambiase, 1978; Sherwood & Creager, 1990; Dalrymple and 70 71 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, 72 1999; Best, 2005; Hendershot et al., 2016; Bradley & Venditti, 2017; Ma et al., 2017; 73 Nagshband & Hoitink, 2020). Dunes are therefore spatially and temporally dynamic and 74 follow coupled flow and sediment transport hysteresis loops (Allen, 1974, 1976; Martin 75 & Jerolmack, 2013; Parsons & Best, 2013), which result in transient morphologic 76 properties (Dalrymple, Knight, & Lambiase, 1978; Bradley & Venditti, 2019a,b). These 77 properties include height (η), wavelength (λ), aspect ratio (AR), lee- and stoss- side 78 79 angles (θ_{lee} and θ_{stoss} , respectively), dimensionality (2 to 3D), roundness, symmetry, and scaling of n and λ to a characteristic flow depth, H, which typically is taken as local 80 (mean) depth, H_{mean} , or maximum depth, H_{max} . Thus, capturing how changes in dune 81 morphology induce variations in flow-fields via form drag (Smith & McLean, 1977; 82

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 89 has divided subaqueous dunes into two categories (Kostaschuk & Villard, 1996, 1999; 90 Venditti, 2013): high-angle dunes (HADs) and low-angle dunes (LADs). High-angle 91 dunes tend to be asymmetric and often are only slightly rounded, with long, gentle θ_{stass} 92 and short, steep θ_{lee} near, or at, the angle-of-repose (~ 25-30°)They have been 93 speculated to commonly be observed in bedload dominated laboratory flumes and 94 shallow (H < 2.5m) rivers (cf. Venditti & Bauer, 2005; Bradley & Venditti, 2017), but 95 have also been sporadically observed in the tidally-dominated Cobequid Bay, Bay of 96 Fundy, Canada (Dalrymple, 1984). In contrast, low-angle dunes possess mean θ_{lee} 97 below the angle-of-repose (typically $\leq 15^{\circ}$ in deep rivers where $H \geq 2.5$ m, Best & 98 Kostaschuk, 2002; Best, 2005; Kostaschuk et al., 2009; Bradley & Venditti, 2017; and ≤ 99 10° in estuarine settings, Dalrymple & Rhodes, 1995), and especially in rivers can be 100 more rounded and symmetric. Furthermore, experimental (Best & Kostaschuk, 2002), 101 numerical (Lefebvre, 2019; Lefebvre & Winter, 2016; Lefebvre et al., 2014a,b), and field 102 studies (Smith & McLean, 1977; Kostaschuk & Villard, 1996; Williams et al., 2003; 103 Holmes & Garcia, 2008; Kostaschuk et al., 2009; Bradley et al., 2013; Cisneros et al., 104

2020), suggest that flow separation over LAD leesides is intermittent ($\theta_{lee} \sim 10-24^\circ$) to 105 non-existent ($\theta_{lee} < 10^{\circ}$), whilst HADs possess continuous leeside flow separation. 106 Three modern environments have received the most attention regarding dune 107 morphology: i) relatively deep ($H \ge 2.5$ m; see Bradley & Venditti, 2017) rivers (e.g., 108 Harbor, 1998; Nittrouer et al., 2008, 2011; Sambrook Smith et al., 2013; de Almeida et 109 al., 2016; Knox & Latrubesse, 2016; Bradley & Venditti, 2017,; Galeazzi et al., 2018; 110 Kostaschuk & Venditti, 2019; Cisneros et al., 2020), ii) shallow marine combined 111 oscillatory wave-tidal settings (Carle & Hill, 2009; Ernstsen et al., 2010; Barnard, 112 Erikson, & Kvitek, 2011; Lefebvre, Ernstsen, & Winter, 2011; Fraccascia et al., 2016; 113 114 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; 115 Dalrymple, 1984; Aliotta & Perillo, 1987; Harris, 1988; Davis & Flemming, 1991; 116 Sherwood & Creager, 1990; Dalrymple & Rhodes, 1995; Gómez, Cuadrado, & Pierini, 117 118 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 119 HADs (e.g., Dalrymple & Rhodes, 1995; Hendershot et al., 2016; Kostaschuk & Venditti, 120 2019; Cisneros et al., 2020), and the re-evaluation of existing η and λ to H scaling 121 relations (Bradley & Venditti, 2017, 2019a; Cisneros et al., 2020). Ever since the 122 proposed original empirical scaling relations ($\eta = 0.17H$; $\lambda = 5H$) by Yalin (1964) and 123 Allen (1982), it has been widely adopted (although not universally accepted; see 124 Dalrymple & Rhodes, 1995; Bradley & Venditti, 2017) that dune η and λ scale to 125 boundary layer thickness (Allen, 1968; Ashley, 1990; Southard & Boguchwal, 1990a,b; 126 Best, 2005), which can be no greater than H_{max} . For instance, Bradley & Venditti (2017) 127

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 135 the bottom of downstream trough) and maximum θ_{lee} (slipface angle) measurements of 136 dunes from modern environments increases, it is apparent that silt- to gravel- bed rivers, 137 estuaries, and deltas are dominated by simple and compound low-angle dunes (e.g., 138 Dalrymple & Rhodes, 1995; Hendershot et al., 2016; Bradley & Venditti, 2017; Ma et al., 139 2017; Kostaschuk & Venditti, 2019; Cisneros et al., 2020). The physical processes 140 driving their formation, however, remain debated (Best et al., 2020; Best & Fielding, 141 2019), with mechanisms for compound low-angle dunes being: i) erosion of primary 142 dune crests and leesides via heightened localised bed shear stresses generated by 143 superimposed bedforms (Allen & Collinson, 1974; Allen, 1978; Reesink & Bridge, 2009), 144 and/or restriction of bedload supply to the primary dune crest by superimposed 145 bedforms, which starves them of sediment needed to maintain steep avalanching 146 slipfaces (Carling et al., 2000; Sukhodolov et al., 2006); ii) superimposed bedforms 147 may suppress flow separation over the crests of primary dunes (Dalrymple & Rhodes, 148 1995); and iii) development of compound dunes at obligue orientations relative to the 149 local flow direction can suppress flow separation due to the apparent leeside angle 150

being smaller than the true leeside angle (Sweet & Kocurek, 1990; Dalrymple & 151 Rhodes, 1995). Whereas the mechanisms for simple low-angle dunes include: i) shift of 152 sediment deposition from dune crests to dune leesides and/or lee-troughs caused by 153 higher rates of suspended-sediment transport, which reduces the sediment available 154 to maintain a steep avalanching slip-face (Smith & McLean, 1977; Bridge & Best, 1988; 155 Kostaschuk & Villard, 1996; Kostaschuk et al., 2009; Hendershot et al., 2016; Bradley & 156 Venditti, 2017; Ma et al., 2017; Nagshband & Hoitink, 2020); ii) downslope lee-face 157 currents resulting from intermittent to non-existent flow separation over dune leesides 158 (Kostaschuk & Venditti, 2019), and iii) liquefied grainflows produced by high excess 159 pore pressure causing failure of dune brinkpoint wedge deposits (Hendershot et al., 160 2016; Kostaschuk & Venditti, 2019), although this mechanism seems unlikely based on 161 162 abundant morphological data (Best et al., 2020). There is insufficient evidence to suggest that any one mechanism (for simple or compound low-angle dunes) 163 unequivocally explains their prevalence (see Kostaschuk & Venditti, 2019; Cisneros et 164 al., 2020) or that the proposed mechanisms for simple dunes do not also play some role 165 in compound dune formation. For example, in sand-bed (200-900µm) rivers, simple and 166 compound low-angle dunes coincide with estimated q_{sus}/q_{bed} ratios of > 2.3 167 (Kostaschuk & Venditti, 2019), whilst in the low Froude number (0.17 < Fr < 0.30) dune 168 experiments of Nagshband & Hoitink (2020) designed to mimic deep rivers, lee-angles 169 progressively lowered via the onset of increasingly higher rates of suspended-sediment 170 transport. Overall, this evidence suggests that the q_{sus}/q_{bed} ratio likely plays some role 171 in both simple and compound low-angle dune formation. 172

Four principal deficiencies persist in our understanding of the morphology of 173 fluvio-tidal dunes. Firstly, although studies tend to report n, λ , aspect ratios, and θ_{lee} , 174 they rarely quantify dimensionality, roundness, and symmetry, via standardised 175 parameters (cf. Harms, 1969; Arnott & Southard, 1990; Oost & Baas, 1994; Dumas, 176 Arnott, & Southard, 2005; Venditti, Church, & Bennett, 2005a,b; Sekiguchi & Yokokawa, 177 178 2008; Perillo et al., 2014a,b,c), but see Dalrymple, Knight, & Lambiase (1978), Dalrymple (1984), Dalrymple & Rhodes (1995), and Hendershot et al. (2016) for 179 exceptions. Secondly, many river and delta studies are spatially restricted to either 180 unsteady and non-uniform 'backwater' channel reaches, or to steady and uniform 181 'normal flow' reaches (but see Harbor, 1998, for an exception). Thus, few studies 182 183 capture any of the broader pattern changes driven by upstream to downstream 184 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 185 temporal windows that are fixed to a narrow range of river stages, which makes it 186 impossible to evaluate sediment transport regime and hysteresis effects caused by 187 discharge variations (cf. Bradley & Venditti, 2019a,b; Naqshband & Hoitink, 2020). 188 Thirdly, estuarine studies tend to be spatially confined to the lowermost backwater 189 190 reaches, which possess relatively low fluvial input and are thus dominated by bidirectional currents (Dalrymple & Rhodes, 1995; Gómez et al., 2010). Furthermore, 191 only a few studies have investigated temporal fluctuations over tidal-cycles (Dalrymple, 192 Knight, & Lambiase, 1978; Sherwood & Creager, 1990; Hendershot et al., 2016), or 193 captured upstream to downstream variations in dune morphology that are linked to 194 195 changes in grain size and spatio-temporal changes in tidal- and fluvial- energy.

196	To address several of these limitations, the present study analyses the primary
197	dunes within the deepest (\geq 9m depth from Mean Sea-Level, MSL, or local $H/H_{max} \geq$
198	0.7) portions of the main channel of the mesotidal Lower Columbia River (LCR),
199	OR/WA, USA, using a National Oceanic and Atmospheric Administration (NOAA)
200	multibeam echo sounder (MBES) dataset collected through ~ 90 river kilometres (rkm)
201	of its fluvial-tidal transition (FTT; Dalrymple & Choi, 2007, or fluvial-tidal zone; Van den
202	Berg, Boersma, & Van Gelder, 2007; Figure 1). Since FTTs are sensitive to sea-level
203	rise and are host to a large proportion of the world's population, spatial characterisation
204	of primary dune morphology can inform hydraulic and morphodynamic models used for
205	assessing flooding risk and ecological sustainability and remediation/dredging
206	strategies. Three questions are addressed herein:
207 208 209 210	1) How does the primary dune morphology in the deepest $(H/H_{max} \ge 0.7)$ channel vary through the LCR fluvial-tidal transition?
211	2) Do any variations in characteristics relate to transitions in
212	hydraulic and sediment transport processes, and/or bed
213	grain size?, and
214 215	3) How does their morphology differ between tidally-dominated
216	estuaries and unidirectional current-dominated rivers and flume experiments?
217	
218	2 STUDY SITE: LOWER COLUMBIA RIVER
219	The Lower Columbia River mean annual discharge, Q_{Wavg} , is ~ 7,000 m ³ s ⁻¹ at the Port

reach (Figures 1 and 2A; Naik & Jay, 2011). Thus, low-river stages possess discharges

220

Westward (Beaver) gauge station (~ rkm 85) near the upstream margin of the study

that are \leq 7,000 m³s⁻¹, whilst high-river stages are > 7,000 m³s⁻¹. Peak modern flows, 222 Q_{Wpeak} , are controlled by dam releases to between ~ 15,000-17,000m³s⁻¹ (Gelfenbaum, 223 1983; Naik & Jay, 2011; Simenstad et al., 2011), which is lower than pre-1900s 'natural' 224 spring freshets (≥ 18,000m³s⁻¹; Naik & Jay, 2011; Figure 2A). The LCR channel bed 225 slope, S_c , is ~ 1.15 x 10⁻⁵ (Hickson, 1912; Figure 2A), and is influenced by mesotidal 226 mixed diurnal and semidiurnal tides. At its mouth, the mean diurnal tidal prism, Q_{tide} , is 227 ~ 11.0 x 10^8 m³ (Walton & Adams, 1976) and the mean tidal range and highest 228 229 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; 230 Simenstad et al., 2011). Its hydrographic ratio, H_g , defined as $(Q_{tide}/Q_{Wavg}) \times 6hrs$, is 231 equal to 7 (where $H_q < 10$ reflects fluvial dominance; Peterson et al., 1984), where this 232 233 fluvial dominance is also supported by hydrodynamic modelling (Hamilton, 1990; Sandbach et al., 2018). Yet, in contrast to this finding, many oceanographic and 234 geologic studies describe the LCR as an estuary (e.g., Hughes and Rattray, 1980; 235 Gelfenbaum, 1983; Fox et al., 1984; Jay, 1984; Hamilton, 1990; Jay & Smith, 1990; Jay 236 et al., 1990; Sherwood & Creager, 1990; Simenstad et al., 2011; Peterson et al., 2013, 237 238 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 239 is that of an 'entrenched' fluvio-deltaic environment possessing a subaqueous deltaic 240 top-set extending to ~ rkm 50, thus supporting fluvial dominance, whilst providing a 241 physical explanation for its initial geomorphic designation as an estuary. 242 With respect to the LCR fluvial-tidal transition, Jay, Giese, & Sherwood (1990) 243

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 245 transition into three hydraulic zones (Figs 2B and 3A): 1) The tidally-dominated regime 246 (TDR; rkm 0-21), or lower delta, which experiences bidirectional tidal-currents, saltwater 247 intrusion, and development of a turbidity maximum (TM; Fox et al., 1984; Sherwood & 248 Creager, 1990); 2) The mixed tidal-fluvial regime (MTFR; ~ rkm 21-56; or upper 249 brackish water delta to lower freshwater tidal river reach; Hoitink & Jay, 2016), where 250 fluvial-currents and bidirectional tidal-currents compete for dominance; and 3) The 251 fluvially-dominated, tidally-influenced regime (FDTIR; ~ rkm 56-235; or mid to upper 252 freshwater tidal river reach; Hoitink & Jay, 2016), which is governed by downstream-253 oriented currents and terminates at the landward most point of tidally-forced variations 254 in water surface elevation, or tidal limit. However, like all FTTs (Dalrymple & Choi, 2007; 255 256 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-257 stages. For example, when transitioning from high- to low- river stage, the upstream 258 boundaries of the TDR, MTFR, and FDTIR, therefore expand to ~ rkm 35, 109, and 235 259 (Bonneville Dam; i.e., tidal limit), respectively, which causes (Figure 3A): i) brackish 260 water intrusion to ~ rkm 50 (Fox et al., 1984; Chawla et al., 2008), ii) fluvial current 261 reversals between rkm 50 to ~ 109 (Clark & Snyder, 1969), and iii) tidally-induced cyclic 262 water surface height variations beyond rkm 172 (Vancouver, WA; Kukulka & Jay, 2003). 263 Thus, during low-river stages, ~ rkm 0 to 35 (zones 1-5; hereafter z1-z5) is mainly 264 dominated by tidal flows (Figure 3B), whilst ~ rkm 35-90 (zones 6-11; hereafter z6-z11) 265 experiences a downstream to upstream increase in fluvial-energy as tidal influence 266 diminishes (Figure 3B). 267

268	The mean annual sediment-load, Q_s , of the Lower Columbia River where $Q_s =$
269	$Q_{sand} + Q_{wash}$ (Q_{sand} represents bedload + suspended-load particles (> 63 to <
270	2000µm), and Q_{wash} represents suspended-load particles \leq 62 µm), is sourced from
271	tributaries upstream of ~ rkm 172 (Jay, Giese, & Sherwood, 1990; Sherwood et al.,
272	1990; Naik & Jay, 2011), whilst negligible Q_s is derived from continental shelf sources
273	(Gelfenbaum et al., 1999; Templeton & Jay, 2013). Prior to the 1900s, its Q_s is
274	estimated to have been ~ 10 Mtyr ⁻¹ (Sherwood et al., 1990; Gelefenbaum et al., 1999;
275	Naik & Jay, 2011), but more recent estimates suggest that after 1970 (post-dam era) Q_s
276	has decreased by ~ 70% to ~ 3.2 Mtyr ⁻¹ (Naik & Jay, 2011). Grain size sampling of
277	deep ($H/H_{max} \ge 0.7$) channel-beds (Fox et al., 1984; Sherwood & Creager, 1990),
278	shows that its D_{50} is fine sand (> 125 to 250µm) from ~ rkm 0 to 32, but increases to
279	medium to coarse sands (> 250 to < 1000 μ m) upstream of rkm 32 (Figure 3B). This D_{50}
280	breakpoint tends to remain spatially fixed during both low- and high- river stages, but
281	during significant river floods, minimal coarsening of the channel-bed from fine to fine-
282	medium sand (> 125 to ~ 275 μ m) may occur between ~ rkm 16-32 (Sherwood &
283	Creager, 1990).

Previous analyses of primary dunes (\geq 9m depth; $H/H_{max} \geq 0.7$) 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 291 suspended-sediment transport context and during high-river stages possessed lee-292 faces that remained downstream oriented over tidal-cycles, but during low-river stages, 293 they maintained upstream orientations over tidal-cycles (Figure 4A, B). The Type-C 294 compound dunes were the coarsest grained and were confined to upstream of ~ rkm 30 295 during low-river flow but extended downstream to ~ rkm 24 at high-river flow (Figure 4A, 296 B). Throughout all tidal-cycles and river-stages, Type-C remained downstream-oriented 297 and formed within lower suspended-sediment transport conditions relative to Type-A 298 and B. These descriptions (Table 1), however, are limited in several ways: i) the 299 number (N) of dunes analysed is not provided, and observations are restricted to ~ rkm 300 0-35, ii) η , λ , and aspect ratios, are reported as general ranges without mean values, 301 302 and mix both low- and high- river stage observations, iii) dune symmetry, roundness, and dimensionality were not quantified via standardised methods, and iv) fundamental 303 properties such as lee-angles and scaling of η and λ to flow depth were not measured. 304 305

306 **3 METHODS OF ANALYSIS**

The low-river stage (< 7,000 m³s⁻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) 314 simultaneously with 2005-2010 dredging of the navigation channel (z1-7, and z9-11), or 315 ii) immediately after 2010 dredging (~ rkm 40-48; z8). Therefore, recently dredged 316 channel reaches outside or within the zones analysed were identified and avoided (see 317 Supporting Information: Type 1 and 2 dredged channel beds) to only evaluate naturally 318 occurring bedforms (Figures 5-8). Next, in each zone, 2D bed transects digitally sub-319 sampled at 1cm horizontal spacing were taken perpendicular to bedform crestlines. The 320 location of these transects are shown in Figures 5-8, whilst example partial sectional-321 profiles (A-A' to K-K' in Figures 5-8) are displayed in Figure 9. Note that all transects, 322 and thus bedforms evaluated, occur at depths \geq 9m (localised $H/H_{max} \geq$ 0.7; Figures 5-323 8) to remain consistent with previous Lower Columbia River bedform analyses. 324

325 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) 326 that identifies consecutive local minimum-maximum-minimum bed elevation points 327 along sectional-profiles (Figure 10A). The n values of all identified bed features were 328 then computed. At this stage, the computed n values of bed features includes both 329 primary and secondary (superimposed) bedforms (Figure 10A). However, to directly 330 compare and integrate the findings of this study with that of the primary dunes 331 examined by Sherwood & Creager (1990), all secondary bedforms were filtered from 332 the zone data. Therefore, a primary dune n cut-off value was established by finding the 333 maximum secondary bedform n within zone transects via hand measurements. For 334 each zone, these values thus represent the η breakpoint between primary dunes and 335 336 secondary bedforms (Figure 10B and Table 3), where the primary dunes of a given

337	zone must possess an η > the maximum η of secondary bedforms. However, since the
338	η of primary dunes, and thus also the maximum η of secondary bedforms, differs
339	between zones (Figures 9 and 10A), a unique primary dune η cut-off value was utilised
340	for each zone (Table 3). After applying the primary dune η cut-offs, the wavelengths, $\textbf{\textit{A}},$
341	of remaining primary dunes were calculated as the absolute lateral distance between
342	upstream and downstream minimum trough elevations (Figure 10B). The primary dunes
343	of each zone were then further characterised by quantifying the following geometric
344	properties.
345 346	Dune aspect ratio was calculated as:
347	$AR = \lambda/\eta \tag{1}$
348 349	whereas dimensionality (crestline sinuosity) was measured following the approach of
350	Venditti, Church, & Bennett (2005a), who defined a non-dimensional span index, NDS,
351	as:
352 353	$NDS = L_c / L_y $ (2)
354 355	where L_{c} represents the actual length of a dune crestline, and L_{y} is the straight-line
356	distance between the ends of this same crestline (Figure 11A). Using natural
357	breakpoints within the data, the following values were used to classify the bedform
358	sinuosity:
359 360 361 362 363	 NDS < 1.08 = 2D NDS ≥ 1.08 < 1.16 = 2.5D NDS ≥ 1.16 = 3D

364	Bedform symmetry and roundness indices (BSI and BRI, respectively) were computed
365	(Tanner, 1967; Perillo et al., 2014a) using:
366 367 368	$BSI = \lambda_{stoss} \lambda_{lee}^{-1} $ (3) $BRI = \lambda_{0.5_{stoss}} \lambda_{stoss}^{-1} $ (4)
369 370	where λ_{stoss} is stoss-side length, λ_{lee} is leeside length, and $\lambda_{0.5_{stoss}}$ represents the length
371	from dune crest to stoss-side at 0.5 η (Figure 11B), where the sectional form boundaries
372	were classified as (Perillo et al., 2014a):
373 374 375 376 377 378 379	 BSI > 1.5 = Asymmetric BSI ~ 1.3-1.5 = Quasi-asymmetric BSI < 1.5 = Symmetric BRI ≥ 0.6 = Rounded BRI < 0.6 = Not rounded
380	Next, maximum and mean $ heta_{lee}$ values were determined utilising the procedures and
381	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

389 Mean Sea-Level (MSL) water elevation.

Tests conducted on data from each zone show that primary dune η , λ , AR, BSI, 390 BRI, and maximum and mean θ_{lee} can be approximated via Gamma distributions (cf. 391 Paola & Borgman, 1991; van der Mark, Blom, & Hulscher, 2008; Cisneros et al., 2020), 392 whilst NDS indices cannot due to the lower number (N) of dunes analysed. Therefore, 393 the zone mean, median, standard deviation, and 25th and 75th guartile values reported 394 herein, for the metrics of η , λ , AR, BSI, BRI, and maximum and mean θ_{lee} , were 395 acquired by applying a Gamma probability density function (PDF) fit to raw 396 measurements, whilst mean zone NDS values and standard deviations were obtained 397 arithmetically from raw measurements. 398

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400 4 RESULTS: PRIMARY DUNES

Two categories of deep (local $H/H_{max} \ge 0.7$) primary dunes were recognised: i) 401 smaller-scale upstream oriented (~ rkm 1-27; z1-5), and ii) further upstream, larger-402 scale downstream oriented (~ rkm 34-85; z6-11). In total, ~ 1,400 dunes were 403 analysed, with the number of dunes examined per zone and their morphologic attributes 404 being reported in Table 4 and as shown graphically in Figure 12A-F. Herein, the 405 average morphology of these two categories is described within the context of past 406 research as well as: i) varying longitudinal position along the fluvial-tidal transition of 407 the LCR, ii) mean channel depths, and iii) channel bed grain size. 408

409 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 (125-250 μ m; 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 413 simple from compound dunes (especially those that are compound and possess 414 superimposed bedforms with $\eta \ge$ their host; Dalrymple & Rhodes, 1995). Detailed 415 stratigraphic evidence (i.e., presence of simple vs compound cross-bedding) is thus 416 needed to unequivocally confirm whether some of the z1-2 dunes are indeed compound 417 in form (Dalrymple, 1984). Without this information, the z1-5 dunes are considered 418 herein as simple in form following the previous observations of Sherwood & Creager 419 (1990; Table 1). Overall, z1-5 dunes are relatively small-scale with mean η and λ 420 ranging from ~ 0.3-0.8m and ~ 13-24m (Figure 13), and thus possess mean aspect 421 ratios of ~ 24-68 (Figure 12A, B). They are predominantly 2D (NDS \leq 1.08), asymmetric 422 (BSIs > 1.5), not rounded (BRIs < 0.6), and possess near equal mean and maximum 423 θ_{lee} (~ 7-11° and ~ 10-15°, respectively; Figure 12C-F). These dunes are therefore low-424 angle dunes possessing leesides with relatively mild dipping slip-faces (i.e., lee form 1; 425 Figure 11C), where their mean θ_{lee} is comparable to both sand-bed river and estuarine 426 LADs ($\leq 10^{\circ}$, Figure 12F). Overall, they resemble the Type-A and B dunes (Table 1) of 427 Sherwood & Creager (1990), where those of z1 and z2 are interpreted to be tidally-428 429 reversing (i.e., Type-A) and those of z3-5 reverse migration in response to the fluvialhydrograph (i.e., Type-B). 430

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 $\eta/H_{mean} = 0.025$ and $\lambda/H_{mean} = 1.0$ 435 . Compared to the H_{mean} and channel bed grain size of other rivers, their aspect ratios 436 are relatively small, but plot within, or above, the range of those in estuaries (Figure 14C 437 and D). Lastly, their relatively low maximum θ_{lee} likely causes intermittent flow 438 separation over crests, which is consistent with other sand-bed river LADs (i.e., grain 439 size > 125 to ≤ 2000 µm; Figure 14E and F).

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441 **4.2 Downstream oriented dunes: zones 6-11**

442 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 443 flows, where bed sediment increases in size to medium to coarse sands (Figure 3B). 444 The first consequence of this shift in hydraulics and grain size is that all primary dunes 445 become seaward-oriented (Figure 9) and larger in size, where average η , λ , and aspect 446 ratios range from ~ 2-3m, ~ 60-110m, and ~ 30-55, respectively (Figure 12A). These 447 448 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 449 tidally-dominated regime (Figure 12B). In contrast to the tidally-dominated regime, the 450 larger and coarser-grained dunes of the mixed tidal-fluvial regime (z6-11) are 451 compound in form (Figure 9 F-F' to K-K', and Figure 15), and thus host superimposed 452 bedforms on their crests, stoss-sides, and occasionally their lee-faces. However, similar 453 454 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, 455 they are mainly 2.5-3D (NDS \geq 1.08; Figure 12E). According to their mean θ_{lee} (~ 11-456 18°), they are low-angle, but their mean θ_{lee} is marginally steeper than those of the 457

tidally-dominated regime and other river and estuarine settings (Figure 12F). A greater 458 mean θ_{lee} is a consequence of their slip-faces becoming: i) more distinct (i.e., transition 459 in dune lee-side morphology to that of lee form 2A and 2B; Figure 11C) and ii) steeper 460 (maximum $\theta_{lee} \approx 18-29^{\circ}$) relative to those of the tidally-dominated regime (Figure 12F). 461 Furthermore, from downstream to upstream, they display a slight increase in both 462 maximum and mean θ_{lee} (z6 to z11; ~ 23-29° and 14-18°; Figure 12F). Overall, the 463 mixed tidal-fluvial regime low-angle dunes closely resemble the Type-C dunes of Table 464 1, and those found in other sand-bed rivers (cf. Harbor, 1998; Nittrouer et al., 2008, 465 2011; Sambrook Smith et al., 2013; de Almeida et al., 2016; Knox & Latrubesse, 2016; 466 467 Bradley & Venditti, 2017,; Galeazzi et al., 2018; Kostaschuk & Venditti, 2019; Cisneros et al., 2020). 468

In contrast to the tidally-dominated regime, scaling of mean η and λ to H_{mean} 469 shows that mixed tidal-fluvial regime dunes plot along, or very close to, the best fit 470 values for deep rivers where their η/H_{mean} and λ/H_{mean} ratios span 0.12-0.15 and 3.7-471 6.2 (Figure 14A and B). However, like their finer-grained downstream counterparts, they 472 possess relatively small aspect ratios for a large river given H_{mean} and grain size, but 473 their aspect ratios tend to be greater than, or equal, to that of estuarine dunes (Figure 474 475 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-476 bed rivers, and thus likely possess intermittent flow separation over their crests (Figure 477 14E and F). However, these same relations for z10 and z11 dunes show that their 478 maximum θ_{lee} are notably steeper than most river dunes (regardless of H_{mean}) with 479 similar, or coarser, bed sediment (Figure 14E and F). It is therefore probable that some 480

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).

484 **5 DISCUSSION**

485 **5.1 Controls on primary dune morphology**

Dune morphologic change is a function of how sediment is redistributed by suspended-486 load and bedload transport, but more specifically it is the duration of time a particular 487 transport rate and mode (or q_{sus}/q_{bed} ratio) persists that determines their adjustment 488 (Baas, 1999; Perillo et al., 2014c). Thus, dunes possess a turnover time (i.e., lag time), 489 T_t , which is equal to A/q_s where A is the cross-sectional area of the dune and q_s is the 490 sediment transport rate per unit width (Myrow et al., 2018). This means that for a 491 constant sediment transport rate and grain size: i) larger dunes possess longer 492 493 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 494 period, and iii) unsteady repetitive flow cycles (i.e., tidal-cycles), and therefore unsteady 495 sediment transport rate and q_{sus}/q_{bed} ratio cycles, produce faster morphologic changes 496 in smaller dunes, whilst the constant changes in sediment transport rates and 497 directionality throughout tidal-cycles limits their size. 498

Within this context, at low-river stage, the deepest $(H/H_{max} \ge 0.7)$ 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 504 q_{sus}/q_{bed} ratio is likely at its highest near peak tidal-velocities, but these characteristics 505 recover as the q_{sus}/q_{bed} ratio, and therefore suspended-sediment concentrations, 506 lowers during the phase shift between ebb- and flood- tidal flows (see figure 6, 507 Hendershot et al., 2016). Thus, the spatio-temporal unsteadiness of flow, sediment 508 transport rates, and the q_{sus}/q_{bed} ratio, restricts the traction-load renourishment of these 509 dunes to a short time window during the waning stages of ebb- and flood- tidal flows 510 when currents are decelerating and formerly intermittently suspended sand is 511 reincorporated into bedload transport (e.g., Sherwood & Creager, 1990; Oost & Baas, 512 513 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), 514 impedes lee-angles from steepening towards the angle-of-repose, and promotes the 515 creation of simple forms that are 2D (e.g., Baas, 1994, 1999; Dalrymple & Rhodes, 516 517 1995; Venditti, Church, & Bennett, 2005a; Rubin, 2012). Secondarily, the suppression of their heights and wavelengths are also likely augmented by internal flow stratification, 518 and thus restriction of boundary layer thickness, caused by either saltwater intrusion or 519 high near-bed suspended-sediment concentrations during peak tidal-current velocities 520 (Jay & Smith, 1990; Dalrymple & Rhodes, 1995; Kay & Jay, 2003). 521 During low-river stages, however, the tidal hysteresis loop of cyclic q_{sus}/q_{bed} 522

ratios is not spatially equal from z1 to z5, due to the upstream development of tidalcurrent 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 densitystratification and tidal-funnelling; Jay, 1984; Jay & Smith, 1990), and a slight coarsening

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of the channel bed. Thus, within z1 and z2, the ebb- and flood- tidal cycle q_{sus}/q_{bed} ratio 527 hysteresis loops may be more symmetric and higher in magnitude, which leads to 528 higher q_{bed} of fine to medium sand during the waning stages of tidal-phases and thus 529 larger dune heights. In comparison, the ebb- and flood- current and q_{sus}/q_{hed} ratio 530 loops will become increasing asymmetric (the flood-tidal phase becomes more 531 dominant) from z3 to z5, although their overall magnitude will be weaker. This causes 532 lower q_{bed} of fine to medium sand during the waning stages of tidal-phases, and 533 consequently leads to smaller heights. Together, these factors are reasoned to cause 534 the downstream to upstream trend in decreasing dune height from z1 to z5 and the 535 switch from tidal-cycle reversing migration (z1 and z2) to upstream-only migration (z3-536 5; Figure 4A). Furthermore, it is thought that the overall greater rate of tidally-induced 537 intermittently suspended sands (higher q_{sus}/q_{hed} ratios) throughout z1-5 is the principal 538 mechanism driving these simple dunes to possess low (and nearly equivalent) 539 maximum and mean lee-angles (i.e., lee form 1 morphology in Figure 11C), or gentle 540 sloping slip-faces (e.g., Smith & McLean, 1977; Bridge & Best, 1988; Kostaschuk & 541 Villard, 1996; Kostaschuk et al., 2009; Hendershot et al., 2016; Bradley & Venditti, 542 2017; Ma et al., 2017; Nagshband & Hoitink, 2020). 543

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 tidalcurrents 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 ~ 550 2.5 to 4 times larger than low-flow discharges (Table 2; Fig. 2A). These greater flows 551 drive considerably greater suspended and bedload transport rates since the LCR 552 discharge-sediment rating curve takes the form of a power-law (Naik & Jay, 2011). 553 These higher transport rates (especially bedload transport rates) will enhance 554 primarydune migration and growth rates (Baas, 1999; Perillo et al., 2014c), which in turn 555 promotes higher order dimensionality (2.5-3D; e.g., Baas, 1994, 1999; Dalrymple & 556 Rhodes, 1995; Venditti, Church, & Bennett, 2005a; Rubin, 2012). Enhanced sediment 557 transport rates, along with the fact that their grain size is near or above the commonly 558 observed threshold of \geq 274µm for compound dune formation (Jackson 1976; Dalrymple 559 1984; Dalrymple & Rhodes, 1995; Bartholdy et al., 2002), will promote the development 560 561 of superimposed bedforms on their stoss-sides and crests thus giving them their compound form. Additionally, higher q_{sus} conditions in conjunction with superimposed 562 bedform migration and growth will tend to drive shallower leeside angles (e.g., 563 Dalrymple & Rhodes, 1995; Carling et al., 2000; Sukhodolov et al., 2006; Reesink & 564 Bridge, 2009; Nagshband & Hoitink, 2020). Furthermore, according to the LCR 565 numerical simulations of Sandbach et al. (2018), such high fluvial-flows eliminate flood-566 tide induced current reversals throughout z6-11, thus leaving seaward-oriented currents 567 as the singular flow constituent affecting dune morphodynamics. Together, the above 568 factors provide a rationale as to why many of the geometric characteristics of z6-11 569 dunes closely resemble those of the middle Columbia River, or in general, those in fully-570 fluvial settings of similar grain size. 571

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However, the 9-8 month long second stage (i.e., low-river period) of the 572 hysteresis loop is interpreted to impact these relict compound dunes in three ways. 573 First, bidirectional tidal-currents, and perhaps especially ebb-currents, likely promote the 574 development of new superimposed bedforms, or sustain previously existing ones, thus 575 helping to maintain their compound forms. Secondly, superimposed bedforms likely 576 suppress flow separation over their host dunes and rework their crests, leesides, and 577 stoss-sides, which helps to maintain their lower mean lee-angles (e.g., Allen & 578 Collinson, 1974; Allen, 1978; Dalrymple & Rhodes, 1995; Carling et al., 2000; 579 Sukhodolov et al., 2006; Reesink & Bridge, 2009; Best et al., 2020; Cisneros et al., 580 2020). Thirdly, the slight increase in their heights from downstream to upstream (z6 to 581 z11; Figure 12A) is reasoned to be the product of decreasing reworking potential of 582 583 superimposed bedforms as a function of decreasing tidal-current energy and the coarsening of bed alluvium (Figure 3B). 584

Several findings from this study diverge from previous research concerning dune 585 morphodynamics. Firstly, the compound low-angle dunes of the mixed tidal-fluvial 586 regime (i.e., dominated by downstream oriented currents) are asymmetric and not 587 rounded, whilst a large percentage of the upstream most dunes (z10 and z11) likely 588 589 display continuous flow separation (i.e., maximum lee-angles $> 25^{\circ}$). Their asymmetry 590 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 591 continuous flow separation over the z10 and z11 dunes runs contrary to those in both 592 river and estuarine settings. This suggests that: i) asymmetry and lack of roundness 593 may be an indicator of tidal-forcing within the mixed fluvial-tidal regimes of fluvial-tidal 594

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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) 602 coarsening of channel bed grain size thus a decrease in q_{sus}/q_{bed} ratios (i.e., bedload 603 transport becomes more dominant in landward direction), and ii) reduction of 604 605 superimposed bedform reworking of compound dunes throughout the mixed tidal-fluvial 606 regime due to the decrease in tidal-current energy. This condition is likely enhanced in the Lower Columbia River since \geq 50% of 'sand' grains are heavy minerals (Specific 607 Gravity \geq 2.8; Whetten, Kelley, & Hanson, 1969; Scheidegger & Phipps, 1976). 608 Therefore, for a given bed shear stress, the LCR possesses a lower q_{sus}/q_{bed} ratio (i.e., 609 610 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 611 et al., 2016; Kostaschuk & Venditti, 2019; Nagshband & Hoitink, 2020), it is unsurprising 612 that the steepest lee-angles (thus steepest slip-faces: lee form 2A or B) occur in the 613 coarsest grained regions of the upper mixed tidal-fluvial regime (z10 and z11) where 614 q_{sus}/q_{bed} ratios are presumably the lowest at low-river stages. Lower overall q_{sus}/q_{bed} 615 ratios may also help to explain why the low-angle dunes of its fluvial-tidal transition 616 tend to be more asymmetric and not rounded like the high-angle dunes of sand-bed 617

shallow rivers and flumes. Although progress has been made in understanding the 618 effects of grain density on sediment sorting (for example, Viparelli et al., 2015), more 619 research is needed to fully understand the effects of heavy minerals on bedform 620 morphology. Additionally, during low-river stages, it is unlikely that the upstream trend 621 in decreasing q_{sus}/q_{bed} ratios is exclusive to the LCR. this suggests that all sand-bed 622 623 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-624 angles. 625

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5.3 Implications of trends in dune morphology

During low-river stages, ~ 86% of the deepest $(H/H_{max} \ge 0.7)$ channel bed of the 629 Lower Columbia River fluvial-tidal transition is populated by larger-scale primary dunes 630 631 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 632 Sherwood & Creager, 1990; Figure 16A, B). Thus, only the most-seaward main 633 channel of the fluvial-tidal transition (maximum of $\sim 14\%$ of total) displays smaller-scale. 634 simple dunes with spatio-temporally fluctuating orientations indicative of a 'fluvial-tidal 635 to tidal' signature (Figure 16A, B). Overall, this large-scale morphology pattern is more 636 comparable to a tidally-influenced fluvio-deltaic environment, rather than a tidally-637 dominated estuary, which further supports the findings of Prokocki et al. (2015, 2020). 638 Furthermore, the low-river stage downstream to upstream trend in increasing dune size 639 through its fluvial-tidal transition also means that there is an accompanying upstream-640

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 647 grained 'fluvial' dunes from ~ rkm 35 to 27, and a 180° orientation reversal (from 648 upstream to downstream directed) of the smaller-scale 'fluvial-tidal' dunes between ~ 649 rkm 12-21. Whereas during the following low-river stage, tidal-energy (especially flood-650 tidal energy and saltwater intrusion) penetrates further upstream causing the 'fluvial-651 652 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 653 16A,B): i) cannibalisation of previously existing high-river stage 'fluvial' dunes between ~ 654 rkm 27-32, and/or ii) fining of channel bed alluvium below the grain-size limit necessary 655 to form the coarser 'fluvial' dunes. These morphology fluctuations will inevitably 656 generate variations in the magnitude of form drag within the 'dynamic morphology' 657 658 reach', which should be parameterised for incorporation into hydraulic and morphodynamic simulations (Unsworth et al., 2020). Moreover, relative to those 659 positioned farther downstream or upstream, the style, thickness, and orientation of 660 primary dune deposits within the 'dynamic morphology reach' likely show the greatest 661 variance, switching between upstream and downstream oriented simple to compound 662 cross-bedding whose cross-sets will vary from thinner to thicker. These unique cross-663

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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.

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669 6 CONCLUSIONS

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The deepest $(H/H_{max} \ge 0.7)$ channel reaches of the Lower Columbia River, and 671 likely other deltaic fluvial-tidal transitions and river-dominated reaches of estuarine 672 fluvial-tidal transitions with similar downstream fining trends (i.e., coarse/medium to fine 673 sand) and hydraulic characteristics, are dominated by low-angle primary dunes, whose 674 heights (and thus form drag) may abruptly increase landward of the high-river stage 675 extent of salinity intrusion or within the mixed tidal-fluvial hydraulic regime. In these 676 fluvial-tidal transitions, seaward-directed currents tend to control dune morphology, 677 whilst the effects of bidirectional tidal-energy are subordinate. Thus, up to 90% of their 678 longitudinal extents will consist of coarse to medium sand, 'fluvial' (large-scale, 2.5-3D, 679 downstream-oriented, and compound) low-angle dunes whose morphology is very 680 similar to those in sand-bed rivers. Only the most seaward reaches (< c. 20% of total 681 longitudinal extent) will display 'fluvial-tidal or tidal' (smaller-scale, 2D, and simple or 682 compound?) low-angle dunes composed of fine sand with reversing orientations 683 caused by tidal-cycles and/or fluvial-discharge fluctuations. The low-angle character of 684 the seaward-most 'fluvial-tidal or tidal' dunes are likely the product of higher 685 suspended-sediment transport, q_{sus} , relative to bedload sediment transport, q_{bed} , (i.e., 686

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 691 downstream (maximum lee-angle ~ 11-15°; mean lee-angle ~ 7-11°) to upstream 692 (maximum lee-angle ~ 18-29°; mean lee-angle ~ 11-18°) steepening trend. This trend 693 suggests that: i) their cross-bed dip-angles may also steepen, and ii) a greater 694 proportion will transition from those with intermittent flow separation (max lee-angles \geq 695 10 to $\leq 25^{\circ}$) to those displaying continuous flow separation (max lee-angles > 25°). this 696 697 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 698 estuarine fluvial-tidal transitions and is potentially a signature of their fluvial-tidal 699 transition. Lastly, Lower Columbia River discharge fluctuations create a 'dynamic 700 morphology reach' within the deepest channel of its fluvial-tidal transition, which 701 702 displays the greatest variability in primary dunes, and thus associated form drag. If the 703 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 704 hydraulic or morphodynamic simulations of their fluvial-tidal transitions. Also, such 705 variance in dune morphology likely causes the style, thickness, and orientation of 706 stacked cross-sets to possess the greatest deviations relative to both seaward and 707 landward channel extents. 708

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