On the origin of crevasse-splay amalgamation in the Huesca fluvial fan (Ebro Basin, Spain)
Implications for connectivity in low net-to-gross fluvial deposits

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1 **On the origin of crevasse-splay amalgamation in the Huesca fluvial fan**
2 **(Ebro Basin, Spain): implications for connectivity in low net-to-gross fluvial**
3 **deposits**

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12 **Abstract**

13 Floodplain deposits are abundant in low-gradient dryland river systems, but their contribution to

14 connected reservoir volumes has not yet been fully acknowledged due to their poor detectability

15 with typical wireline log suites and relatively-lower reservoir quality. This study presents an analysis

16 of stacked crevasse splays in the distal part of the Miocene Huesca fluvial fan (Ebro Basin, Spain).

17 Vertical stacking of crevasse splays implies local aggradation of the active channel belt. Lateral

18 amalgamation of crevasse splays created an elevated rim around their feeder channel, raising its

19 bankfull height. Subsequent crevasse splays were deposited on top of their predecessors, creating

20 sand-on-sand contact through incision and further raising the active channel belt. This process of

21 channel-belt super-elevation repeated until an upstream avulsion occurred. Amalgamated crevasse

22 splays constitute connected reservoir volumes up to $\sim10^7$ m^3. Despite their lower reservoir quality,

23 they effectively connect channel deposits in low net-to-gross fluvial stratigraphy, and hence, their

24 contribution to producible volumes should be considered. Unswept intervals of amalgamated

25 crevasse splays may constitute a secondary source of natural gas. Their interval thickness can serve

26   as a proxy for feeder-channel dimensions, which can in turn be used to estimate the degree of

27   stratigraphic connectivity.

28   **Keywords**

29   Low-gradient dryland river systems; low net-to-gross fluvial stratigraphy; Huesca fluvial fan; crevasse

30   splays; stacking mechanism; connectivity

1. Introduction

Floodplain deposits in low net-to-gross fluvial stratigraphy contain fine-grained reservoir bodies (e.g., Donselaar et al., 2011; McKie, 2011a; Ford and Pyles, 2014). These sub-seismic-resolution deposits are difficult to distinguish on well logs (e.g., Passey et al., 2004; Bridge, 2006) and are thought to represent a relatively small proportion of overall sandstone volumes with a reservoir quality that is generally lower than that of coarser-grained fluvial facies (e.g., Pranter et al., 2008). As a result, intervals of floodplain deposits have been discarded as 'waste zones' in conventional reservoir development (Donselaar et al., 2011) and published research on their accumulation and reservoir architecture is limited (Bridge, 2006). Smith et al. (1989), van Gelder et al. (1994), Tooth (2005), Donselaar et al. (2013), and Li et al. (2014) have studied the deposition of crevasse splays in active river systems through field work and time-lapse satellite data analysis. Fisher et al. (2007), Hampton and Horton (2007), Jones and Hajek (2007), Nichols and Fisher (2007), and Gulliford et al. (2014) have provided qualitative characterisations of ancient fluvial floodplain deposits exposed in outcrop and proposed and/or applied conceptual models for their stratigraphic evolution. Jordan and Pryor (1992), Pranter et al. (2008, 2009), McKie (2011a), and Ford and Pyles (2014) have presented quantitative reservoir-architecture studies of heterogeneous fluvial intervals based on outcrop and subsurface data. These authors have acknowledged the contribution of crevasse-splay sheet sands to connected reservoir volumes, but focussed on channel and bar deposits which constitute the highest-quality reservoir sandstones. A dedicated study of overbank splay geometries has been conducted by Mjøs et al. (1993), who established geometric ratios and rudimentary volume estimates based on outcrop analogues. A more in-depth examination is needed to better assess the (secondary) reservoir potential of crevasse splays and their impact on reservoir connectivity.

Floodplain deposits are abundant in the distal part of low-gradient dryland river systems fringing endorheic basins, such as the Huesca fluvial fan, Ebro Basin, Spain (e.g., Nichols and Fisher, 2007). The distal part of the fluvial system is characterised by one single active channel with a downstream-

56 decreasing bankfull capacity (Tooth, 2000, 2005; Nichols and Fisher, 2007; Weissmann et al., 2010;

57 Donselaar et al., 2013; Li and Bristow, 2015). This has been attributed to: (1) a gradient-induced loss

58 in flow energy, and (2) a transmission loss due to high percolation and evapo-transpiration rates. The

59 downstream decrease in cross-sectional area makes the channel prone to extensive overbank

60 deposition during short episodes of peak discharge (Donselaar et al., 2013). Similar depositional and

61 climatological conditions characterised Permo-Triassic basins throughout the Central and North

62 Atlantic margins (Williams and McKie, 2009; McKie, 2011b), including now gas-prolific plays in

63 Northwest Europe (e.g., Geluk, 2007a, b; Donselaar et al., 2011).

64 In this study, an analysis is presented of thin-bedded floodplain deposits in the distal part of the

65 Huesca fluvial fan. Depositional mechanisms explaining the occurrence of intervals of stacked

66 crevasse splays will be proposed. Understanding these processes and their preserved sedimentary

67 architecture improves connectivity estimations of subsurface floodplain reservoirs and aids the

68 interpretation of low net-to-gross fluvial stratigraphy.

69 **2. Geological setting**

70 The Huesca fluvial fan is located on the northern fringe of the Ebro Foreland Basin in northeast Spain

71 (Fig. 1a). The basin formed during the Cenozoic in the Pyrenean Phase of the Alpine Orogeny. It is

72 bounded by the Sierras Marginales thrust front to the north and the Iberian and Catalan coastal

73 ranges to its SW and SE, respectively. From the late Oligocene to the late Miocene, the centre of the

74 basin was occupied by a lake undergoing cycles of water-level fluctuations, which caused km-scale

75 migrations of the shoreline over a low-gradient coastal plain. Alternations of lacustrine clastic

76 sequences with carbonates and evaporites led Arenas and Pardo (1999) to associate high lake levels

77 with a relatively humid paleoclimate, whereas low lake levels were linked to a more arid playa-lake

78 environment.

79 The Huesca fluvial system derived its sediment from the high Pyrenees as well as the uplifted South

80 Pyrenean Foreland Basin to its north. Sediments of the Huesca fluvial system entered the Ebro basin

81 through a 15-20 km wide gap formed by a fractured zone in the Sierras Marginales thrust sheet,

82 which acted as its line source (Donselaar and Schmidt, 2005). The fan has a ~60 km radius and shows

83 a change in fluvial style from amalgamated braided streams in its proximal reaches to meandering

84 channels and eventually unconfined terminal lobes towards its distal fringe (Nichols and Fisher, 2007;

85 Fisher et al., 2007) (Fig. 1a).

86 Low net-to-gross floodplain deposits are abundant in the distal part of the fluvial system (Nichols and

87 Fisher, 2007). Here, channel deposits comprise <10 % of the overall stratigraphy (Hirst, 1991). Their

88 low-sinuous ribbon geometry (width-to-thickness ratio <15) has been attributed to a limited

89 longevity, as channels avulsed prior to any substantial lateral migration (e.g., Friend et al., 1979;

90 Hirst, 1991; Nichols and Fisher, 2007). Thin-bedded sandstone sheets are common and frequently

91 extend from the top of channel deposits as 'wings', continuing into the channel-fill sandstone (Friend

92 et al., 1986; Hirst, 1991; Nichols and Fisher, 2007; Fisher et al., 2007). These laterally-extensive

93 sandstone bodies have been interpreted as the result of unconfined sheet flow (Friend et al., 1986),

94 i.e., overbank levees and crevasse splays (Hirst, 1991; Fisher et al., 2007).

95 **3. Methodology**

96 The study area is located to the northeast of Huesca and comprises two outcrop localities spaced

97 approximately 1.6 km apart (Fig. 1b). Natural cliff faces at 1.3 km to the NNW of the *Castillo de*

98 *Montearagón* fortress (Fig. 1d) cover an approximately 35 m thick interval (Fig. 2a) over an area of

99 0.2 km^2. The western cut slope of the *Presa de Montearagón* reservoir dam (Fig. 1c) exposes a 52 m

100 thick succession (Fig. 2b) over a length of 550 m.

101 The local structural dip was reconstructed in order to accurately correlate data between outcrop

102 localities. A well-developed paleosol at the base of the succession was chosen as a reference horizon,

based on the assumption that it developed on a near-horizontal floodplain during a sustained period

of inactivity (Kraus, 2002). Sixteen point locations were measured along this horizon with sub-

centimetre accuracy using a *Trimble R7* differential-GPS (dGPS) set (Parkinson and Enge, 1996). These

were used in a least-squares regression analysis to estimate the orientation of the planar surface,

yielding a structural dip of 0.96° with an azimuth of 227° towards the SW. This is consistent with

earlier studies (e.g., Hirst, 1991) and compass validations. The corresponding coefficient of

determination (R^2) is 0.99 with a vertical standard deviation (σ_z) of 0.14 m, supporting the

assumption that the reference paleosol is planar. The structural dip was used in conjunction with

dGPS anchor points to stratigraphically correlate the outcrop localities, revealing a common

stratigraphic interval of approximately 34 m (Fig. 2).

Sedimentological logs comprising bed thickness, nature of contacts, grain size and colour,

sedimentary structures, paleoflow measurements, and bioturbation were recorded at 1:50 scale. The

apparent bed thickness on sloping sections was restored to true thickness. Log correlation was aided

with digital photo panels, a laser range finder (0.1 m accuracy), and dip-corrected dGPS

measurements.

Samples were collected at 50-cm vertical increments from the top of selected intervals for grain-size

analysis using a *Helos KR Sympatico* laser particle sizer (Blott et al., 2004). The polymodal grain-size

distributions (GSDs) represent mixtures of overlapping components: i.e., bed-load (saltation) and

suspended-load sediments. The GSDs were therefore decomposed by curve fitting using Weibull

distribution functions (Sun et al., 2002; Weltje and Prins, 2007).

4. Facies description and interpretation

The sedimentary logs of both outcrop localities show similar lithologies within their common

stratigraphic interval, particularly for intervals of stacked siltstone and sandstone sheets (Fig. 2). The

126   relative location of the outcrops (Fig. 1b) is approximately parallel to the paleoflow direction

127   measured in these deposits (Fig. 3), suggesting that one is proximal to the other by ~1.6 km.

128   *4.1 Thin-bedded crevasse splays*

129   Ochre-to-grey thin-bedded siltstone and sandstone sheets are observed throughout the sections in

130   both localities (Fig. 2), commonly exceeding outcrop dimensions in lateral extent. Their thickness

131   ranges from 0.05 to 0.6 m with a sharp and locally-undulating base. The tops are diffuse, typically

132   heavily bioturbated, and blue-grey in colour (Fig. 4a). Individual beds fine upwards from fine-grained

133   sand to silt and frequently contain <1 cm mud pebbles at their base. Climbing ripples and horizontal

134   laminae are common, whereas cross bedding infrequently occurs where beds incise into underlying

135   deposits. Sedimentary structures and scour orientations indicate a paleocurrent direction towards

136   the N to NW (Fig. 3).

137   The facies described here are interpreted as crevasse-splay deposits. Individual crevasse splays

138   originated from a breach point in the previously-deposited levee of their feeder channel, forming

139   lobate or elongated sheet sands with surface areas up to several square kilometres (Mjøs et al.,

140   1993; Li and Bristow, 2015). Their sharp (and partly erosional) basal contact is indicative of a high

141   flow energy directly after crevassing, when sediment-laden water entered the floodplain under a

142   waxing flow regime. This is supported by the occurrence of rip-up mud clasts, which originated from

143   the desiccated floodplain surface and underwent short-distance bedload transport (Rust and Nanson,

144   1989). Low sinuous crevasse-splay channels focussed the outflow and may have incised into the

145   substrate (Ford and Pyles, 2014). Further downstream in the crevasse splay, they bifurcate

146   (Donselaar et al., 2013) and end in subtle terminal mouth bars (Bridge, 2006). Abundant climbing

147   ripples indicate that fine sand and silt were deposited under a waning flow regime. The diffuse top

148   suggests a gradual transition to deposition from suspension. This occurred when the floodplain was

149   inundated and suspended-load sediment started to form conformable drapes of floodplain fines.

150   *4.2 Floodplain fines*

151   Horizontally-laminated claystone and siltstone constitute 50 to 60% of the total stratigraphy in the

152   sedimentary logs (Fig. 2). Continuous successions are up to 5 m thick (Fig. 2) and often heavily

153   weathered or vegetated. The colour of these deposits ranges from ochre to beige (Fig.4c), displaying

154   abundant red mottling. Structureless horizons of claystone and fine siltstone with diffuse contacts

155   occur throughout the sections, varying in thickness (up to 0.5 m) and lateral extent. Their colour

156   ranges from pink to purple-red and they are frequently overlain by a blue-grey or white top.

157   Laminated claystone and siltstone are interpreted as floodplain fines. These sediments settled out of

158   suspension during floodplain inundation, forming a conformable drape on top of previous deposits.

159   They were subsequently subaerially exposed, causing widespread mottling. Extended periods of

160   pedogenesis formed paleosols, manifested as red to purple-red-coloured horizons. This required a

161   prolonged absence of sediment influx, signifying local floodplain inactivity (Kraus, 2002). The

162   paleosols in the study area match the descriptions of the Bolea (inceptisol) and Erla (entisol)

163   pedotypes, as defined by Hamer et al. (2007). Their blue-grey tops have been attributed to burial

164   gleization (Retallack, 1991).

165   *4.3 Ribbon channel sands*

166   Ribbon-shaped sandstone bodies range from 1.0 to 2.8 m in thickness, with the maximum probability

167   density at 2.0 m (standard deviation (σ) = 0.78 m; n = 6) (Fig. 5). They may be stacked up to 5 m in

168   height with clear reactivation surfaces (Fig. 4d). Their base is sharp and erosional, overlain by

169   medium-to-coarse-grained poorly-sorted sand and rip-up mud clasts. Trough cross bedding with set

170   heights of 20-40 cm occur throughout the sandstone body. At the top, grain size decreases to fine-to-

171   medium-grained sand with abundant bioturbation. The orientation of the low-sinuous ribbon axes

172   and sedimentary structures indicate a paleoflow direction towards the W to NW (Fig. 3), which is in

173   accordance with earlier studies (e.g., Hirst, 1991) and oblique to the paleoflow direction of the

8

174 crevasse splays (Fig. 3). These sandstone ribbons typically truncate intervals of stacked thin-bedded

175 crevasse splays (Fig. 4d), creating sand-on-sand contact at their erosional base.

176 Ribbon-shaped sandstone bodies are interpreted as fluvial channels whose low width-to-thickness

177 ratios (<10) and lack of accretionary surfaces indicate limited lateral migration (Friend et al., 1979;

178 Hirst, 1991; Nichols and Fisher, 2007). Their cross-bedded channel fill overlies an erosive base,

179 suggesting incision and subsequent aggradation of channel-lag deposits up to 2.8 m thick. Vertical

180 stacking of the channel ribbons into multi-storey sandstone bodies is caused by repeated

181 reoccupation of the remnant channel depression (Slingerland and Smith, 2004). Friend et al. (1986),

182 Hirst (1991), Nichols and Fisher (2007), and Fisher et al. (2007) have observed crevasse splays

183 extending from the top of ribbon channels as 'wings', forming a continuous sandstone body. This

184 could not be unambiguously verified in the outcrop localities of this study due to the poor exposure

185 of these interfaces.

186 **5. Crevasse-splay amalgamation**

187 *5.1 Lateral amalgamation*

188 Donselaar et al. (2013) and Li et al. (2014) have observed compensational stacking of crevasse splays

189 in the modern-day Río Colorado fluvial system (Altiplano Basin, Bolivia). The lobate geometry of

190 crevasse splays forms a dm-scale topography on the otherwise planar floodplain. Subsequent splays

191 are deposited in the subtle topographic lows between adjacent crevasse splays. The proximally

192 erosive base of the newly deposited splay truncates the finer-grained top of its previously deposited

193 neighbours, creating sand-on-sand contact. Further away from the feeder channel, the depth of

194 incision is less, preserving the finer-grained top of the underlying sheets in an onlapping geometry.

195 This process of lateral amalgamation creates large areas of interconnected sand sheets up to ~10^7 m^2

196 (Li and Bristow, 2015), elevating the floodplain proximal to the feeder channel. The mechanism is

197 thought to be generic , but its expression is difficult to unambiguously identify in the study area due

198 to the limited extent of outcrop exposure and lack of grain-size contrast at lateral amalgamation

199 surfaces.

200 *5.2 Vertical stacking*

201 Individual crevasse splays are frequently stacked (Fig. 4b) to a combined thickness of up to 2.35 m,

202 with the maximum probability density at 1.8 m ($\sigma = 0.37$ m; n = 36) (Fig. 5). Such intervals are

203 typically underlain by a red to purple-red paleosol with diffuse contacts, separated by several

204 centimetres of ochre to red laminated floodplain claystone and siltstone. The lowermost crevasse

205 splay has a sharp planar base, whereas subsequent splays show an increasingly-undulating base with

206 an undulation relief of <30 cm (Fig. 6). Proximal to the feeder channel, these bed bases cut tens of

207 centimetres into underlying crevasse splays, having removed their fine-grained top and creating

208 sand-on-sand contact. The grain-size contrast at these contacts is dependent on the depth of incision

209 and the interface is frequently marked by <1 cm mud clasts separated by a sand matrix. The

210 amalgamation surfaces exceed outcrop dimensions and are only locally interbedded with remnant

211 drapes of floodplain fines. The depth of incision decreases in the paleocurrent direction, where the

212 finer-grained top of the underlying sheet is preserved up to several hundreds of metres from its

213 distal fringe. This results in a downstream separation of individual beds, which finger out and lose

214 grain-size contrast with interbedded parallel-laminated floodplain fines (Fig. 7). Laser particle size

215 analysis of samples from these intervals reveals grain sizes ranging from 2 to 250 µm in diameter, i.e.,

216 clay to fine-grained sand (Wentworth, 1922). The grain-size distributions are decomposed into

217 component A, ranging from clay to very-fine silt and displaying no obvious trends, and component B,

218 ranging from medium silt to fine sand and showing a subtle coarsening-up trend within consecutive

219 sheets (Fig. 8).

220 *5.3 Depositional mechanisms*

221 Vertical stacking of crevasse splays implies local aggradation of the active channel belt, i.e., super-

222 elevation above the surrounding floodplain. Successive avulsions caused the active channel belt to

223   periodically shift and enter previously-inactive lower-lying parts of the floodplain in a process of

224   large-scale compensational stacking (Slingerland and Smith, 2004). Local accommodation

225   downstream of each avulsion point induced super-elevation of the active channel belt, allowing

226   crevasse splays to stack until a subsequent avulsion terminated the influx of sediment (e.g., Mohrig

227   et al., 2000; Dalman et al., 2015).

228   Prolonged periods of local floodplain inactivity and subaerial exposure allowed for extensive

229   pedogenesis, resulting in mature paleosols (Kraus, 2002) (Figs. 4c, 6, 9a). Depocentre shifts within the

230   fluvial system periodically reactivated the local floodplain, which was manifested by deposition of

231   floodplain fines (Figs. 4c, 6). Reactivation was occasionally followed by (partial) avulsion onto the

232   local floodplain when a channel reached a critical threshold for avulsion (e.g., Mohrig et al., 2000;

233   Slingerland and Smith, 2004; Hajek and Wolinsky, 2012). Prolific crevasse splays prograded onto the

234   lower-lying floodplain adjacent to the active fluvial ridge, receiving a gradually increasing share of the

235   total discharge. These preludes to avulsion have been termed associated non-coeval splays by Ford

236   and Pyles (2014). As their crevasse-splay channels stabilised and decreased in number, the increasing

237   flow volume was accommodated through headward incision and levee development (Smith et al.,

238   1989; Hajek and Wolinsky, 2012) (Fig. 9b).

239   Even before the gradual avulsion was fully completed, levee build-up caused the bankfull height of

240   the new channel to rise. This was then mirrored by aggradation of the newly-established channel

241   when flow volumes stabilised and the hydraulic capacity became more or less constant (Fig. 9c).

242   Weak points in the levees were breached during peak discharge events, after which their sediment

243   was reworked into crevasse splays (termed associated coeval splays by Ford and Pyles, 2014). The

244   initial crevasse splays were deposited onto a relatively flat floodplain surface, resulting in a planar

245   base (Figs. 6, 9c). Amalgamation with adjacent lobes (Li et al., 2014) created an elevated rim

246   extending up to several kilometres away from the active channel (Donselaar et al., 2013) (Fig. 9d).

247   Levees continued to build on top of the proximally-aggraded floodplain and were in turn reworked

248   into crevasse splays, which stacked on top of precursory splays and further raised the floodplain

249   proximal to the active channel. The subsequent rise in bankfull height was again mirrored by an

250   aggradation of the channel thalweg. The resulting increased overbank gradient caused successive

251   crevasse spays to longer retain their energy. This is manifested by deeper incision of crevasse-splay

252   channels close to their feeder channel (Fig. 6), an increase in grain size (Fig. 8), and sediment

253   transport further onto the floodplain, resembling a subaerial analogue of a prograding high-stand

254   delta (Fig. 9e). This iterative process of channel-belt aggradation and widening continued until an

255   upstream avulsion occurred.

256   The expression of stacked crevasse splays in this study (Figs. 6, 7) and the proposed depositional

257   mechanisms (Fig. 9) seem to closely match the inferences of Dalman et al. (2015). These authors

258   used a numerical model with subgrid parameterisation of channel evolution (crevasses, avulsions,

259   bifurcations) to simulate super-elevation of a fluvio-deltaic system onto a low-gradient shelf as a

260   result of progradation (regression), where accommodation is created through a lengthening of the

261   fluvial equilibrium profile and a reduction of its gradient. In fact, fluvial system aggradation (i.e.,

262   creation of accommodation) may occur throughout a supply-accommodation cycle (sequence),

263   except when relative sea level falls below a morphologically-conspicuous shelf break (e.g., Koss et al.,

264   1994). During this late stage of the falling stage systems tract (FSST), progradational units with a

265   fundamentally different geometry are created, characterized by the absence of avulsions and an

266   abundance of bifurcations (Karamitopoulos et al., 2014). Throughout the rest of a supply-

267   accommodation cycle, stacking of crevasse splays owing to channel-belt aggradation is the rule in

268   low-gradient fluvial systems (e.g., Hajek and Wolinsky, 2012; Dalman et al., 2015).

6. Implications for reservoir evaluation

6.1 Connectivity

Despite their large areal extent (up to ~10^6 m^2) , the limited thickness (~10^{-1} m) of individual crevasse splays only yields sediment volumes on the order of ~10^5 m^3. Lateral amalgamation connects individual splays into aerially-extensive sand sheets (Donselaar et al., 2013; Li et al., 2014) (Fig. 10a), increasing the aerial extent and, hence, connected sediment volumes by one order of magnitude, i.e., to ~10^6 m^3. Vertical stacking of crevasse splays creates sand-on-sand contact through local erosion of underlying beds (Fig. 10b), mainly proximal to their feeder channel. In this study, two intervals of stacked crevasse splays have been correlated between the outcrop localities (Fig. 2), well within the maximum areal extent of crevasse splays (Mjøs et al., 1993; Li and Bristow, 2015). Vertical stacking up to several metres thickness may increase the connected sediment volume of these intervals by another order of magnitude, i.e., to ~10^7 m^3.

Crevasse splays are connected to channel-lag and heterolithic channel-fill deposits, which typically constitute the lithologies with best reservoir quality (e.g., Fielding and Crane, 1987; Pranter et al., 2008). Channel-lag deposits in the crevasse-splay channels connect as 'wings' to their coeval feeder-channel fill (Friend et al., 1986; Hirst, 1991; Mjøs et al., 1993; Nichols and Fisher, 2007; Fisher et al., 2007) (Fig. 10c). Truncation of previously-deposited crevasse splays (i.e., non-associated or associated non-coeval splays; Ford and Pyles, 2014) by later channels may further increase connectivity (Mjøs et al., 1993) (Fig. 10d). This contradicts the notion that laterally-restricted fluvial reservoirs such as the channel ribbons in this study are isolated from each other by floodplain deposits (e.g., Fielding and Crane, 1987). Crevasse splays in low net-to-gross fluvial intervals effectively connect channel deposits, despite their lower reservoir quality. Their contribution to producible volumes should therefore be considered in order to avoid an underestimation of net reservoir volume and long-term production rates. Similar issues are of concern in deep-marine

293   systems, where thin-bedded lobate deposits contribute to connected reservoir volumes (de Ruig and

294   Hubbard, 2006).

295   Effective connectivity may be limited to a channel-belt scale. Compensational stacking in an

296   aggradational setting causes onlap of successive channel belts. The floodplain fines at their lateral

297   fringes are likely to be preserved in absence of erosion. These floodplain fines form barriers to flow,

298   effectively compartmentalising individual channel belts unless they are connected through their

299   channel-lag deposits.

300   *6.2 Secondary reservoir potential*

301   Thin-bedded crevasse splays may constitute a secondary source of natural gas in areas where

302   production is in decline, such as the Northwest European gas province. A prolongation of production

303   by several percent at a low investment cost (e.g., Donselaar et al., 2011) would allow to postpone

304   abandonment, significantly increasing the net present value (NPV) of a field. Existing well

305   penetrations that produce from conventional reservoir intervals may be re-evaluated in order to

306   detect complexes of amalgamated crevasse splays. Borehole image logs with a high spatial resolution

307   contribute to an improved characterisation of these thin-bedded deposits (e.g., Donselaar and

308   Schmidt, 2005), which are difficult to distinguish on conventional well logs (Bridge, 2006). Low-cost

309   development scenarios may include reperforation of existing wells and side-track or infill drilling in

310   order to unlock the unswept volumes in low net-to-gross fluvial stratigraphy.

311   Intervals of amalgamated crevasse splays can also be used to estimate the dimensions of their feeder

312   channel in intervals where the latter is not directly penetrated by a well. The proposed mechanism

313   for the stacking of crevasse splays (Fig. 9) implies that the thickness of these crevasse splay

314   complexes is approximately equal to the thickness of their aggraded feeder-channel lag, given that its

315   hydraulic capacity was more or less constant. Thickness measurements in the *Castillo de*

316   *Montearagón* outcrop locality are compliant with this proportionality (Fig. 5). Larue and Hovadik

317   (2006) used channel dimensions and the net-to-gross ratio in wells to predict geobody connectivity.

7. Conclusions

The studied stratigraphic interval comprises crevasse splays, floodplain fines, and low-sinuous channel sandstone ribbons, representing the distal part of the Huesca fluvial fan. Stacked crevasse splays are encountered in intervals of up to 2.35 m thick and can be correlated between outcrop localities over a distance of 2 km.

Vertical stacking of crevasse splays implies local aggradation of the active channel belt, which may have been induced by avulsion onto a lower-lying part of the floodplain. Levees were reworked into crevasse splays with surface areas of up to several square kilometres. Lateral amalgamation created an interconnected fringe around the active channel, raising its bankfull height and the feeder-channel thalweg. This process of channel-belt super-elevation repeated until an upstream avulsion occurred.

Lateral amalgamation and vertical stacking of crevasse splays significantly increases their connected reservoir volume up to ~10^7 m^3. Despite their lower reservoir quality, these thin-bedded sheets effectively connect channel deposits in low net-to-gross fluvial stratigraphy, and hence, their contribution to producible volumes should be considered. Unswept intervals of amalgamated crevasse splays may constitute a secondary source of natural gas in brown field development situations. Their interval thickness can serve as a proxy for feeder-channel dimensions, which can in turn be used to estimate the degree of stratigraphic connectivity.

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468

Captions

Fig. 1. Study area. **(a)** Miocene paleogeography of the Huesca fluvial fan showing its line source (1),

proximal braid plain (2), distal meandering channels (3), and terminal lobes at the fan fringe (4)

(modified from Donselaar and Schmidt (2005)). Inset: map of Spain; black circle indicates location of

paleogeographic map. **(b)** Overview map of outcrop localities NE of Huesca, indicated by boxes

(Google Earth Pro). **(c)** Satellite image of the *Presa de Montearagón* outcrop locality, with the

western cut slope encircled in red (Google Earth Pro). **(d)** Satellite image of the *Castillo de*

Montearagón outcrop locality, with the studied cliff faces encircled in red (Google Earth Pro).

Fig. 2. Correlation of the sedimentary type logs over a distance of approximately 2 km. **(a)**

Sedimentary type log of the *Castillo de Montearagón* outcrop locality. **(b)** Sedimentary type log of

the *Presa de Montearagón* outcrop locality.

Fig. 3. Current roses (level II and III in Fig. 2). **(a)** Paleoflow direction of ribbon-shaped sandstone

bodies. **(b)** Paleoflow direction of thin-bedded siltstone and sandstone sheets.

Fig. 4. Lithofacies. **(a)** Thin-bedded siltstone and sandstone sheet with bed boundaries indicated in

white dotted lines. Note the ripple foresets. **(b)** Stacked thin-bedded siltstone and sandstone sheets

with interval boundaries indicated by white dotted lines. **(c)** Horizontally-laminated claystones and

siltstones with pink to purple-red structureless horizons. **(d)** Ribbon-shaped sandstone body with

reactivation surfaces indicated by white dotted lines.

Fig. 5. Left-closed histogram of thickness measurements (bars; left y-axis) and their associated

probability density functions (dashed lines; right y-axis) of channel sandstone bodies (red) and

intervals of stacked crevasse splays (blue). Note that channel thickness might be overestimated due

to unidentified reactivation surfaces.

491 **Fig. 6.** Section through stacked thin-bedded siltstone and sandstone sheets perpendicular to the

492 paleoflow direction (level II in Fig. 2). **(a)** Outcrop photo with bed boundaries marked in continuous

493 (sharp) or dotted (diffuse) black lines. Sedimentary log location indicated in black. **(b)** Interpretation

494 panel including sedimentary log. Colours according to key in Fig. 2.

495 **Fig. 7.** Section through stacked thin-bedded siltstone and sandstone sheets parallel to the paleoflow

496 direction (level I in Fig. 2). **(a)** Outcrop photo with interval of interest bounded by black lines.

497 Sedimentary log locations indicated in black. **(b)** Interpretation panel including sedimentary logs.

498 Colours according to key in Fig. 2.

499 **Fig. 8.** Distribution of median grain size in quartiles after decomposition at 0.5 m intervals from the

500 top of stacked thin-bedded siltstone and sandstone sheets (level III in Fig. 2). Component A in grey;

501 component B in orange.

502 **Fig. 9.** Schematic representation of the proposed mechanism for stacking of crevasse splays (not to

503 scale). Colours according to key in Fig. 2.; water in blue. **(a)** Pedogenesis of the inactive floodplain,

504 creating an extensive paleosol. **(b)** Increased deposition of floodplain fines announcing floodplain

505 reactivation. Upon avulsing, the active channel incises into the substrate and starts building up levees

506 through overbank deposition. **(c)** Levees are reworked into crevasse splays when they are breached.

507 The rise in bankfull height is mirrored by the elevation of the channel thalweg. **(d)** Lateral

508 amalgamation of crevasse splays creates an elevated rim around the active channel. **(e)** Levees

509 continue to build on top of the proximally-aggraded floodplain and are in turn redeposited into

510 crevasse splays, stacking on top of their precursors and building out further onto the floodplain,

511 resembling a subaerial analogy of a prograding high-stand delta. The process terminates when an

512 upstream avulsion occurs. Note that the thickness of the stacked crevasse-splay interval matches

513 that of the aggraded channel lag (orange-yellow).

514 **Fig. 10.** Conceptual diagram illustrating the nature and types of connectivity (not to scale). Colours

515 according to key in Fig. 2. Continuous lines indicate erosional surfaces; dotted lines indicate gradual

516   bed boundaries. **(a)** Lateral amalgamation and **(b)** vertical stacking proximal to the feeder channel

517   where bed bases are erosional, creating sand-on-sand contact. **(c)** Crevasse splays continuing into the

518   top of aggraded channel-lag sandstone. **(d)** Truncation of crevasse splays by channel-lag sandstone,

519   creating sand-on-sand contact. **(e)** Heterolithic channel-fill.

Highlights

- Depositional mechanisms for crevasse-splay amalgamation are proposed.

- Interconnected crevasse-splay complexes have secondary reservoir potential.

- Crevasse splays connect to channel deposits and contribute to producible volumes.

- The thickness of crevasse-splay complexes scales with feeder channel dimensions.

Fig. 1

Fig. 2

Log height [m]

A   B

level III

level II

level I

Grain size:
- medium
- fine
- very fine
- silt
- clay

Fig. 3

Channels (n = 13)   Crevasse splays (n = 28)

Fig. 4

Fig. 5

Fig. 6

Fig. 7

SSE — NNW

A

40 m

100 m

B

1 m

Fig. 8

Fig. 9

Fig. 10