Quantifying the relationship between structural deformation and the morphology of submarine channels on the Niger Delta continental slope

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Abstract
The processes and deposits of deep-water submarine channels are known to be influenced by a wide variety of controlling factors, both allocyclic and autocyclic. However, unlike their fluvial counterparts whose dynamics are well-studied, the factors that control the long-term behaviour of submarine channels, particularly on slopes undergoing active deformation, remain poorly understood. We combine seismic techniques with concepts from landscape dynamics to investigate quantitatively how the growth of gravitational-collapse structures at or near the seabed in the Niger Delta have influenced the morphology of submarine channels along their length from the shelf edge to their deep-water counterpart. From a three dimensional (3D), time-migrated seismic-reflection volume, which extends over 120 km from the shelf edge to the base of slope, we mapped the present-day geomorphic expression of two submarine channels and active structures at the seabed, and created a Digital Elevation Model (DEM). A second geomorphic surface and DEM raster—interpreted to closer approximate the most recent active channel geometries—were created through removing the thickness of hemipelagic drape across the study area. The DEM rasters were used to extract the longitudinal profiles of channel systems with seabed expression, and we evaluate the evolution of channel widths, depths and slopes at fixed intervals downslope as the channels interact with growing structures. Results show that the channel long profiles have a relatively linear form with localized steepening associated with seabed structures. We demonstrate that channel morphologies and their constituent architectural elements are sensitive to active seafloor deformation, and we use the geomorphic data to infer a likely distribution of bed shear stresses and flow velocities from the shelf edge to deep water. Our results give new insights into the erosional dynamics of submarine channels, allow us to quantify the extent to which submarine channels can keep pace with growing structures, and help us to constrain the delivery and distribution of sediment to deep-water settings.
1 | INTRODUCTION

1.1 | Rationale and study aims

The processes and deposits of submarine channels are influenced by a wide variety of external and internal boundary conditions (Picot, Droz, Marsset, Dennielou, & Bez, 2016; Wynn et al., 2012). The former comprises extrinsic factors such as eustasy, climate change, changes in sediment supply and alterations to seafloor bathymetry through tectonic activity. The latter includes morphological parameters such as channel gradient, which affects the ability of contained gravity flows to incise, transport and deposit sediment; channel width and depth, which control the size and height of the contained flow; and sinuosity (Wynn et al., 2012). Unlike fluvial geomorphology where the importance of intrinsic and extrinsic controlling variables is well-documented and quantified by a number of power-law relationships (Kirby & Whipple, 2012; Whittaker, 2012; Whittaker, Attal, Cowie, Tucker, & Roberts, 2008), the importance and interaction of controlling variables in submarine channel evolution remains poorly constrained (Covault, Fildani, Romans, & McHargue, 2011; Jolly, Whittaker, & Lonergan, 2017). In particular, the factors which control the long-term dynamics of submarine channels on complex slopes undergoing active deformation remain poorly understood.

A submarine channel's present-day geometry and longitudinal profile is sculpted by time-integrated processes of erosion and sedimentation (Pirmez, Beaubouef, Friedmann, & Mohrig, 2000; Heiniö & Davies, 2007; Talling et al., 2015). Its morphology records a history of the competing processes of incision and deposition, and the channel's response to deformation throughout its development (Covault et al., 2011; Ferry, Mulder, Parize, & Raillard, 2005; Jolly et al., 2017; Pirmez et al., 2000). In principle, a channel's long profile and morphology could be used to make inferences about tectonic activity and the formation of structures at the seabed, analogous to the way rivers are used to indicate subaerial faulting (Mitchell, 2006; Tucker & Whipple, 2002; Whittaker, 2012). To achieve this, an improved understanding of the geomorphic response of submarine channels to tectonic perturbation is required. However, a quantitative model for the long-term evolution of submarine channels remains elusive, due to the lack of direct measurements of turbidity currents (Konsoer, Zinger, & Parker, 2013; Talling et al., 2015), a limited understanding of the time-integrated erosion of submarine channels (Pirmez & Imran, 2003; Talling et al., 2015; Talling, Paull, & Piper, 2013), and the limited availability of data sets which quantify channel response to structural deformation over substantial down-system distances, particularly where a number of structural domains are traversed (e.g. Deptuck, Sylvester, & O'Byrne, 2012; Jolly et al., 2017).

Existing studies of submarine channel geomorphology frequently use seafloor bathymetry data (Covault et al., 2011; Gerber, Amblas, Wolinsky, Pratson, & Canals, 2009; Jolly et al., 2017; Konsoer et al., 2013; Mitchell, 2006) or mapped seismic-reflection surfaces (Deptuck, Steffens, Barton, & Pirmez, 2003; Deptuck et al., 2012; Deptuck, Sylvester, Pirmez, & O'Byrne, 2007; Ferry et al., 2005) to evaluate channel response to structural deformation. Bathymetric data are suitable when the submarine channel is active or very recently active, and therefore the impact of subsequent hemipelagic infill is negligible. However, if this is not the case, the seafloor may not be an accurate representation of the active channel geometry. Use of mapped seismic-reflection surfaces is likewise susceptible to interpretation errors if the impact of compaction and post-abandonment deformation is not considered, as both may significantly modify channel geometry data and long profile.

In this study, we use submarine channels with sea-bed expression in the southern lobe of the Niger Delta to quantify channel-structure interactions in time and space (Figure 1). This study significantly expands on previous work (e.g. Jolly et al., 2017) by addressing the entire channel length of the system from the shelf edge (~200 m water depth) to the base of slope (~2,850 m water depth). In doing this we incorporate a greater range of gravity-driven structural styles, including a relatively undeformed translational domain and a strike-slip transfer zone. We use three-dimensional (3D) seismic-reflection data, which extends ~120 km from the edge of the Niger Delta shelf, to quantitively constrain the geomorphic response of two submarine channels to a series of active faults and folds (Figure 2). Channel longitudinal profiles and geometries are extracted for the present-day seafloor and a seismic-reflection surface at the base of a hemipelagic sequence in the uppermost channel fill. We consider the suitability of geomorphic surfaces to measure channel morphological
parameters and we evaluate the influence of hemipelagic fill on modelled channel erosional dynamics and shear stress. Finally we constrain the response of these channel systems to deformation before considering the extent to which the channels had achieved topographic, steady state in relation to their tectonic boundary conditions.

1.2 Models of submarine channel evolution

Despite the volume of research on submarine channels in slope and deep-water settings around the world (e.g. Clark & Cartwright, 2009; Deptuck et al., 2003; Deptuck et al., 2007; Gerber et al., 2009; Gerber et al., 2008; Kostic, Parker, & Marr, 2002; Mayall, Jones, & Casey, 2006; Mayall et al., 2010; Pirmez et al., 2000; Pirmez & Imran, 2003; Posamentier & Kolla, 2003; Prattson et al., 2007; Puig, Ogston, Mullench, Nittouer, & Sternberg, 2003; Wynn, Cronin, & Peakall, 2007), only a subset of studies investigate the control of margin tectonics on the morphology and longitudinal profile of submarine channels (Covault et al., 2011; Ferry et al., 2005; Georgiopoulou & Cartwright, 2013; Huyghe, Foata, Deville, & Masle, 2004; Pirmez et al., 2000).

To understand the morphological evolution of a submarine channel with respect to structure, it is necessary to consider the morphology and longitudinal profile adopted in the absence of deformation (Covault et al., 2011; Ferry et al., 2005; Mitchell, 2006; Pirmez et al., 2000). Several studies have proposed an ‘equilibrium profile’ for submarine channels (Covault et al., 2011; Ferry et al., 2005; Georgiopoulou & Cartwright, 2013; Hodgson, de Celma, Brunt, & Flint, 2011 McHargue et al., 2011; Pirmez et al., 2000; Prather, Booth, Steffens & Craig, 1998) similar to fluvial systems where channel geometry and planform are mutually adjusted in response to changes in the contained flow. This results in a longitudinal profile that tends towards an equilibrium condition where neither significant erosion nor deposition occurs (e.g. Covault, Sylvester, Hubbard, Jobe, & Sech, 2016). Any ‘equilibrium’ profile form should reflect the interplay of sedimentary processes and tectonic deformation. Where sedimentary processes dominate

FIGURE 1 Map showing the location of the study area covering blocks OPL 245 and 256 (highlighted in orange) on the southern lobe of the Niger Delta. The locations of major structures are also shown.
tectonic deformation, it is argued that channels will adopt a smooth, concave-up profile (Pirmez et al., 2000). Where rates of tectonic deformation are high, more complex equilibrium profiles occur, reflecting both sedimentation and movement on active structures. The use of equilibrium profiles to account for erosion and deposition in turbidite systems has led to terms such as ‘above-grade’, ‘out-of-grade’ and ‘out-of-equilibrium’ for slope systems considered to have antecedent topography steeper than that of the ‘equilibrium profile’ (Georgiopoulou & Cartwright, 2013; Prather, 2003; Prather et al., 1998; Prather, O’Byrne, Pirmez, & Sylvester, 2016). Yet, despite the frequency of use, there is little consensus over what constitutes (a) an 'equilibrium profile'; (b) transient or out-of-equilibrium features, and (c) the method of readjustment in response to boundary condition change (Georgiopoulou & Cartwright, 2013; Jolly et al., 2017). While many of the concepts on which the equilibrium profile is based are taken directly from rivers, the extent to which we can transfer these concepts between environments is contentious. An examination of the disparities in fluid dynamics (e.g. decrease in effective gravity acting on subaqueous flows, and the dramatic increase in upper-boundary friction in subaqueous turbulent flows) suggests that substantial differences should be expected (Jobe et al., 2016; Konsoer et al., 2013; Peakall, McCaffrey, & Kneller, 2000; Peakall & Sumner, 2015; Straub, Mohrig, & Pirmez, 2012). In fact, there is no single predictive model for the long-term configuration of submarine channels in the way that hydraulic scaling and stream power relationships explain long profiles and morphologies in rivers (Leopold & Maddock, 1953; Whipple & Tucker, 2002; Whittaker, 2012).

Recently, there has been an increased focus on the evolution of knickpoints, representing ‘transient’ steep sections of a submarine channel long profiles, as diagnostic markers of erosion processes (Amblas et al., 2011; Covault et al., 2017; Covault et al., 2016; Heinio & Davies, 2007; Mitchell, 2005, 2006; Sylvester & Covault, 2016). Although many of these studies consider relatively limited timescales, addressing concepts such as flow segregation and hydraulic jumps (Toniolo & Cantelli, 2007), knickpoints can also represent the response of submarine channels to tectonic perturbation (Jolly et al., 2017). Mitchell (2006) presented simple, time-integrated bed-erosion models analogous to advective and diffusive models in fluvial settings to compare with submarine long profiles. These numerical models base erosion as a simple function of bed gradient or curvature to simulate whether channel topography (knickpoints) should advect i.e. migrate upstream, or diffuse without retreat. The results showed that aspects of both diffusive and advective models can replicate channels on the Barbados accretionary prism; steep reaches tended to be located upstream of active features, indicating advection, but over small lengthscale profile smoothness was consistent with diffusion. This approach provides a numerical method of estimating erosional processes within a submarine channel over time. However, the method is idealized and does not address the longer term morphodynamic evolution of the turbidity flow, such as transport flux.
variations, incorporation or deposition of suspended load, and resultant changes in flow velocity and erosivity. Given that direct measurements of turbidity, current erosivity are hard to obtain (Xu, 2010; Sumner & Paull, 2014; Talling et al., 2015; Jolly et al., 2017), Konsoer et al. (2013) proposed that a driving force argument, akin to that used in rivers, may also be used to predict the hydraulic geometry of submarine channels (cf. Jolly et al., 2017). Their results suggest that comparable power-law trends with sediment discharge may be drawn between submarine and fluvial systems, although the underlying processes and universality of this scaling remain debated (e.g. Shumaker et al., 2018). Moreover, process analogue models are yet to be tested in submarine channels interacting with active deformation in areas where tectonic rates are constrained independently (e.g. Jolly, Lonergan, & Whittaker, 2016; Pizzi, Lonergan, Whittaker, & Mayall, 2020).

Here, we address this challenge by quantifying submarine channel response to actively growing structures in the Niger Delta using good quality 3D seismic-reflection data. The documented changes in channel planform morphology and slope allow us to critically evaluate current models for submarine channel morphology; the method of readjustment to a tectonic perturbation; and give new insights into submarine channel erosional dynamics and the use of process analogues as a means to reconstruct turbidity current flow characteristics.

2 GEOLOGICAL SETTING AND STUDY AREA

The Niger Delta, in the Gulf of Guinea, is one of the largest regressive delta systems in the world with an area of ca. 140,000 km² and a sedimentary wedge ca. 12 km thick (Damuth, 1994; Kulke, 1995). The delta stratigraphy can be subdivided into three diachronous lithostratigraphic units named the Akata, Agbada and Benin Formations (Doustit & Omatsola, 1989). Down-system, the deltaic deposits of the Agbada Formation transition into sheet sands, mass transport complexes (MTCs), hemipelagic mudstones and the deep-water channel-levee systems on which this study will focus (Krueger & Grant, 2011).

Gravity-driven deformation along regional detachments within the Akata shales has significantly affected the sedimentary wedge, creating an upslope extensional domain of synthetic and antithetic listric growth faults across the shelf and uppermost slope, compensated downslope by a compressional fold-thrust system (Figure 1) (Corredor, Shaw, & Bilotti, 2005; Damuth, 1994; Rouby et al., 2011). From the Oligocene to the early Miocene, the deformation front was restricted to the inner fold and thrust belt, creating basinward-verging imbricate thrusts (Krueger & Grant, 2011). Deformation advanced in the middle Miocene forming the present-day outer fold-and-thrust belt (Krueger & Grant, 2011; Rouby et al., 2011), characterized by both basinward- and landward-verging thrusts and associated folds (Corredor et al., 2005; Jolly et al., 2016; Pizzi et al., 2020). A translational domain with deformation restricted to a few moderate-relief detachment folds separates the thrust belts (Corredor et al., 2005; Rouby et al., 2011).

The Niger Delta lobes are linked laterally by a series of transfer faults, which also detach in the mobile shale, and extend across the slope from the shelf to the toe of slope, locally offsetting thrust-fold structures (Benesh, Plesch, & Shaw, 2014; Leduc, Davies, Densmore, & Imber, 2012; Morgan, 2004). Wu et al. (2015) propose that these large strike-slip faults partition deformation between the major sedimentary lobes following major oceanic fracture zones. Smaller examples, which have also been interpreted as tear faults, accommodate strain gradients between thrust sheets (Benesh et al., 2014).

The variations in structural style make the Niger Delta an excellent location in which to investigate the interaction between submarine channels and active structures. This study covers a 6,200 km² area on the southern lobe of the Niger Delta and incorporates a range of structures from the inner to the outer fold-and-thrust belt. The seafloor bathymetry map shows two channel-levee systems with significant seabed expression traversing down regional dip (NE-SW) and interacting with a range of growing structures (Figure 3a). The geomorphic expression of the channel systems becomes markedly less pronounced up-system and they appear partially infilled with sediment, which based on its seismic expression as parallel, continuous reflections, we interpret to be a hemipelagic drape (see Section 3.2).

The northernmost part of the dataset images basinward-verging folds and thrusts of the inner fold-and-thrust belt, and locally in this area the folds trend E-W (Figures 4 and 3a). The thrusts do not propagate to the surface but their associated folds deform the seabed. To the south lies a 20 km wide, asymmetric detachment fold of the translational domain (Figure 3a); relief created by the detachment fold has been filled and the slope ‘healed’.

The south-west of the area is dominated by closely spaced, fault-propagation folds. The folds have hinge lines which strike NW-SE, perpendicular to the regional slope. Many of the structures are buried and appear to have little/no impact on present-day seafloor bathymetry. However, several of the most laterally persistent folds are marked by seabed scarps at their crests and actively deform the seabed. The structural and tectonic evolution of these structures have been documented by Jolly et al. (2016) and Pizzi et al. (2020): measurements of shortening and cumulative strain were calculated across four of the major thrust folds generating topographic relief. Line-length balancing was
FIGURE 3 (a) Three-dimensional (3-D) bathymetric map of the seabed showing the geomorphic expression of the two modern channels (black dashed lines), and structures that occur at, or near, the seabed (labelled a–g). Red lines show the location of seismic sections shown in Figure 5. (b) Close up image of the N-S transfer zone that transects the study area. Linked extensional faults form an elongate bathymetric low (transtensional pull-apart basin). X-X’ shows the location of the seismic line shown in Figure 4.
undertaken on mapped seismic horizons with their corresponding ages (derived from confidential well data provided by Shell) to investigate the variation in strain in time and space. Jolly et al. (2016) showed that these structures initiated between 15 and 10 Ma and experienced relatively rapid growth until 4 Ma. Subsequently, a slower phase of growth continued to the present. The latter period of slow structural growth is relevant for the studied channels with seabed expression.

Finally, a complex, N-S trending transtensional fault transects the study area, oblique to the regional slope dip. Features associated with the structure can be observed on the present-day seafloor where a series of en-echelon, extensional faults form a small pull-apart basin (Figure 3b). The normal faults on the basin margins NNW-SSE—oblique to the major N-S fault—and systematically offset the strike of the fault. The strike-slip deformation is interpreted to have a sinistral sense of motion as the NNW-SSE trending extensional faults indicate that the local minimum principle stress ($\sigma_3$) trended NE-SW at the time of fault movement.

3 | DATASET AND METHODS

3.1 | Seismic-reflection dataset and interpretation

A time-migrated, 3D, seismic-reflection dataset was provided by Petroleum Geo-Services (PGS). The data were migrated using Kirchhoff prestack migration and bending-ray, poststack migration to generate a 12.5 m by 12.5 m grid with a 4 ms sampling interval and were displayed every 4 in-lines and cross-lines giving a bin size of 50 m, which constrains the maximum horizontal resolution. The data were processed to near zero-phase and are displayed using SEG-normal polarity with a positive amplitude or peak representing an increase in acoustic impedance. The average frequency of the full-stack data is 50 Hz, resulting in a vertical resolution of approximately 8.25 m (assuming vertical resolution is equal to a quarter of the wavelength) and using an average interval velocity of 1,650 ms$^{-1}$ (see Supplementary Material). The horizontal resolution of the seismic-reflection volume (50 m) determines the maximum grid cell resolution (50 x 50 m) of any Digital-Elevation Model (DEM) derived from this data.

To evaluate the relationship between active structures and the submarine channels, three key seismic-reflection surfaces were interpreted: (a) the seabed; (b) a seismic-reflection surface at the base hemipelagic drape (see below); and (c) a surface interpreted to be 3.7 Ma, also documented in Jolly et al. (2016), Jolly et al. (2017). The surfaces were depth converted using a seismic interval velocity of 1,490 ms$^{-1}$ for seawater and 1,650 ms$^{-1}$ for the shallow stratigraphy isochore.

The seismic-reflection surface referred to as “the base hemipelagic surface”, was mapped due to the considerable change in hemipelagic thickness observed in the seismic-reflection data across the study area. The deposits were identified by their distinct seismic-reflection expression of low amplitude, continuous reflections that occur directly below the seafloor event and that can be traced regionally. As the two seismic sections in Figure 5 illustrate, hemipelagic thicknesses vary across the study area with greater thicknesses being observed to the north. To evaluate the impact of hemipelagic fill on the submarine channel geometries (below), analysis of the seabed and base hemipelagic surfaces was undertaken, allowing us to contrast the results from each of the surfaces.

Finally, the 3D seismic-reflection data were calibrated with seven stratigraphic horizons corresponding to ca. 3.7, 7.4, 9.5, 12.8, 15, 23.2 and 27.8 Ma, derived from Jolly et al. (2016).

3.2 | DEM derivation and channel geometries

The seabed and the base hemipelagic surface were exported to ArcGIS, georeferenced and used to create a DEM for each surface with a grid size of 50 x 50 m. Elevations within the DEMs were used to establish flow networks.
and flow accumulations using the Arc hydrology toolbox. For each of the surfaces, longitudinal profiles were extracted from cost paths that follow lines of successive low values within the DEM, starting from a user defined point source located up-system. Longitudinal profiles are presented as the bathymetric depth of the channel thalweg from sea level, $Z_m$, against down-system channel distance, $L$. Channel gradient, $S (dZ/dL)$, at any downstream distance was obtained by differentiating a high-order polynomial fitted to the channel long profile, to smooth steps and remove noise (Figure 6).

Channel width measurements were measured horizontally between external levee crests at an approximate interval of 1.5 km down system. Where levees are asymmetric, channel width was measured as the maximum length of a horizontal line at the height of the lower crest (Figure 6). In the few locations where significant terraces exist on channel margins, the bankfull width including the terrace was recorded. Values for channel depth were measured at the same locations as channel width measurements by measuring the vertical distance between the lower levee crest and floor of the channel thalweg. Channel cross-sectional areas were estimated using the channel width and depth measured between the channel levee crests and assuming a half ellipsoid channel shape to approximate the dominant channel form.

### 3.3 Distribution of strain and structural uplift

The strain analysis undertaken by Jolly et al. (2016) was extended to incorporate Fold A (Figure 3a). Strain was measured using classic line length balancing (Dahlstrom, 1969) on two sections perpendicular to the trend of the structure and approximately parallel to the flow direction of channel one, via the following equation:

$$e = \frac{(L_f - L_o)}{L_o}$$  \hspace{1cm} (1)

where $L_f$ is the final i.e. present-day length of the section and $L_o$ is the original length of the restored horizon. Values of strain were obtained for seven stratigraphic horizons for which age data are available (27.8, 23.2, 15, 12.8, 9.5, 7.4 and 3.7 Ma) and from these a cumulative strain curve was constructed to illustrate the fold growth history. The sum of the interval strains were used to construct the cumulative strain curve. The calculation involves several assumptions, the most important being that deformation occurs through layer-parallel slip with no layer parallel pure shear occurring (i.e. layer thickness and bed length are conserved during deformation). Recent work in the southern lobe of the Niger Delta shows that this approach is associated with measurement error of ±15% (Pizzi et al., 2020).

Seismic sections along the channel thalwegs were used in conjunction with the biostratigraphic data to determine the distribution of sediment thickness since 3.7 Ma (see Supplementary Material). The 3.7 Ma horizon was chosen as it is the youngest of the regional surfaces and most relevant for understanding the young channel systems. Vertical sediment thicknesses between the seabed and the 3.7 Ma seismic-reflection surface were used to derive a normalized distribution of structural uplift every 1 km along the channel lengths. The sediment thicknesses are normalized to the point of maximum crestal uplift, at which the dimensionless, normalized relative uplift, $U$, is equal to one, and the maximum sediment thickness, at which the normalized uplift is zero. Consequently,

$$U = \frac{(t_{\text{max}} - t)}{t_{\text{max}}}$$  \hspace{1cm} (2)
where \( t = t_l - t_{\text{min}} \), \( t_l \) thickness at a location, \( t_{\text{min}} \) minimum thickness at location of max uplift (see Supplementary Material). This method assumes that changes in sediment thickness relate to changes in structural uplift alone, and that sedimentation rate of prechannel stratigraphy remained constant along the channel length. Given the thrust-fold structures present, we suggest that most changes in sediment thickness are tectonically controlled.

### 3.4 Estimating submarine channel shear stress and flow velocity

Measurements of submarine channel shear stresses and flow velocities in the deep-water Niger Delta do not exist (cf. Jolly et al., 2017; Talling et al., 2015). However, Konsoer et al. (2013) showed that a channel’s bankfull geometry and gradient, \( S \), may be used to estimate the driving force per unit volume, \( f \), of a turbidity current. This can then be used, as illustrated in Jolly et al. (2017), to derive the distribution of typical bed shear stresses along a channel’s length. If the driving force, \( f \), of a turbidity flow is multiplied by the channel depth, \( H \), an estimate for the maximum value of bed shear stress, \( \tau_b \), can be obtained from:

\[
\tau_b = \rho_w R C g \bar{H} S
\]

where \( R \) is the sediment excess density (a factor of 1.65), \( \rho_w \) is the water density and \( C \) is the sediment concentration (see also supplementary material). Combining the modified Chezy equation for the bankfull flow velocity of a turbidity current, \( u \), with the above equation has been shown to approximate a minimum fluid velocity (Jolly et al., 2017) i.e.

\[
\frac{u}{C} \sim \sqrt{\frac{\tau_b}{2\rho_w C}}
\]

where the bed friction generated at the base of the flow surpasses the friction generated at the mixing surface between the flow and the above ambient fluid, and assuming flow depth corresponds to channel depth, \( H \). Equations (3) and (4) were used to approximate the variation in maximum bed shear stress and minimum fluid velocity along the entire length of two Pleistocene to recent submarine channels. This approach...
requires an estimate of the initial sediment concentration, $C$. As there are no direct measurements of deep-water turbidity current sediment concentrations measured in natural systems (Shumaker et al., 2018; Talling, Masson, Sumer, & Malgesini, 2012; Xu, 2011) a plausible range of concentrations was used, $0.002 \leq C \leq 0.004$. The values are similar to those estimated by Konsoer et al. (2013), and are comparable but at the lower end of values estimated by Pirmez and Imran (2003) (0.006–0.025).

It is necessary to consider variations in flow density along the system and whether the turbulent flow will ‘bulk’ in response to an increase in channel cross-sectional area in order to maintain the sediment concentration, or dilute reducing the sediment concentration. Two solutions are presented (cf. Jolly et al., 2017). The first assumes a constant sediment concentration implying local flow bulking at sites of increased channel area. The second assumes that sediment concentration is proportional to the cross-sectional area of the channel, thus conserving the mass of sediment and sediment discharge within the flow down-system, allowing sediment concentration to increase proportionally as the cross-sectional area decreases. Given the paucity of data in natural systems, the ability of turbidity currents to ‘bulk’ remains unknown. Therefore, we choose to present two plausible end-member models and stress that real turbulent flows are expected to lie in between these end-member solutions. Our modelling of shear stress and flow velocity assumes that bankfull approach of Konsoer et al. (2013) is appropriate and we return to these assumptions and the applicability of fluvial process analogues for submarine systems in the discussion (see Section 5.4).

4 | RESULTS

4.1 | Structural analysis

Figure 7a summarizes the strain for four folds in the area, including three from Jolly et al. (2016) and fold A, which is new to this study. Fold A is the first major fault-propagation-fold of the outer fold-and-thrust belt. Cumulative strain for fold A, like those of Jolly et al. (2016) across folds C, E and G, illustrates that fold growth rate and hence strain accumulation) varied both temporally between stratigraphic horizons (changes in the gradient of the curve) and spatially along the strike of the structure (Figure 7b). Compared to the structures measured by Jolly et al. (2016), deformation initiated earlier on fold A at ca. 23 Ma. The structure underwent relatively moderate growth until 10 Ma at which point it entered a high strain rate period. This phase of rapid structural growth continued until ca. 3.7 Ma when strain rate decreased and remained low to the present day. The maximum total cumulative strain across Fold A is $-0.32$ (32%) which is the largest of any of the measured structures. However strain for the interval of 3.7 Ma to now, which is most relevant for this study, is $-0.01$ (fold A), $-0.024$ (fold C), $-0.02$ (fold E) and $-0.026$ (fold G) which would imply shortening of 56 m for fold A over a line length of 5.1 km.

The second series of structures which influenced the sediment transport field are the extensional faults of the transfer zone (Figure 3b). A typical section along the strike of the fault zone displays a sub-vertical fault array which branches upward forming a negative flower structure, characteristic of
transtension, which results in the elongate bathymetric low of the pull-apart basin (see supplementary material). The expression of the fault changes rapidly along its length, both in terms of style of deformation and displacement, reflecting multiple phases of deformation at several points along the fault’s length. Activity on the fault zone pre-dates the two channel systems—see supplementary material—but has consisted of several phases of deformation which initiated at various locations along the faults length, forming numerous overlapping extensional faults. Throw measured on the bathymetry map across the extensional faults at the present-day seafloor is up to 110 m.

4.2 Submarine channel morphology and the influence of hemipelagic fill

The isochore map (Figure 8) shows the spatial distribution of hemipelagic deposits across the study area. Regionally, the thickness of hemipelagic drape decreases down-system forming a convergent, wedge-shaped sediment package, typical of delta-sourced depositional mud belts (Prather et al., 2016). Locally, hemipelagic thickness varies in relation to tectonic (thrust-related folds) and depositional (levee crests) bathymetric highs, thinning and onlapping onto the respective crests. Deposits correspondingly thicken into tectonic and depositional bathymetric lows such as the elongate pull-apart basin and channel axes. Proximal to the shelf edge, hemipelagic sediment thicknesses of up to 150 m occur, whilst below 2,000 m water depth, in the outer fold and thrust belt, thicknesses are typically <15 m. The thickening in the uppermost channel reaches likely reflects deposition of mud-rich turbidity currents being transported along the channels.

The recent- to present-day expression of the two channel systems range from 80 to 120 km in downstream extent with a NE-SW trend consistent with regional dip (Figure 9a). A confluence point occurs in the south of the study area. The spatial variations in cross-sectional geometry provides a comparison between the seabed (black lines), and the base hemipelagic DEM (coloured lines) (Figure 9b,c). Channel depths are greater with the removal of the hemipelagic drape, with an average increase in channel depth of >20% over measurements taken from the seabed. In contrast, the channel widths are similar, with variations >300 m limited to a few locations where the channel margins are shallow dipping (Figure 9c, Section 7). Figure 9d shows the downstream evolution of width and depth for channel one to illustrate these differences. While the trends in the data are similar, in the channel’s up-system section, the difference in measured depth between the present-day seafloor and the base hemipelagic surface is large, reflecting a substantially greater hemipelagic thickness in the channel axis as opposed to above the levees. However, with increasing downstream distance the differences in channel depths reduce significantly. Channel width is similar for both seabed and base hemipelagic DEMs.

The channel longitudinal profiles are linear to quasi-linear at a regional scale for both the present-day seafloor and the base hemipelagic surface (Figure 10). The base hemipelagic longitudinal profile for channel one is identical in length to its present-day seafloor channel equivalent. However, removing the hemipelagic thickness allowed channel two to be traced for 20 km up system onto the shelf edge and the structural highs of the inner fold-and-thrust belt. The shelf edge produces the most significant topography observed, with an elevated channel gradient and a significant inflexion point at approximately 18 km downstream distance. Despite being relatively linear at a regional scale and across the translational domain, both long profiles are characterized by steepened reaches or ‘knickpoints’ which form local convexities across the outer fold-and-thrust belt. These locations of elevated gradient typically coincide with the location of underlying structures, whether it be thrust folds or extensional faults associated with the transfer zone (see Section 4.3 for more details). While gradient changes

**FIGURE 8** Isochore map calculated between the seabed and the base hemipelagic seismic surface across the study area illustrating the regional trend in hemipelagic thickness and local variations with respect to tectonic (thrust-fold crests) and depositional (levee crests) bathymetric highs across which the deposits thin (dark blue).
at such locations are typically sharp and well-defined on the base hemipelagic surface, deposition of the hemipelagic drape has smoothed many of the inherent topographic irregularities, reducing the elevation changes observed. Nevertheless, the results of the surfaces are broadly similar, with average gradients within 5% of one another over the same channel length.

To directly compare the longitudinal profiles with one another, they can be normalized by their elevation range and maximum downstream distance. Figure 10b illustrates that at a regional scale the two profiles for the base hemipelagic surface differ in shape. Overall, profile concavity is close to zero for channel 1, and is −0.1 for channel 2. These results are consistent with other deep-water channels of the Niger Delta.
(e.g. Covault et al., 2011; Deptuck et al., 2012; Hansen et al., 2017), where linear to quasilinear profiles display steepened reaches in the vicinity of structure. As the hemipelagic infill dampens the present-day channel depths in their northernmost parts proximal to the shelf break, we interpret the base-hemipelagic DEM to be a more robust measure of the channels’ original geometry. Consequently, the base hemipelagic DEM is used for the subsequent analyses.

4.3 Most recent channel response to structural deformation

Figure 11 shows the normalized relative structural uplift, $U$, along the channel thalwegs since 3.7 Ma, and slope, width, depth measurements for the base hemipelagic surface.

There is a substantial range of channel depths from 5 to 170 m. Channel one is significantly deeper and more heavily entrenched than channel two (40 ± 30 m), with a mean channel depth of 80 ± 40 m. Channel depth generally increases down system in both channels, however, it is highly sensitive to underlying structure, with maxima typically being localized to a structure or a series of closely spaced structures. Conversely, the channel width data generally increase down-system in the case of channel 1, and are relatively stable in the case of channel 2. Channel two is the narrower system with a mean channel width of 1,500 ± 650 m compared to 2,700 ± 850 m of channel one.

In channel system one the mean channel gradient is 0.013 (0.7°), however, across several knickpoints which occur above the crest of major structures, channel gradient can locally be up to 0.025 (1.4°). Channel one interacts with five major thrust-related structures along its length, a single fault-propagation fold at 41 km (fold A), three closely-spaced structures between 52 and 65 km (folds B, C and F), and another individual fault-propagation fold at 75 km (fold G). Channel gradient increases sharply at each fold and reaches a maximum above the crest of the structure, which corresponds to a peak in the normalized relative uplift. The increase in channel gradient over the thrust folds is accompanied by a prominent increase in channel depth, with an increase of 30 m (46%) across fold A and by up to 63 m (60%) across folds B, C and F. Notably, over the
same regions of elevated uplift, the response of channel width does not show a consistent signal—width increases over some structures and decreases over others. Across the crest of fold A at 41 km downstream distance, channel width increases significantly by over 700 m before rapidly decreasing downdip of the structure by 1,000 m; however, over fold F at 63 km, channel width decreases to ~1,000 m over the crest of the structure before increasing down-system. Despite lacking a systematic response, the bankfull widths of channel one are generally larger and more variable in the vicinity of the fold structures. The channel segments interacting with the outer fold-and-thrust belt also show a marked increase in the channel sinuosity.

Channel two can be divided into two sections. In the updip section (0–70 km) the channel is confined by extensional faults associated with the N-S trending transfer zone. Strike-slip displacement along the fault has controlled the location and configuration of the channel that has a N-S trend over this 70 km section—parallel to the fault axis. Furthermore, the channel is diverted around several of the normal fault structures with larger throws (at 12, 24 and 35 km on Figure 9a). The confinement and diversion of the channel course suggest deformation on these structures pre-date the channel development and controlled the slope gradient at the time of channel development. Fault-related offset of the channel margins across segments where the channel runs parallel to fault trends i.e. 10–20 km and 32–37 km, results in marked reductions in the apparent channel depth. Despite the initially high gradient of the channel proximal to the shelf break, the upper channel section is less steep with an average gradient of 0.014 (0.8°). The reduced channel gradient reflects the channel orientation over this length, which is oblique to regional NE-SW dip. Excluding the fault-controlled shallowing, channel width and
depth remain fairly consistent over the channel’s upper section averaging 1,450 ± 580 m and 35 ± 16 m, respectively. Minor knickpoints as the channel cross-cuts extensional faults at 20, 37 and 44 km see an increase in the normalized uplift and the channel gradient increases locally (over 10² m length scales) up to 0.02 (1.2°) (Figure 11b). The structurally elevated gradients result in relatively moderate increases in channel depth of less than 20 m, and no discernible change in width.

In contrast, the lower section is not confined by the transfer zone, and has a mean channel gradient of 0.019 (1.1°), somewhat higher than its up-dip equivalent, as observed on the channel long profile. Channel two interacts with three major structural zones within this section, fold A at 74 km along channel distance, folds D, E and F between 94–101 km, and fold G at 108 km downstream distance (Figure 11b). Channel gradient reaches a maximum above the crest of the three major structural zones, increasing by up to a factor of 2.8 (fold G). Across fold A, near the thrust’s eastern tip, there is a moderate 10 m increase in channel depth and a negligible change width. The region of elevated uplift around fold E, however, is accompanied by a rapid 2,350 m (160%) increase in channel width to a maximum value of 3,800 m (Figure 9c, transect 7) and a 80 m (105%) increase in channel depth. Fold G likewise results in an increase in channel width and depth, albeit to a more moderate extent—700 m and 10 m change in width and depth, respectively.

Figure 12 provides a comparison of the aspect ratios of the two channels across the translational domain and the outer fold-and-thrust belt. Regionally, aspect ratios decrease downstream in both systems, reflecting the deepening of both channels as they traverse the slope. Across the translational domain, the aspect ratios of channel one and two remain relatively high with means of 42 ± 11 and 42 ± 14, respectively. The outer fold-and-thrust belt sees channel aspect ratios decrease to 28 ± 9 and 26 ± 5 for channel one and two, respectively. A Kolmogorov–Smirnov test demonstrates that aspect ratios from the two domains represent different probability distributions (see supplementary material). This decrease in aspect ratios as the slope transitions from the unstructured translational domain to the stepped fold-and-thrust belt reflects the substantial increase in system depth observed between the two domains.

Overall, the data show that channel gradient and depths are the most sensitive morphological parameters to deformation. Knickpoint development and increases in channel depth over short-length scales are limited to channel reaches associated with underlying structure and elevated uplift. In contrast, channel width lacks a consistent response to tectonic perturbation.

4.4 | Estimated channel shear stress and flow velocity

Two estimates for bed shear stress and flow velocity down system are presented using (1) a constant sediment concentration, and (2) varying sediment concentration, modelled to be inversely proportional to the cross-sectional area and maintaining constant sediment flux (see supplementary material). The varying sediment concentration model produces lower values for both shear stress and flow velocity, particularly in areas impacted by structure, due to dilution of the modelled flow at locations of large cross-sectional area. The two solutions represent end-members and illustrate the distribution of shear stresses and flow velocities for a bankfull flow.

The shear stresses and flow velocities differ between the upper translational domain and the outer fold-and-thrust belt. In the translational domain, shear stress and flow velocity for channels one and two—using both constant and variable sediment concentrations—remain remarkably constant and within

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**Figure 12** Plot of channel aspect ratios (Width/Depth) against a downstream distance. The measurements of channel one have been adjusted to show its aspect ratio relative to channel two to allow direct comparison of the systems across the translational domain and the outer fold-and-thrust belt as their lengths differ in absolute terms. Both systems are therefore referenced to the point at which channel two enters the translational domain, 18 km downstream distance.
a narrow range of values from 40–100 Pa and 1.5–3 ms\(^{-1}\) and 20–100 Pa and 1–2.8 ms\(^{-1}\), respectively. In contrast, the results of the outer fold-and-thrust belt are characterized by both large downstream fluctuations in shear stress and flow velocity and significant variability between the results of the two models. Maxima, with values raised by up to a factor of 3, are limited to locations of high normalized uplift above structures and channel erositivity (Figure 13a,b), and typically have significantly greater range of values. Channel one shows a maximum predicted shear stress of 100 to 230 Pa and maximum flow velocity ~2 ms\(^{-1}\) to ~4.5 ms\(^{-1}\) for constant values of sediment concentration between 0.002 and 0.004, respectively. The location of the maximum values occur at 40–42 km and 52–68 km down-system (Figure 13a, yellow envelope). Whilst the results for the varying sediment concentration model also show increases at these locations, the increase is less marked, with a peak value of 150 Pa at ~41 km and a substantially dampened increase between 52 and 68 km, due to flow dilution (Figure 13a, purple envelope). The locations of maximum shear stress and flow velocity at 40–42 km and 52–63 km down-system clearly coincide with sections of the channel which correspond to increased time-integrated uplift associated with underlying structures of the outer fold-and-thrust belt; correspondingly they have raised channel gradients. Results for channel two similarly show high values for the maximum predicted shear stress and flow velocity across several thrust-fold structures between 95–100 km down-system, with a peak shear stress of ~270 Pa and peak velocity of ~5.1 ms\(^{-1}\), for the constant sediment concentration model (Figure 13b, yellow envelope). However, the variable sediment concentration model shows an opposing relationship with underlying structure, decreasing shear stress and flow velocity over the crest of folds (Figure 13b, purple envelope). This outcome signifies dilution—a decrease in sediment concentration associated with an increase in channel cross-sectional area (consequence of maintaining constant sediment discharge)—has a greater effect on the modelled flow than the structurally elevated gradient, creating an overall decrease in shear stress and flow velocity in locations impacted by structure.

These data indicate that for sediment concentrations of 0.002–0.004, calculated shear stresses and flow velocities are relatively consistent across the translational domain ranging from 20–100 Pa and 1–3 ms\(^{-1}\) for both models. However, across the outer fold-and-thrust belt values increase up to 20–270 Pa and 1–5 ms\(^{-1}\). The locations of maxima coincide with underlying structure. Consequently the documented modification of channel morphology results in increased shear stress and flow velocities, allowing the channels to keep pace with the developing structures.

5 | DISCUSSION

5.1 | Influence of hemipelagic drape

This study analyses channel morphology using two geomorphic surfaces, the present-day seafloor and a seismic-reflection surface, traced regionally, at the base of the recent hemipelagic
deposits which partially infill the two systems. Whilst the results obtained are broadly similar, they represent different stages in channel development; the present-day seafloor data represents the systems at some stage during post-abandonment infill, whilst the base hemipelagic surface reflects the systems at a recent active stage. The hemipelagic deposits affect channel morphology in two ways, significantly decreasing present-day seafloor channel depths in the proximal reaches and smoothing structurally controlled topographic irregularities in the long profiles. The seismic stratigraphy of the hemipelagic sequence is very similar to that observed by Jobe et al. (2015) in the Y channel system on the western lobe of the Niger Delta. They demonstrated that the latter stage of ‘phase 4’ of channel Y’s development was dominated by mud-rich, dilute turbidity flows, which draped onto the abandoned channel section. Cored thicknesses were up to 20 m at 1,000 m water depth, comparable to those observed here. We propose a similar source for the hemipelagic deposits, with a combination of deposition of mud-rich, dilute turbidity flows as a result of waning flow speeds with increasing run out distance and pelagic fall out. Jobe et al. (2015) date ‘phase 4’ of the Y channel from 44 ka to abandonment at ca.19 ka on the basis of radiocarbon dating of Globigerinoides ruber. We suggest a similar age for our hemipelagic drape, although we cannot rule out the previous glacial cycle (MIS 5e or 6) as the point at which drape accumulation started. The fact that these channels have pronounced seabed expression with relatively little infill would not be consistent with a mid Pleistocene or earlier abandonment. The low strain rates since 3.7 Ma (56 m shortening for fold A over this time period) implies the hemipelagic surface has not been significantly deformed over timescales of <10⁵ years.

Exclusion of hemipelagic drape in analyses of seafloor submarine channel morphology is desirable due to the smoothing of long profiles and shallowing of channel depths proximal to the shelf edge. Its inclusion in analyses may lead to the misinterpretation of channel geometries, underestimation of a channel’s ability to incise, and over estimation of the importance of diffusive, rather than advective processes in long profile evolution.

5.2 | Morphology of channels and the influence of gravity-driven tectonics

This study demonstrates the planform and cross-sectional geometries of submarine channels are sensitive to both tectonic boundary conditions and long-term erosional dynamics. Several previous studies have outlined both qualitative (Ferry et al., 2005; Georgiopoulou & Cartwright, 2013; Huyghe et al., 2004; Pirmez et al., 2000) and quantitative (Mitchell, 2005, 2006) models for submarine channel readjustment to tectonic perturbation. The morphological data presented here allow us to critically test several of the conceptual models for channel readjustment and equilibrium profile development.

The geometry of knickpoints has the potential to reveal the time-integrated style of erosion and deposition and the degree to which channel incision is detachment- or transport-limited (Mitchell, 2006). The initial ‘equilibrium profile’ model of Pirmez et al. (2000) envisaged topographic irregularities in submarine long profiles to be smoothed out through the positive feedback of enhanced upstream erosion and downstream deposition. The model, analogous to transport-limited erosion models in fluvial settings, would see the irregularity diffuse without retreat in a process governed by longitudinal profile curvature. Erosion and deposition would occur in convex and concave up segments, respectively. Other studies (Ferry et al., 2005; Georgiopoulou & Cartwright, 2013; Mitchell, 2006) have supported this method of longitudinal profile readjustment and expanded upon it to include changes in channel width and sinuosity. A contrasting model for long profile readjustment is demonstrated in the numerical models of Mitchell (2006) and described by Heiniö and Davies (2007) where topographic irregularities (e.g. knickpoints) migrate upstream. The model bases erosion as a simple function of channel gradient and, as illustrated by Mitchell (2006), results in advection of the convexity upstream in a way similar to detachment-limited knickpoint migration in river systems (cf. Whittaker, Cowie, Attal, Tucker, & Roberts, 2007). The presence of structurally-controlled knickpoints in the outer fold-and-thrust belt section of channel systems one and two allow us to test qualitatively the various erosion models described and evaluate which of the numerical solutions presented in Mitchell (2006) best represents the erosion exhibited in the channels.

By contrasting the gradient of the background slope profile of the outer fold-and-thrust belt to the long profile of channel one, we can interpret the system’s erosional dynamics and the method of long profile readjustment to a tectonic perturbation. Figure 14 shows the slope longitudinal profile in addition to the base hemipelagic and present-day seafloor long profiles for channel one. Of the three profiles the overbank slope profile has the most substantial topographic irregularities across folds A (92 m) and F (51 m), whilst folds B and C have more muted topographic signals due to the proximity of the section to the fault tips. The channel topography on both geomorphic surfaces is substantially smoother across the structures A and F than the driving tectonic topography represented by the slope profile. Hence, we propose the time-integrated erosion of the channel, and the system’s longitudinal profile must have a diffusive component. However, whilst the channel appears to have predominantly smoothed the knickpoint located above fold A, the lip also appears to have advected approximately 400 m upstream. Furthermore, across folds B and C the related knickpoints occur 800 m and 200 m upstream of their overbank equivalent, respectively. Upstream advection of knickpoints from folds B and C in response to tectonic uplift may explain how high channel
depths are sustained across the region despite channel gradient increases being shorter and higher frequency. Migration of knickpoints from folds B and/or C toward the large knickpoint above fold A would result in low present-day gradients with high channel depths generated when the knickpoint was more proximal. The knickpoint geometries indicate that neither advective nor diffusive erosional dynamics alone generated the long profile topography observed.

Mitchell (2006) similarly found both advective and diffusive components to be well represented in submarine channels in the Barbados accretionary prism. However, whilst these numerical models provide a means to estimate the time-integrated erosivity of a submarine channel, they apply only to the channel long profile and gradient. A key outcome of Figure 14 is that the suitability of the present day seafloor as a geomorphic surface must be considered when interpreting knickpoint morphology and channel erosional dynamics. The present-day long profile highlights the smoothing effect of hemipelagic fill on knickpoints, therefore studies where channel activity has not been considered or studies on inactive systems using bathymetry data will likely over-represent the impact of diffusive processes in channel erosion. Our results demonstrate that the response of a submarine channel to tectonic deformation is through locally increasing channel gradient and noticeably increasing channel depth, and that over time bed-erosion in relation to the topographic irregularity will display both diffusive and advective components. The migration of knickpoints upstream supports a model of long-term erosivity that is both advective and gradient-dependent, advocating the suitability of using bed-shear stresses for erosivity.

A final key implication of our results is that the observed changes in channel aspect ratio in relation to channel gradient, and hence underlying structures, is related to channel deepening as opposed to a reduction in width. Channel widths lack a systematic response to structurally elevated gradients whilst channel depths increase substantially, decreasing channel aspect ratio. Shumaker et al., (2018) similarly interpret submarine channel depth to be the primary control on aspect ratio, citing stable levees and their resistance to erosion as factors limiting system width. However, this relationship is in stark contrast with fluvial systems, where decreases in channel width in response to uplift and local gradient increases is well-documented (e.g. Schumm, Dumont, & Holbrook, 2002; Whittaker et al., 2007; Attal et al., 2008). Previous work on bedrock rivers (e.g. Whittaker et al., 2007) has shown that tectonic, slope-driven narrowing drives change in aspect ratios and is an intrinsic process by which fluvial systems maximize erosional response through increasing shear stress. In contrast, our results indicate that channel deepening is the primary response of submarine channels to tectonically elevated gradients, driving changes in aspect ratios, and that deepening is the principal way in which channels increase their shear stress and stream power in order to keep pace with growing structures.

5.3 Thrusting versus transtension

The two systems studied here give new insights into channel response to seafloor deformation from structures associated with both extension and contraction.
Throughout the translational zone (0–70 km downstream distance) channel two is structurally confined by the elongate bathymetric low associated with transtension along the transfer fault (Figure 15). The presence of multiple pre-existing, overlapping extensional faults results in the regional diversion of the channel oblique to the slope dip, creating a significant convexity in the long profile (54–70 km) (Figure 10b). Further, it is apparent from the bathymetry map that motion along the fault zone has also resulted in periodic reorganization of the sediment transport field, with channel one being offset from its up-dip source by the kinematically linked normal faults associated with the transfer zone (Figure 15). The initiation point of channel one is located on the footwall high of one of the extensional faults which separates it from the bathymetric low containing channel two. Given the proximity of the two channel systems—separated only by a fault plane—it seems likely the uppermost part of channel two once transported sediment into channel one, before fault development resulted in the abandonment of channel one and an avulsion with sediment being rerouted down channel two. This abandonment and avulsion of channel two out of channel one in response to extensional fault development is in contrast to its response to the thrust-fold structures occurring down-dip where the channel is able to modify its geometry—increasing shear stress up to 270 Pa— and incise across the growing structures allowing channel one to keep pace with the growing thrust-fold structures. This outcome suggests that drainage capture, analogous to fluvial systems, plays an important but often overlooked role in determining the routing of channels to deep water.

Clark and Cartwright (2009) attribute channel diversion to the presence of pre-existing structural relief, oriented at a high angle to regional dip. In contrast, these results indicate that the growth of a fault plane, oriented oblique to regional dip, is able to sufficiently change flow paths such that channel one is abandoned and flow diverts down channel two, whilst the thrust-fold structures perpendicular to regional dip are not. Moreover, relief created through both sets of structures precedes the inferred initiation of channel one. We propose that the ability of the extensional faults to divert the channel systems instead relates to: an interplay of their orientation relative to structural dip, the uplift profile and growth rate across the structures and the power of the entrained flow. The low angle of the structures relative to regional dip allows the flow to continue downslope albeit at a slightly lower gradient. In contrast, were the channels to be diverted by one of the thrust structures, the system would be required to flow perpendicular to regional dip for >10 km. Furthermore, unlike the thrust-fold structures where channel geometries appear largely adjusted, deformation rates across the extensional faults may be too high for the channel to sufficiently modify its morphology to keep pace with steep uplift profile of a fault plane with surface expression. While deformation rates are difficult to quantify, the seabed DEM indicates throws across these structures of up to 110 m, compared to <60 m of shortening over 3.7 Ma for fold A over a line length of 5 km.

5.4 The application of bankfull analysis in submarine channel systems

Fluvial process analogues, such as the bankfull analysis undertaken in this study, may provide a means of estimating the dynamics and formative processes of submarine channels based on their morphological characteristics (Gerber et al., 2009; Konsoer et al., 2013). However, debate remains as to the extent to which the morphometric scaling relationships that underpin fluvial geomorphology may be applied to
submarine channels with discussion centering on the differences in fluid dynamics (effective gravity, shapes of velocity and sediment concentration profiles) (Konsoer et al., 2013; Peakall et al., 2000; Straub et al., 2012) and the relationship between flow discharge and channel distance (Shumaker et al., 2018). Shumaker et al. (2018) document opposing scale-dependency of channel aspect ratios in fluvial and submarine channels with the latter often exhibiting a decrease in aspect ratio with increasing channel size. They propose submarine channels with large cross-sectional areas and low aspect ratios are linked to the larger sediment discharges associated with proximal settings, whilst higher channel aspect ratios and smaller cross-sectional areas are associated with greater downstream distances where discharges are lower. This detachment of the submarine channel width- and depth-discharge relationships to that of fluvial systems is in opposition to the findings of Konsoer et al. (2013). However, Shumaker et al. (2018) stress that channels on the Niger Delta provide a counter-example to this trend and exhibit complex, nonlinear fluctuations in channel aspect ratio and scale down system. Our study demonstrates that channel aspect ratio is explicitly tied to the underlying structural template and channel gradient. While both channels studied here exhibit an inverse relationship between channel size and aspect ratio, channel size increases significantly down system in both channels and is primarily a function of channel depth associated with structurally elevated slopes—a factor not considered in Shumaker et al. (2018) (Figure 11; see supplementary material).

Given (i) the complex disparities between fluvial and submarine channel morphometric scaling relationships and (ii) the sparsity of data on submarine channel discharges and sediment concentration and velocity profiles in natural systems, we argue the bankfull end-member models presented provide a useful first-order evaluation of the structural influence on submarine channel dynamics. Bankfull depth and channel gradient provide consistent, quantifiable geomorphic markers across the study area. The assumption of bankfull flow may be limiting if the aim is the accurate reconstruction of a specific flow, but this does not nullify the use of the models as end-member solutions. Velocity and sediment concentration maxima of turbidity currents typically occur near the base of the flow (Pirmez & Imran, 2003; Shumaker et al., 2018). This high-velocity core is typically thinner than the full channel or canyon form, and is the portion of the flow responsible for actively sculpting the meandering planform shape of the channel (Pirmez & Imran, 2003). In the models presented, where a flow is not bankfull—the existence of inner levee deposits indicate that some were not—then the shear stress and velocity calculations for the constant sediment concentration model will be reduced as the input flow height, $H$, is less. However, the effect of a nonbankfull flow on the variable sediment concentration model would yield contrasting results. While a decrease in flow height would decrease shear stress and flow velocity proportionally, the reduced cross-sectional area occupied by the flow will increase sediment concentration, and hence the output shear stress and flow velocity, by an amount approximately proportional to the square of height decrease. Consequently, the constant and variable sediment concentration models illustrate plausible distributions of bed shear stress and flow velocity for turbiditic flows that ‘bulk’ in response to increases in channel cross-sectional area and that maintain constant sediment discharge, respectively.

5.5 Implications for the concept of an equilibrium profile

An important question when evaluating the geometries and longitudinal profile of a submarine channel is the extent to which the system is in equilibrium or topographic ‘steady state’ with respect to its boundary conditions. It is frequently assumed that submarine channels will revert toward an equilibrium state represented by a concave-up longitudinal profile (Covault et al., 2016; Deptuck et al., 2012; Ferry et al., 2005; Pirmez et al., 2000). As a result, the presence of knickpoints is often used to suggest that a channel is in a transient state. However, several studies have demonstrated that a range of profile shapes exist from strongly concave (i.e. the Laurentian system) to convex (i.e. the Brazos-Trinity system) (Mallarino et al., 2006), even in margins which are not tectonically perturbed (Covault et al., 2011). In fact, it is knickpoint migration that is a key piece of evidence for transient long profile evolution.

One of the key observations is that throughout the translational domain channel one's longitudinal profile is remarkably linear and channel width, depth, gradient and sinuosity remain stable (Figures 9 and 11a). The channel undergoes a relatively linear increase in channel depth, but values stay within a moderate range of $54 \pm 11$ m, and with few fluctuations away from the mean width, $2,150 \pm 550$ m. Consequently, channel aspect ratios likewise remain consistent, yet high from $42 \pm 10$ (channel 1) throughout the translational domain. The stability of channel morphological parameters is reflected in the estimated shear stresses and flow velocities which remain within a narrow range of values from $40–100$ Pa and $1.5–3$ m s$^{-1}$ for both constant and variable sediment concentrations. Our results show, that in the absence of significant deformation, submarine channels may develop linear longitudinal profiles and maintain stable ranges for width, depth and gradient at relatively high aspect ratios and in apparent topographic steady state.

The transition from the graded slope profile of the translational domain to the stepped slope of the outer fold-and-thrust belt marks the transition to a setting where deformation
dominates the long profile evolution. We interpret the up-stream migration of knickpoints to imply that the channels had not reached topographic steady state across the full uplift profile of the structures despite the degree of incision at the crest indicating they are capable of matching uplift locally as demonstrated by Jolly et al. (2017). We therefore interpret the recent hemipelagic sediment drape to cover a system that was undergoing a transient response to active deformation near the sea bed.

6 | CONCLUSIONS

Using 3D seismic-reflection data, this study quantifies the geomorphic response of two late Pleistocene submarine channels to a range of gravity-driven structural styles on the Niger Delta slope from water depths of 200 m to >2,800 m. The channels cross an undeformed translational domain, a strike-slip transfer zone, and a deep-water fold-and-thrust belt enabling us to evaluate their geomorphic response to a varying structural template. Cumulative strain profiles, independently reconstructed using line-length balancing, for several folds of the outer fold-and-thrust belt allowed changes in channel planform and cross-sectional geometry to be analysed with a quantitative link to their tectonic boundary conditions. Channel longitudinal profiles and geometries were measured on two geomorphic surfaces, the present-day seafloor and a seismic-reflection surface which removed recent hemipelagic deposition, each representing a different stages in the channels’ evolution. The present-day seafloor data represents the system at some stage during post-abandonment infill, whilst the base hemipelagic surface is interpreted to reflect the system at a recent active stage.

An isochrone of the hemipelagic deposits highlights the down-system thinning with localized variations in thickness, relating to tectonic and depositional bathymetric highs, superimposed on the regional trend. This impacts channel morphology in two key ways, decreasing present-day seafloor channel depths in their upper reaches and smoothing structurally-controlled topography in the long profiles. The longitudinal profiles of the two channel systems are linear to quasilinear at regional scale with concavities ranging from approximately zero for channel one to ~0.1 for channel two. However, while linear across the unstructured translational domain, profiles are characterized by short length-scale convexities relating to structure across the outer fold-and-thrust belt. Locations of structurally elevated channel gradient are typically associated with a doubling of channel depth, which may represent the system acting to increase the confinement of entrained flows upstream of active structures. In contrast, channel widths, whilst increasingly variable in proximity to structure, lack a systematic response.

Using morphological data, we estimate channel bed shear stress and flow velocity for a constant and variable sediment concentration, and consider the application of process analogues in estimating the behaviour and formative processes of submarine channels. Shear stress and flow velocity remain stable over the unstructured translational domain ranging from 20–100 Pa and 1–3 ms\(^{-1}\) for sediment concentrations of 0.002 to 0.004. However, they increase up to a maximum of ca. 270 Pa and 5 ms\(^{-1}\) across the fold-and-thrust belt.

Our results demonstrate planform and cross-sectional geometries of submarine channels are sensitive to both the tectonic template and the channels’ long-term erosional dynamics, with knickpoint development and incision being unambiguously linked to variations in uplift across underlying structures. Structurally-controlled knickpoints in channels one and two over the outer fold-and-thrust belt allowed us to test current models of submarine channel erosion. Knickpoint geometries, which appear to have migrated upstream but smoothed over short-length scales, indicate that neither advective nor diffusive erosional dynamics alone generate the long profile topography observed. Upstream migration of knickpoints indicates that the channels were yet to reach topographic steady state across the full uplift profile, and therefore were in a transient state at the onset of hemipelagic deposition.

Our results provide new constraints on the time-integrated dynamics of submarine channel response to tectonic perturbation and offer fresh insight into the use of process analogues as a means to reconstruct flow characteristics from seismically-derived geomorphic data.

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DATA AVAILABILITY STATEMENT

The data that support the findings of this study are provided in the supplementary material.

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**SUPPORTING INFORMATION**

Additional supporting information may be found online in the Supporting Information section.