The formation and evolution of submarine headless channels

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[1]Energy and Environment Institute, University of Hull, Hull, HU6 7RX, UK

[2]Department of Geography, Geology and Environment, University of Hull, Hull, HU6 7RX, UK

[3]Departments of Earth Sciences and Geography, University of Durham, Stockton Road, Durham, DH1 3LE, UK

[4]School of Ocean and Earth Sciences, University of Southampton, Southampton SO14 3ZH, UK

[5]Ocean Biogeoscience, National Oceanography Centre, Southampton SO14 3ZH, UK

[6]Center for Coastal and Ocean Mapping, University of New Hampshire, Durham, New Hampshire 03824, USA

[7]Natural Resource Canada, Geological Survey of Canada, Sidney, British Columbia V8L 4B2, Canada

[8]Department of Geoscience, University of Calgary, Calgary, T2N 1N4 AB, Canada

[9]Faculty of Civil Engineering and Geosciences, Delft University of Technology, Delft, The Netherlands

Email: Ye.Chen-2016@hull.ac.uk

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[4]Ocean Biogeoscience, National Oceanography Centre, Southampton SO14 3ZH, UK

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[7]Natural Resource Canada, Geological Survey of Canada, Sidney, British Columbia V8L 4B2, Canada

[8]Department of Geoscience, University of Calgary, Calgary, T2N 1N4 AB, Canada

[9]Faculty of Civil Engineering and Geosciences, Delft University of Technology, Delft, The Netherlands

Abstract: The scale of submarine channels can rival or exceed those formed on land and they form many of the largest sedimentary deposits on Earth. Turbidity currents that carve submarine channels pose a major hazard to offshore cables and pipelines, and transport globally significant amounts of organic carbon. Alongside the primary channels, many systems also exhibit a range of headless channels, which often abruptly terminate at steep headscarps. These enigmatic features are widespread in lakes and ocean floors, either as branches off the main submarine channel thalweg or as isolated secondary channels. Prior research has proposed that headless channels may be associated with early and incipient stages of channel development, but their formation and evolution remain poorly understood. Here, we investigate the morphology, origin and development of headless channels by examining repeat bathymetric surveys spanning a period from 1986 to 2018, in Bute Inlet, Canada. We show how channel switching processes, the extension of turbidity currents across distal fans, along with overbanking turbidity currents, are able to initiate headless channels in submarine settings. We discuss how the evolution of headless channels plays an important role in shaping submarine channels, promoting channel extension and modifying the overall longitudinal profile, as well as impacting the character of sedimentary records in channel-lobe transition zones.

INTRODUCTION

Submarine channels act as the primary conduits for the delivery of globally significant amounts of sediment, pollutant and organic carbon into the deep-sea (e.g., Bouma, 2000; Paull et al., 2011; Hage et al., 2020; Zhong and Peng, 2021). They extend over distances of hundreds of kilometres (e.g. Talling et al., 2013) and pose significant hazards to offshore infrastructure (Carter et al., 2014). Although submarine channel inception and subsequent development have drawn significant attention in the last few decades (Imran et al., 1998; Covault et al., 2014; De Leeuw et al., 2016), limited work has been undertaken on the formation and evolution of headless channels. These enigmatic features can branch off the main channel thalweg or exist as isolated secondary channels, often exhibiting a steep break in slope at the channel head, which can often take the form of a headscarp. They have been observed both on lake floors (Girardclos et al., 2012; Turmel et al., 2015; Corella et al., 2016) and within submarine systems (Paull et al., 2011; Gales et al., 2019).

Despite their wide occurrence, the role of headless channel in the generation and evolution of submarine channel systems is uncertain, especially as headless channels do not connect to obvious sources for turbidity currents, such as river mouths. Previous research has suggested that submarine channels themselves are initiated or extended by trains of scours that are likely formed by the interaction of overriding turbidity currents with seafloor perturbations (Fildani and Normark, 2004; Covault et al, 2014). Zones where turbidity currents spread laterally as a result of a transition from confined to unconfined environments are favourable locations for new channel formation (e.g. Fildani et al., 2006). Channelization across subaqueous fans by overpassing turbidity currents have been suggested as a mechanism that can induce avulsion (e.g. Yu et al., 2006) with channel abandonment and channel filling thereby resulting in the formation of abandoned headless channels (Hamilton et al., 2015). Additionally, Fildani and Normark (2004) have suggested that initiation of such new flow pathways is often preceded by discontinuous channel features, consisting of a series of scours produced by the overspilling turbidity currents (Maier et al., 2011; Fildani et al., 2013; Covault et al., 2014). These incipient channels can form basinward of the main channel system, with their connectivity leading to overall downstream channel extension (Fildani et al., 2013;

Hamilton et al, 2015; Pohl et al., 2019). As such, the formation and evolution of headless channels is likely to play an important role in channel inception and longer-term downstream extension of channel systems as well as controlling long-term sedimentary architecture.

In this study, we examine the morphodynamics of submarine headless channels using a >30 years (1986-2018) record of bathymetric surveys, extending from source to sink, from the submarine channel system in Bute Inlet, Canada. Our aims are to understand: (1) the spatial and temporal distribution and evolution of submarine headless channels; (2) how headless channels are formed in different settings, such as channel-lobe transition zone (CLTZ), within distal fan region and within channel overbank areas, and (3) how headless channels affect downstream channel extension and wider channel system evolution, and the subsequent deposits that are preserved in the sedimentary record.

STUDY SITE AND METHODS

Bute Inlet is a fjord located on the SW coastline of British Columbia, Canada (Fig. 1). A sandy floored submarine channel is incised into Holocene fjord-bottom silts and clays, created by turbidity currents from the Homathko and Southgate River deltas at the fjord head (Zeng et al., 1991; Heijnen et al., 2020). The submarine channel extends over 44 km and reaches water depths of >600 m (Prior et al., 1987; Chen et al., 2021).

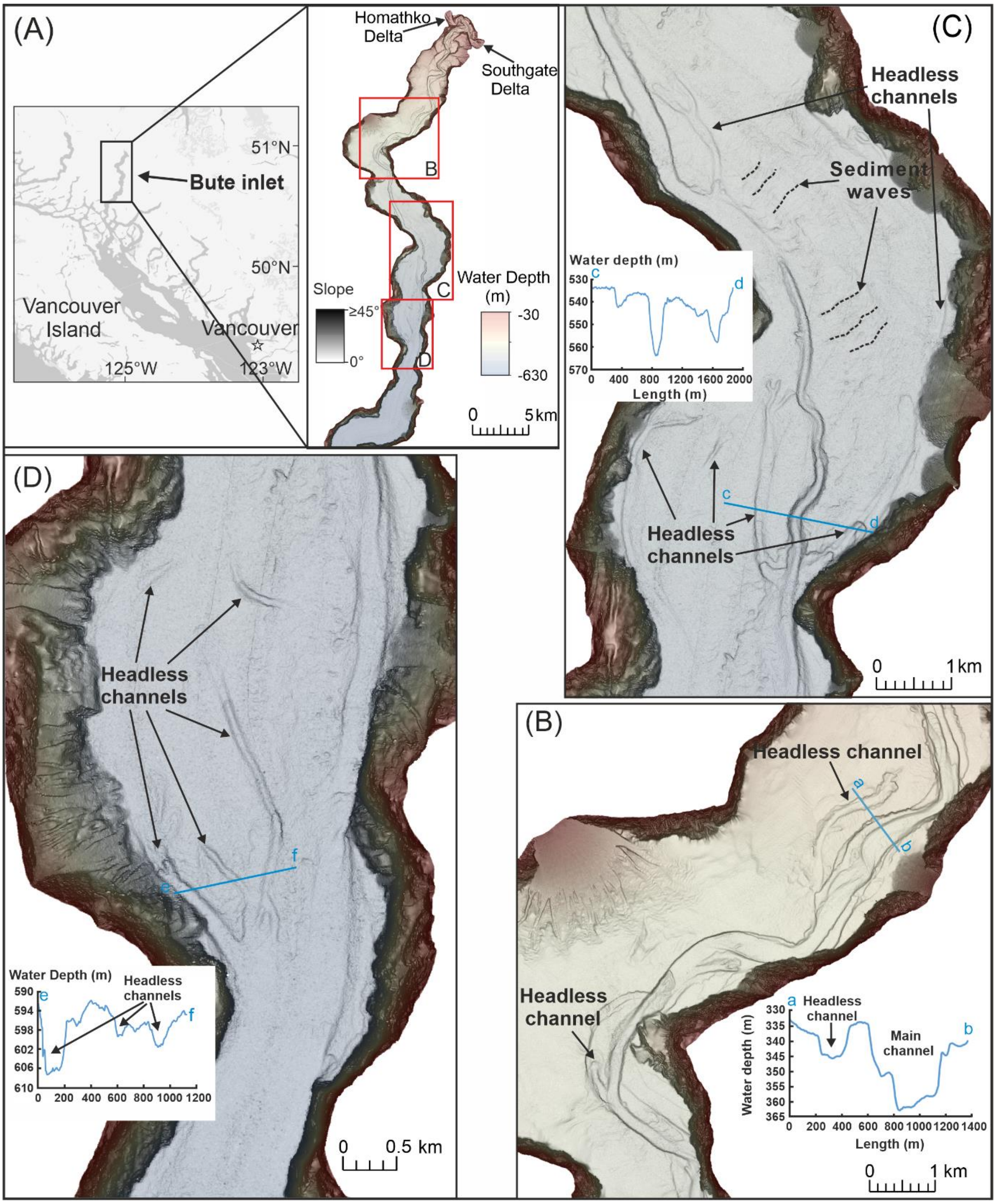

Figure 1. Submarine channel morphology and locations of headless channels within the fjord. (A) Location of Bute Inlet in British Columbia, Canada. (B-D) mark locations of headless channels and sediment waves from the middle to lower reaches of the submarine channel system. The cross-sectional profiles of headless channels and main channel are shown inset within (B) (C) and (D).

An unprecedented dataset comprising a sequence of repeat high-resolution bathymetric surveys has allowed quantification of the fjord seafloor morphodynamics from 2008 to 2018. These data were collected using either a Kongsberg-Simard EM1002 or more recently a Kongsberg EM710 multibeam system, with a vertical resolution of ~0.5% and horizontal resolution of ~3% of water depth (Heijnen et al., 2020). Data were processed using CARIS-HIPS software, with the employment of POSPac processing suite to improve the positional accuracy of the vessel using coincident data from fixed regional GPS stations. In addition to these more recent bathymetric surveys, published results from a previous field survey, which employed position interpolated sidescan sonar (Prior et al., 1986, 1987), are included in this study in order to explore the longer-term planform channel changes over an extended 30-year period. Some caution is needed in terms of the positional accuracy and the detailed interpretations of the older sidescan survey. However, comparison of identified features and broader pattern between the 1986 survey and more recent bathymetric surveys shows significant, larger-scale changes in channel location, as well as downstream channel extension.

RESULTS

The overall fjord seafloor profile exhibits a concave-upward section proximal to the deltas with a reduction in mean gradient through the middle and lower parts of the system (Fig. 2). The channel system is herein classified into five sections (Fig. 2A): (1) at the fjord-head, two freshwater delta systems clinoforms with superimposed crescentic bedforms; (2) an upper, sinuous confined channel (up to 40 m depth) that extends from the deltas to ~15 km along the channel course with a lack of headless channel features throughout the zone; (3) a middle-part, lower-gradient section that extends from ~15 km to ~29 km, with a number of headless channels on the main channel flanks; (4) a lower channel part comprising a channel-lobe transition zone (CLTZ), extending from ~29 km to 41 km, where headless channels are common; and (5) a broad distal depositional fan at the end of the channel.

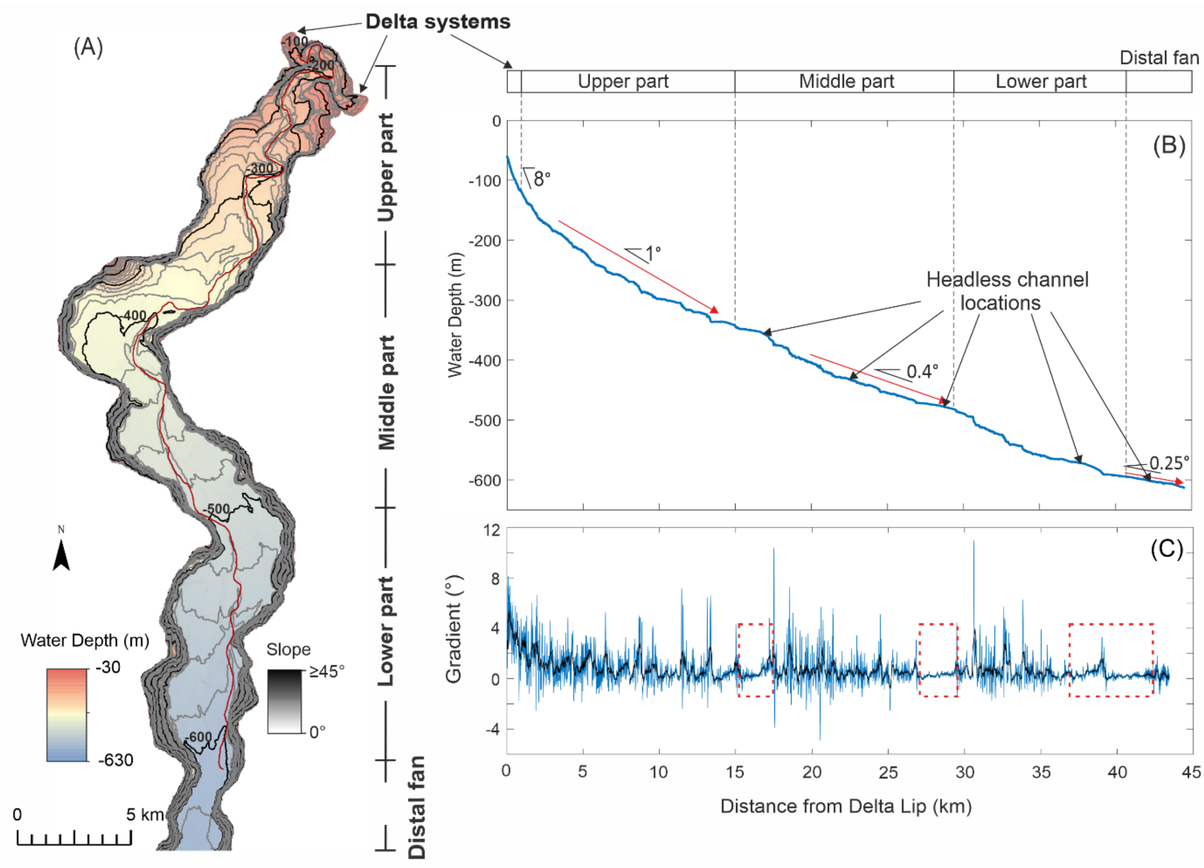

Figure 2. (A) Bathymetric map of Bute Inlet showing the broad channel classification system adopted herein. Red line denotes the channel thalweg from October 2016; (B) Profile along the channel thalweg and locations of headless channels, demonstrating that channel slope gradient decreases from ~8° at the delta to below 0.25° at the distal area; (C) Blue line denotes the gradient of channel longitudinal profile derived from each 1 m pixel. Black line denotes a moving average (10 midpoint) of slope gradient. The red dashed boxes highlight three low-gradient regions along the profile.

Headless channels are thus predominantly observed from the middle to lower part of the channel system (Fig. 1B, C, D), where slope gradients are generally below 0.5°. The submarine channel becomes less entrenched downstream and is also dominated by large knickpoints (Heijnen et al., 2020; Chen et al., 2021). Sediment waves are also present in the overbank areas along the lower part of the channel (Fig. 1C).

Comparison of the field data from 1986 to 2008 reveal clear morphological changes in the channel location within the lower channel section and across the distal fan (Fig. 3A and Fig. 3B). The 1986 sidescan survey was interpreted to show two main channels along both margins of the fjord seafloor (Fig. 3A), although there is uncertainty over how they connected in this lower resolution sidescan data (Prior et al., 1987). The bathymetric data, from 2008-2018, reveals that a small headless channel was initially formed on the central fjord seafloor but was disconnected from main channel in the 2008 survey (Fig. 3B, 3C). The evolution of this headless channel is captured by subsequent surveys showing how the feature gradually migrated upstream to connect with the main channel, and result in a single extended overall channel flow pathway (Fig. 3C). This newly-formed channel section deepened and widened from 2008 to 2018, until it was similar in cross-sectional area to the upper channel in the 2018 survey. The surveys also reveal how distributary channels are abandoned, highlighting how remnants of both distributaries can be preserved as headless channels on the seafloor (Fig. 3B). The shift of the submarine channel across the CLTZ over this 30-year period resulted from channel switching processes, with the initiation and upstream propagation of headless channels and the subsequent connection with primary channel.

DISCUSSION

Here, comparison between 1986 and 2008 surveys suggests that channel switching processes through time in the channel-lobe transition zone (CLTZ) promotes the formation of a new flow pathway and abandonment of former channels (Fig. 3A and B). Similar channel evolution processes have been observed in sublacustrine channels, in which mass transport deposits were inferred to regularly cause channel avulsions (Turmel et al., 2015; Corella et al., 2016) and in river systems, where sediment supply has been linked to avulsion frequency (e.g. Carlson et al., 2019). In Bute Inlet, the relicts of the former channels are preserved as a suite of headless channels (Fig. 3B) carved into the fjord floor. We reason, therefore, that channel switching processes likely cause the formation of submarine headless channels across the CLTZ (Fig. 4A), and that these system dynamics will be captured and ultimately preserved into the geological record (e.g. Haughton et al., 2002).

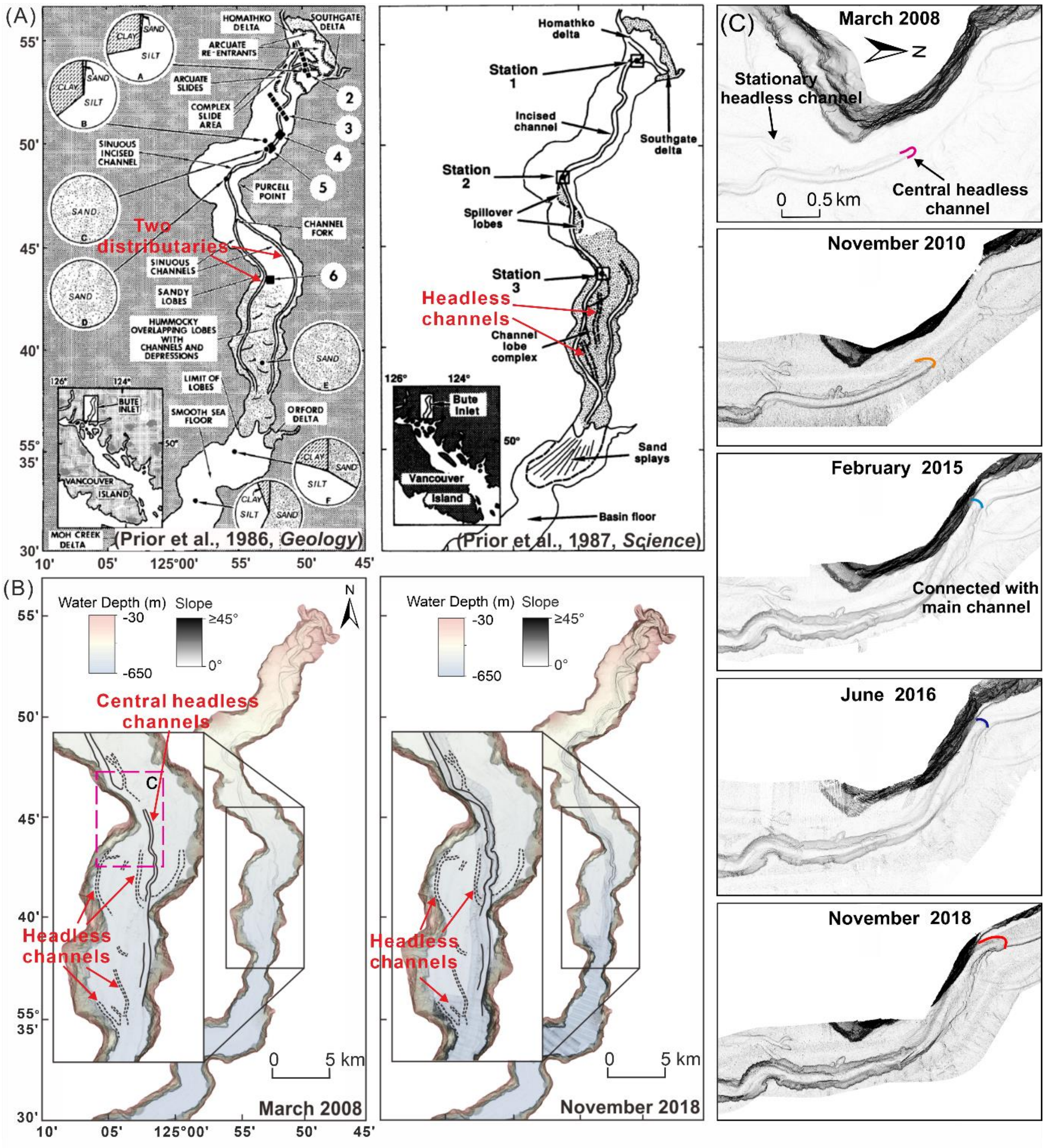

Figure 3. Morphology evolution of CLTZ in Bute Inlet. (A) Published results from1986 survey. From left to right: Prior et al., 1986; Prior et al., 1987; (B) Data in this study. From left to right: March 2008; November 2018. (C) Upstream migration of central headless channel from 2008 to 2018.

The repeat bathymetric data from 2008 to 2018 suggest that the upstream migration of erosional headless channels is associated with the loss of turbidity current confinement with the passage through CLTZ and into distal fan region. Laboratory experiments have additionally shown how loss of lateral confinement can trigger a lowering of the velocity maximum and an increase in the basal shear stress, which may also cause scours to develop (Pohl et al., 2019). It is suggested here that the initiation and subsequent propagation of headless channels seem to result from the migration of these distal erosional scours, which are broadly similar in form to some megaflutes (Hiscott et al., 2013) or the upstream migration of channel knickpoints (Heijnen et al., 2020). We reason these scour features may form incipient headless channels that subsequently migrate upstream, with these small-scale headless channels eroding up-system during turbidity current flow events. Under such circumstances headless channels may connect with the primary channel and create an extended conduit, resulting in channel extension across the distal fan (Fig. 4B, C). Other headless channels will stop migrating when turbidity currents are newly captured within the new dominant headless-channel pathway (Fig. 4B). This process results in a series of abandoned headless channels in the distal fan which will have high preservation potential as the overall channel system progrades and extends.

Headless channels can also form as a result of turbidity currents overspilling channel banks and re-channelize downslope (e.g., Normark and Piper et al., 1991). Previous analysis of the submarine channel in Bute Inlet suggests that depositional features (e.g. spill-over lobes), observed from the middle to lower reaches of the channel and occurred due to turbidity current flow stripping from the main channel (Fig. 3A, Prior et al., 1986; Zeng et al., 1991). Elsewhere, linear series of net-erosional scour-shaped depressions/steps are interpreted as incipient channels offshore California in Monterey East Channel (Fildani et al., 2013), Eel Canyon (Lamb et al., 2008), and Lucia Chica Channel (Maier et al., 2011). The interactions between turbidity current overflows and seafloor perturbations in overbank regions could therefore initiate headless channels, or form trains of scours/steps which subsequently merge to form headless channels (Fildani et al., 2013). The presence of sediment waves in the lower part of Bute Inlet (Fig. 1C) suggests that turbidity current overflow has occurred and likely contributed to the formation of headless channels when overspill flows interact with seafloor perturbations or meet confinement at fjord sides.

We propose a general model for the formation and evolution of headless channels and their interactions with the main channel, which promotes downstream channel extension via the following steps: (1) a few small headless channels are formed on the distal fan due to upstream migration of scours, driven by overpassing turbidity current becoming unconfined; (2) one of the headless channels gradually becomes established and connects with the main channel, where the initial headscarp of the headless channel may be preserved as a break in slope (i.e. a knickpoint) that continues to migrate up the main channel; (3) the new channel section is gradually deepened and widened, and develops terraces (Fig. 4B). Over time, this channel extension cycle could control both channel longitudinal profile and propagating knickpoints, and thus dominate long-term submarine channel evolution (Fig. 4C).

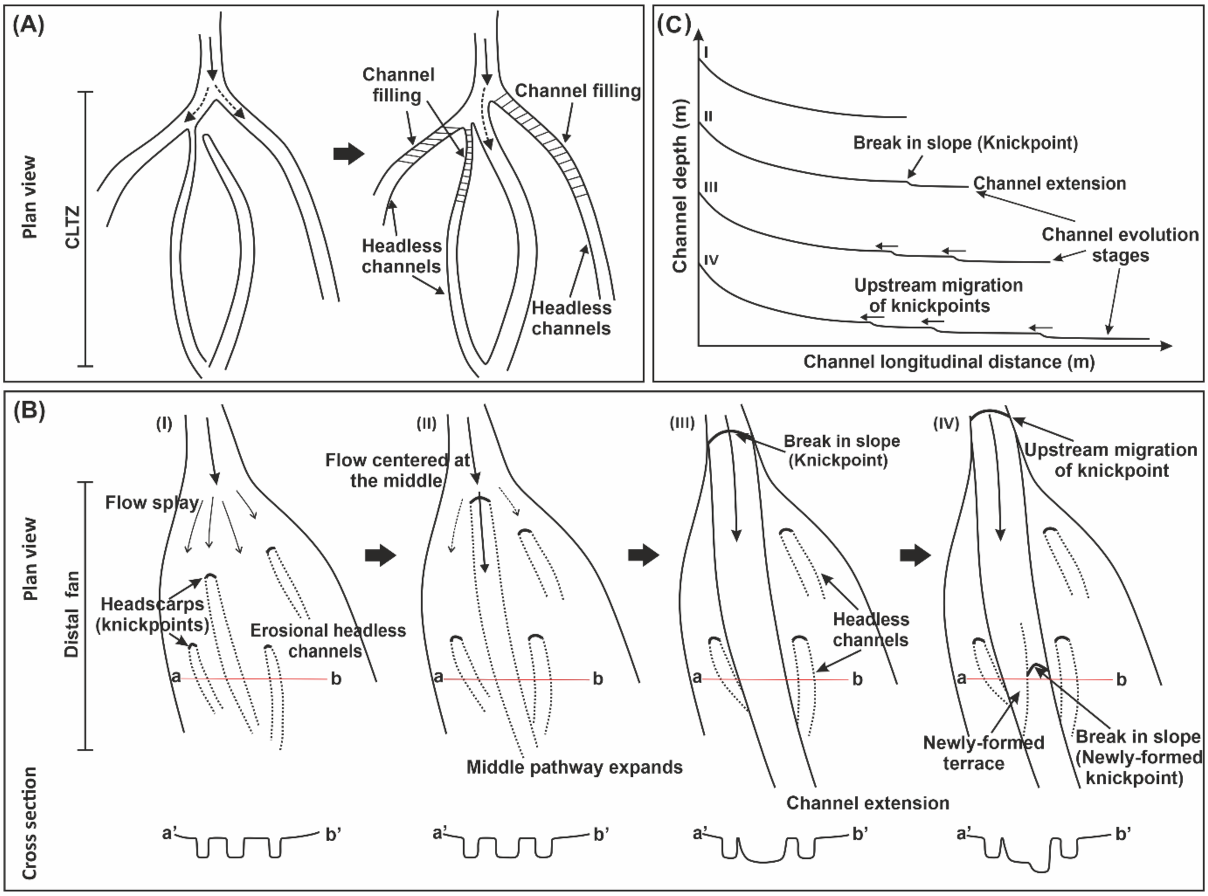

Figure 4. Conceptual models of (A) channel switching processes; (B) the temporal evolution of headless channels and interactions with main channel across the distal fan, resulting in channel extension and the formation of knickpoint; (C) the evolution of channel longitudinal profile with stages showing channel extension processes and the upstream migration of knickpoints.

Headless channels also likely record a distinct seafloor geomorphological signature within the geological record, particularly in terms of the different development stages of submarine channels (Corella et al., 2016; Gales et al., 2019). Outcrop studies have demonstrated the presence of disconnected channels and giant scours, which are interpreted as evidence of a temporal shifting of CLTZ (e.g. Hofstra et al., 2015; Brooks et al., 2018). We therefore hypothesize that the presence of a suite of headless channels may be used to help infer the locations and ancient movements of CLTZ, which in turn could be used to interpret various processes of submarine channel avulsion, channel extension and overall channel system extension. A range of scales of headless channel have been observed in a variety of submarine environments globally (e.g. Girardclos et al., 2012; Gales et al., 2019) and these features likely hold significant potential for improved interpretation of channelization and CLTZ dynamics across continental margins.

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Figure 1. Submarine channel morphology and locations of headless channels within the fjord. (A) Location of Bute Inlet in British Columbia, Canada. (B-D) mark locations of headless channels and sediment waves from the middle to lower reaches of the submarine channel system. The cross-sectional profiles of headless channels and main channel are shown inset within (B) (C) and (D).

Figure 2. (A) Bathymetric map of Bute Inlet showing the broad channel classification system adopted herein. Red line denotes the channel thalweg from October 2016; (B) Profile along the channel thalweg and locations of headless channels, demonstrating that channel slope gradient decreases from ~8° at the delta to below 0.25° at the distal area; (C) Blue line denotes the gradient of channel longitudinal profile derived from each 1 m pixel. Black line denotes a moving average (10 midpoint) of slope gradient. The red dashed boxes highlight three low-gradient regions along the profile.

Figure 3. Morphology evolution of CLTZ in Bute Inlet. (A) Published results from1986 survey. From left to right: Prior et al., 1986; Prior et al., 1987; (B) Data in this study. From left to right: March 2008; November 2018. (C) Upstream migration of central headless channel from 2008 to 2018.

Figure 4. Conceptual models of (A) channel switching processes; (B) the temporal evolution of headless channels and interactions with main channel across the distal fan, resulting in channel extension and the formation of knickpoint; (C) the evolution of channel longitudinal profile with stages showing channel extension processes and the upstream migration of knickpoints.