Clinoform growth and sediment flux into late Cenozoic Qiongdongnan shelf margin, South China Sea

Si Chen1 | Ronald Steel2 | Hua Wang1 | Rui Zhao1,2 | Cornel Olariu2

1Key Laboratory of Tectonics and Petroleum Resources, MOE, China University of Geosciences, Wuhan, China
2Jackson School of Geosciences, The University of Texas at Austin, Austin, TX, USA

Correspondence
Si Chen, Key Laboratory of Tectonics and Petroleum Resources, MOE, China University of Geosciences, Wuhan, China. Email: sichen720@hotmail.com

Abstract
The South China Sea continental margin in the Qiongdongnan Basin (QDNB) area has incrementally prograded since 10.5 Ma generating a margin sediment prism more than 4km-thick and 150–200 km wide above the well-dated T40 stratigraphic surface. Core and well log data, as well as clinoform morphology and growth patterns along 28 2D seismic reflection lines, illustrate the evolving architecture and margin morphology; through five main seismic-stratigraphic surfaces (T40, T30, T27, T20 and T0) frame 15 clinothems in the southwest that reduce over some 200 km to 8 clinoforms in the northeast. The overall margin geometry shows a remarkable change from sigmoidal, strongly progradational and aggradational in the west to weakly progradational in the east. Vertical sediment accumulation rate increased significantly across the entire margin after 2.4 Ma, with a marked increase in mud content in the succession. Furthermore, an estimate of sediment flux across successive clinoforms on each of the three selected seismic cross sections indicate an overall decrease in sediment discharge west to east, away from the Red River depocenter, as well as a decrease in the percentage of total discharge crossing the shelf break in this same direction. The QDNB Late Cenozoic continental margin growth, with its overall increased sediment flux, responded to the climate-induced, gradual cooling and falling global sea level during this icehouse period.

KEYWORDS
clinoform growth, Qiongdongnan Basin, sediment flux, shelf margin

1 | INTRODUCTION AND OBJECTIVES

Deepwater shelf margins or continental margin sedimentary prisms are important archives of clinoform architecture and evolving sedimentary regime during their construction (Chen, Steel, Olariu, & Li, 2018; Gomis-Cartesio, Poyatos-Moré, Hodgson, & Flint, 2018; Johannessen & Steel, 2005; Lin et al., 2018; Patruno & Helland-Hansen, 2018; Ryan, Helland-Hansen, Johannessen, & Steel, 2009). Specific aspects of margins have been highlighted such as (1) the geometry and sequence stratigraphy of individual high-frequency clinothems (Lobo & Ridente, 2014; Pellegrini et al., 2018), the sand-dominated (Carvajal, Steel, & Petter, 2009) vs. mud-dominated (Poyatos-Moré et al., 2016) character of the shelf break reaches of clinothems, (2) the relative effectiveness of storm waves, rivers, or tidal currents for dispersing sand onto the deepwater slope (Dixon, Steel, & Olariu, 2012), (3) the generation of sandy fan systems emplaced in clinothem bottomsets during rising and highstand of sea level (Carvajal & Steel, 2006; Covault, Normark, Romans, & Graham, 2007) and (4) forward numerical modelling of the topset width of
shelf margin clinoforms (Burgess, Steel, & Granjeon, 2008). However, what is rare or absent in clinoform margin research are studies of depositional strike variability (e.g. grain size or sediment accumulation rates) along shelf margins. Although several clinoform margin studies have documented depositional strike variability in architecture and process regime (Jones, Hodgson, & Flint, 2015; Pellegrini et al., 2018; Sanchez, Fulthorpe, & Steel, 2012), none have combined lateral changes in depositional architecture with grain size and vertical sediment accumulation rates over Myr time scales, tied to global climate and regional tectonic change.

Our main objective herein is to document an unusual example of lateral change along a Neogene segment of the South China Sea margin. Specifically, we will (1) illustrate changing clinoform architectures, and stacking patterns during the evolution of the Qiongdongnan Basin margin, (2) document a significant along-strike change in the geometry of the continental margin sediment prism, as well as a vertical increase in the sediment accumulation rate and in the mud content, especially above the Pliocene-Pleistocene boundary; and (3) provide a quantitative approach for the evaluation of sediment flux across this late Cenozoic shelf, and discuss the mismatch between the calculated high sediment flux to the shelf and the evidently small Hainan Island catchments that could possibly have fed the margin transversely. The above points will also be discussed in the context of Neogene climate cooling and increased amplitude of eustatic sea level changes.

2 | GEOLOGICAL BACKGROUND

2.1 | Geodynamic setting

The South China Sea (SCS) is a typical large marginal sea basin that opened since late Paleogene (Nagel, Granjeon, Willett, & Lin, 2015) and was located on the intersection of the Eurasian, Indian-Australian, Filipino and Pacific plates. The SCS region is composed by the Yinggehai Basin (YGHB) to the west of Hainan Island, Qiongdongnan Basin (QDNB) to the south of Hainan Island, and Pearl River Mouth Basin in the east (Figure 1a,b). The general evolution of the entire region includes a post-rift thermal subsiding stage (Early-Middle Miocene) and then an accelerated subsidence stage to the present (Wang et al., 2015; Zhao, Sun, Sun, Wang, & Sun, 2016). In the Pliocene, the YGHB and the QDNB merged and synchronously experienced relatively rapid subsidence (Figure 1a,b; Cao, Jiang, Wang, Zhang, & Sun, 2015; Clift & Sun, 2006).

The deepwater QDNB development was subdivided (Lei & Ren, 2016) into four stages: syn-rift, fault-depression, post-rift thermal subsidence and accelerated subsidence. Ren et al. (2014) subdivided the margin architecture into nine clinoform intervals. He, Xie, Kneller, Wang, and Li (2013) documented the unidirectional migration of some submarine canyons, attributed to the activity of an outer shelf warm current. Su et al. (2014) demonstrated a segmentation of the axial submarine canyon, developed since late Miocene (Figure 1a,b), with a turbidite channel complex in the western head, and a mass transport complex in the east, derived from the northern continental slope. The paleo-Red River was considered to be the main source passing across the YGHB and feeding the submarine canyon. Xie, Müller, Ren, Jiang, and Zhang (2008) described the prograding slope clinoforms

### FIGURE 1
(a) Modern topographic continental shelf margin of QDNB, SCS with the shelf edge location of T40, T30, T27, T20 and T0 (Gong et al., 2016; Ren et al., 2014; Shi et al., 2013; Sun et al., 2015; Zhao et al., 2016, 2015). AA’, BB’ and CC’ mark the 3 illustrated seismic profiles out of a group of 28 parallel seismic lines available to this study. Note the positions of the shelf break during the study interval and the widening of the shelf especially from early Pliocene (T27). Published lithology data from well YC35-1–2 is used in this study. (b) River systems from southern China and Vietnam (Cao et al., 2015; Milliman & Farnsworth, 2011). QDNB = Qiongdongnan Basin, YGHB = Yinggehai Basin; Pearl River Mouth Basin lies immediately east of QDNB. Yellow line in QDNB is the trace of the deepwater submarine canyon active since late Miocene.

### Highlights
- There is a dramatic lateral, along-strike change in growth pattern from west to east in architecture of the Late Miocene-Pleistocene QDNB shelf margin.
- Time change in clinoform geometry shows initially strong shoreline progradational patterns across a low-angle ramp morphology became aggradational, and eventually, strongly rising trajectory during the margin growth.
- There was a dramatic increase in sediment accumulation rate after 2.4 Ma. Less than half of the succession was deposited within the 8 My before the Pleistocene whereas more than half accumulated within the subsequent 2.4 My.
- During the Pleistocene, there was a marked increase in mud content in the succession likely caused by long-distance, alongshore currents with high content of mud.
- Laterally, the total sediment flux onto the margin shows a dramatic decline from west to east while moving away from the Red River depocenter.
(a) Hainan Island

Legend

Survey boundary
Seismic lines
Shelf edges
Wells
Central Canyon

(b) Red River Fault Zone

Elevation (m)

1000-3000
500-1000
100-500
<100

Nanliujiang
Thai Binh
Gulf of Tonkin
Song Hong (Red River)
Ma
Nam San
Ca

YGHB
QDNB

Nandujiang
Changhuajiang

Qiongzhou Strait
Gulf of Tonkin

Red River Fault Zone

Nanliujiang
Thai Binh
Gulf of Tonkin
Song Hong (Red River)
Ma
Nam San
Ca

YGHB
QDNB

Nandujiang
Changhuajiang

Qiongzhou Strait
Gulf of Tonkin
and the rapid seaward shift of shelf edges of YGHB, suggesting that this was due to high sediment supply, compared to the adjacent aggradational stacking pattern in QDNB, where there was insufficient sediment supply. Gong, Steel, Wang, Lin, and Olariu (2016) focused on the sediment flux to the deepwater parts of this basin. Previous research has therefore shown that the overall Neogene QDNB shelf margin development in time was characterized as progradational, progradational/aggradational and then aggradational, providing the basis for a discussion of relationship between sediment flux and dynamic stratigraphy.

2.2 | Age of the main stratigraphic surfaces on the margin

A critical part of the analysis herein, is the subdivision of the thick continental margin succession into stratal units that can be correlated and dated along-strike on the margin, through the 28 available depositional dip seismic lines (Figure 1a). A main stratal framework from continental shelf to the deepwater QDNB basin floor has long been known via a series of key surfaces whose ages have been determined and published by earlier researchers (Lei et al., 2011; Sun, Wu, Lü, & Yuan, 2010; Xie et al., 2008). To be consistent with regional correlations, the present study recognizes and uses these reflection surfaces in the re-examined seismic dataset of the current project. However, we note some disagreement on age determination among previous researchers. The ages of the key surfaces are based on a nannofossil biostratigraphy, for example, planktonic foraminifera, calcareous nanoplankton and dinoflagellates, collected from more than 20 CNOOC borehole drilling sites in the YGHB and QDNB (Lei et al., 2011; Xie et al., 2008). The key stratigraphic surfaces have usually been identified by stratal unconformities or correlative conformities as seen by stratal truncation, downlap and onlap of seismic reflectors. The surfaces utilized herein are T40, T30, T27 and T20, and these are widely used and mapped in previous works in the QDNB (Figure 2, Figures A and B in Supplementary material) (Sun et al., 2010). The unconformity surface T40 (10.5 Ma) (Xie et al., 2008) is considered to be the time of important change to a passive margin stage in the basin, though the thermal subsiding stage started earlier at about 23 Ma (Li et al., 2017; Wang et al., 2015). Some other researchers estimate the age of surface T40 as 11.6 Ma (e.g., Lei & Ren, 2016; Wang et al., 2015). The central submarine canyon system that runs sub-parallel along the base of slope in the deepwater area (yellow line in QDNB, Figure 1a) developed from T30 (5.5 Ma) (Sun et al., 2010; Yuan et al., 2009) to T28 (3.8 Ma). Some other researchers estimate T30 age as 5.6Ma (e.g. Lin et al., 2001) or 5.7Ma (e.g. Su et al., 2014). Within the overlying succession, the main surfaces are T27 (2.4 Ma) and T20 (1.9 Ma) (Sun et al., 2010; Yuan et al., 2009) to Present. Some researchers have estimated T27 to 2.7Ma (Li et al., 2017; Su et al., 2014), T20 as 1.6Ma (Liu et al., 2015; Wang et al., 2015) or 1.8Ma (Lin et al., 2001; Shi, Xie, Wang, Li, & Tong, 2013; Su et al., 2014).

The above dating is important to the present project, but most of the disagreements are relatively minor in the longer time scale of the current research.

2.3 | Main morphological features of the basinal area

The NE–SW oriented QDNB is a petroleum basin characterized by syn-rift structure in the lower part, and post-rift broad subsidence in the upper part of the sedimentary infill, the latter being the margin in focus for the present work. The QDNB has a length of 250–450 km NE–SW, a width of 150–200 km NW–SE and a total area of 8 × 10⁴ km² (Su et al., 2014). The primary morphological feature imaged on Figure 2, from earliest Pliocene to present, is the shelf break, separating the landward extending shelf from the continental slope. The QDNB shelf extended for an unknown distance back to the northwest and the outermost 100km of this shelf is well seen on Figure 2. The margin clinoforms create a significant vertical relief (between shelf break and basin floor), up to 2 km since the early Pliocene, though the continental slope is irregular in places. The slope segment of any clinothem becomes gently dipping as it approaches the basin floor. It is noticeable that the oldest clinoforms, developed between T40 and T30, have a much lower amplitude and are more strongly progradational than the younger ones (Figure 2), giving the impression that the earliest stage of margin growth was in a ramp rather than a shelf break setting. A prominent feature of the basin floor is the NNE-SSW oriented ‘central submarine canyon’, a deeply incised feature of early Pliocene age (T30) on the basin floor, running nearly parallel with the toe of slope (Figure 1a).

The QDNB clinoforms are, by scale and setting, clearly continental margin clinoforms in the classification of Patruno and Helland-Hansen (2018), and are the key growth increments of the larger scale SCS continental margin. However, it should be noted that the QDNB clinoforms are still several hundred km away from the ocean–continent boundary to the southeast and there is some significant rising (but deep subaqueous) topography, probably pre-Late Miocene rifted remnants, lying between the high-relief clinoforms and the oceanic crust. Some of this topographic complexity can be seen to the southeast of the clinoformed margin in Figure 2, Figures A and B in Supplementary material.

2.4 | Potential sediment supply to the margin

The SCS margin was fed by a large drainage system with runoff greater than 0.43 m/year (Milliman & Farnsworth, 2011; Milliman & Syvitski, 1992; Syvitski & Milliman, 2007). The potential river sources for QDNB Margin are Ca
River (suspended sediment (TSS) and dissolved-solid (TDS) are 4 Mt/year and 2.7 Mt/year), Ma River (TSS = 3 Mt/year; TDS = 2.2 Mt/year), Nam San River, Song Hong (Red) River (TSS = 50–110 Mt/year; TDS = 20 Mt/year), Thai Binh River (TSS = 1 Mt/year; TDS = 1.9 Mt/year), which run into YGHB from Vietnam; and Nanliujiang River (TSS = 1.1 Mt/year) from south coast of China runs into Gulf of Tonkin, as well as the two rivers from Hainan Island (Changhuajiang river (TSS = 0.08 Mt/year) runs towards northwest into YGHB and Nanduijiang River (TSS = 1.1 Mt/year) runs into northern Qiongzhou strait) (Milliman & Farnsworth, 2011) (Figure 1b).

3 | SEISMIC DATABASE

The database includes 28 depositional dip oriented (NW–SE) seismic lines and 23 along-strike (NE–SW) lines (all lying between AA′ and CC′ in Figure 1a). Only three main seismic lines (Line AA′, BB′, CC′ in Figure 1a) are used in the discussions below and in the sediment flux calculation, because the method used is a 2D-based wave equation calculation giving a result for each of the seismic profiles. Petter, Steel, Mohrig, Kim, and Carvajal (2013) tested this method successfully with different numbers of laterally coeval seismic
lines and on different margins where the sediment discharge had been independently calculated by other, more time-consuming methods. The advantage of this method, compared for example with the method outlined by Carvajal and Steel (2012), is that no 3D survey is required.

The studied stratigraphic interval (T40-present, duration ca. 10.5 My) is subdivided into 15 clinothems in Line AA’, 12 in Line BB’ and 8 in Line CC’. The clinoforms have been interpreted by mapping the strong amplitude reflectors on dip oriented cross sections between the well-defined key surfaces. The interpreted surfaces were followed from the gently dipping reflectors (on the shelf), down into the more steeply dipping reflectors (slope) and eventually the less steep reflectors on the basin floor. The key surfaces T40, T30, T27 and T20 were mapped and recognized across the 28 dip lines in the whole QDNB volume and tied to well logs. However, some clinothems pinch out laterally across the 160 km strike distance, and so the numbering of clinothems above and below the key stratigraphic surfaces in Figure 2, A, and B varies slightly. This lateral variability is unimportant for the sediment flux calculation, as each seismic line is considered separately. A seismic velocity of 1.54 m/ms has been used in the 2D seismic profile time-to-depth conversion (1500 m/s in Gong et al., 2016; 2000 m/s in Han, Leng, & Wang, 2016; P-wave velocities are 1.7–4.5 km/s in the sedimentary layer in Qiu et al., 2013; Zhao et al., 2015).

4 | STRATIGRAPHIC ARCHITECTURE OF QDNB CONTINENTAL MARGIN

4.1 | Clinoforms, stacking patterns and shelf edge trajectories

The continental margin clinoforms of QDNB have been building basinward since 10.5 Ma, generating a sedimentary prism more than 4 km-thick above the T40 surface (Figure 2). The clinoforms are well imaged by the reflectors on the seismic profiles (Figure 2, Figures A and B in Supplementary material). As with most shelf margins, the clinothems stack basinwards in an incremental manner, with fairly flat ‘tramline’ reflectors in the topset deposits suggesting repeated progradation–retrogradation transits of shorelines during shelf growth (Steel, Olariu, Zhang, & Chen, 2019 this volume). The landward extent of transgressions on the shelf is difficult to evaluate because of a lack of lithology information. The more steeply dipping segments of clinoform reflectors denote the deepwater continental slope and the outermost, gently dipping parts of clinoforms represent the basin-floor deposits. The growth trajectory made by the shelf break during the entire growth period of the margin is somewhat irregular, and shows several flat (progradation) and rising (aggradation) segments. Within the three chosen seismic inlines (Figure 2, Figures A and B in Supplementary material), the well-known T40 surface marks the beginning of an accelerated subsidence stage, separating the overlying passive continental shelf margin from underlying post-rift interval of thermal subsidence (Wang et al., 2015; Zhao et al., 2016).

The shelf break points for each clinothem were picked where the change of gradient on the outermost shelf reaches a maximum value (Olariu & Steel, 2009; Patruno, Hampson, Jackson, & Dreyer, 2015). Likewise the toe of slope is located where the gradient change from the more flat-lying bottomset up to the slope reaches a maximum value (Anell & Midtkandal, 2017; Anell, Midtkandal, & Braathen, 2014; Pirmez, Pratson, & Steckler, 1998).

On Line AA’, the shelf edge trajectory begins to rise significantly at T30 (Figure 2), that is, the T30 surface marks the end of the ‘ramp’ period with strongly prograding clinoforms, after which there is a well-developed shelf break with rising trajectory pattern. Above the T30 surface, shelf aggradation increased markedly. The toe of slope trajectory on line AA’ shows some irregularity around T20 and T30. The shelf edge trajectory angle increases significantly after clinothem 12 (Figure 2), and this allows the overall shelf margin in the west to be subdivided into a progradation stage with average shelf edge trajectory angle of −0.7° (clinothems 1–4), a progradation + aggradation stage with average shelf edge trajectory angle of 2.2° (clinothems 5–11), and an aggradation stage with average shelf edge trajectory angle of 4.7° (clinothems 12–15) (Figure 2).

For Line BB’, the shelf edge trajectory shows two ‘flat-to-rise’ intervals, with the most progradational interval just above the T40 surface in clinothem 1, and another below the T20 in clinothem 8 (Figure A in Supplementary material), exhibited as two low-angle segments of the shelf edge trajectory (Figure A(B) in Supplementary material). The average shelf edge trajectory angle is −2.4° for clinothem 1, 6.6° for clinothems 2–7, 0.7° for clinothem 8 and 7.9° for clinothems 9–12.

For Line CC’, the entire margin and shelf edge trajectory shows an aggradational pattern with very small sediment volumes on the deepwater slope (Figure B in Supplementary material). The average shelf edge trajectory is very steep (clinothems 1–8) with an angle of 70°.

The T40-T30 stratal interval (Figure 2) is a transitional stage to the overlying passive margin, and this strongly prograding (modestly aggrading) interval defines the early ‘ramp’ setting, as noted above. This early ‘ramp’ stage of margin clinoform development has been described also from other deepwater basins (e.g. Santra, Steel, Olariu, & Sweet, 2013). A series of channelized incisions on the T40 surface are discussed further below, as well as the longitudinal submarine canyon system developed from T30 to T28. The canyon system has a W–E orientation, a length of about 425 km and width of 3–12 km as described by Gong and others (2015).
4.2 | Clinoform types

The clinoform shapes have been defined by the changes in dominant thickness along any clinothem (Figures C and D in Supplementary material) and this can change along-strike. They can be classified as: F-type with thick basin-floor deposits but relatively thin deposits on the shelf and deepwater slope; S-type with relatively thick slope deposits but thinner deposits on both shelf and basin-floor segments; and O-type with thick shelf deposits but relatively thin deposits on the slope and basin floor (see Chen et al., 2018).

On the seismic Line AA’, the distribution is F-type (on clinothems 1, 2, 5 and 11), S-type (clinothems 7, 8, 10 and 12–15) and O-type (clinothems 3, 4, 6 and 9) (Figure C in Supplementary material). Through time, the clinothems changed from dominantly F-type low in the stratigraphy, to O-type later and then to S-type towards the top (Figure D in Supplementary material). The slope gradient of T40 surface is 2.6°. For the seismic Line BB’, the slope gradient of T40 surface is 3.4°. Below the level of T27, the dominant clinothem shape is F-type with a relatively high partitioning of sediment beyond shelf edge. For the seismic Line CC’, the slope gradient of T40 surface is 3.7°. The biggest volumes of sediment are preserved either on the shelf or the basin floor. O-type becomes the dominant type in line CC’ due to the extreme aggradational style of the margin there.

The T40 slope gradient increases from line AA’ towards CC’ in the east. The clinoform types match with the sediment partitioning on the shelf margin (Figure E in Supplementary material). The F-type associates with a strong partitioning of sediments towards deepwater area, associated with basinward shifting and progradational clinoform style (blue dot in Figure E(A and B) in Supplementary material). The O-type reflects on-shelf storage of sediments (low shelf edge flux percentage, pink dot in Figure E(A in Supplementary material) and/or a high aggradational pattern (pink dot in Figure E(C) in Supplementary material). The S-type usually suggests moderate sediment flux beyond the shelf edge compared to the other types, consistent with transitional stage between aggradational and progradational stacking patterns (green and orange dot in Figure E in Supplementary material).

4.3 | Depositional systems interpretation from seismic data

The accreting study clinothems are especially clear above T30; each of them has topset (morphological shelf), deepwater slope and basin-floor segments (Figure 2, Figures A and B in Supplementary material). The topset reaches of the margin are interpreted to contain distributary channels, swamps and flood-basin deposits in the proximal northerly areas, and shallow-marine delta, estuary, shoreface and shelf deposits in the distal parts of topsets (Li et al., 2017; Sun et al., 2014). Only the distributary channels are obvious on the seismic data (Figure 3) as discussed below. Coeval with the topset deposits, the more steeply dipping slope segments of the clinoforms are relatively muddy, but also channelized, and dominated by sediment gravity flow deposits; the basin-floor segments are also likely to be muddy or show sandy, turbiditic submarine-fan development (Cao et al., 2015; Li et al., 2017; Liu et al., 2015; Su et al., 2014; Wang et al., 2015). During the time interval T40 to T30 there was strong progradation of the deltaic systems, less distinct shelf breaks and large volumes of sandy sediment were bypassed along the ramp margin from the shelf and into the basin-floor area at this early stage of margin growth. The sediment rich ‘ramp’ connected with a V shaped canyon near the base of slope (Figures 1a and 2).

On seismic line AA’, the seismic reflections show strong amplitudes above T20 both on-shelf and basin-floor segments. There are some discontinuous strong reflections, probably distributary channels on the shelf above T27 (Figure 3a) and some coeval chaotic seismic events on the deepwater slope (Figure 3a), interpreted as mass transport complexes. The central canyon near the toe of slope at level T30 cuts down 400 m, has a width of 6 km, developed in W–E orientation (Figure 1a), and contains channels filled with turbidites and mass transport deposits, which have been previously interpreted by well lithology, log curves and seismic reflections (well log and seismic data in Lin et al., 2001; Gong et al., 2015; seismic data in He et al., 2013) and can be seen on all three seismic lines.

On seismic line BB’, there are also some discontinuous strong reflectors suggesting channels on the shelf above surface T27 and coeval chaotic reflections on the upper slope, interpreted as slumps and slides near areas of decreased slope gradient (Figure 3b, Figure A in Supplementary material). The obvious U-shaped downcutting reflector at the basin floor near the T30 surface, the central canyon, has a width of about 7.5 km and depth of 700 m, which shows multiple downcutting and infill phases within the canyon container (Figure 4a).

On seismic line CC’, there are again small, U-shaped downcutting channel reflectors (about 4.5 km in width 200 m in depth) on the outer shelf between T40 and T30, and some discontinuous strong reflections on the inner shelf above T27. There are also very chaotic strata on the narrow upper slope (Figure 3c). Some wedge and lenticular-shaped units overlap with each other on the lower slope and basin floor with their landward reaches on-lapping the toe of slope, and their basinward part pinching out (Figure B in Supplementary material). The large-scale incised canyon downcutting on the basin floor near T30 shows multiple channelized surfaces within its U-shaped container (about 10 km in width and 900 m in depth) and possible levee system flanking on both sides of the canyon (Figure 4b).
4.4  |  Lateral changes along the continental margin

Along the depositional strike direction (SW–NE) in the study region, there is a spectacular change from southwest to northeast in the margin architecture, shelf width and vertical sediment accumulation rate. This change is likely related to the western location of the main sediment supply fairway, the Red River, while the eastern reaches of the system represent the eastern flank of this system. A prominent feature is that the margin growth in the west is sigmoidal with strong progradation and aggradation (Figure 5AA’), whereas the clinoformal slope in the east becomes steeper and shorter (Figure 5CC’).

The total stratal thickness of the post-Miocene succession decreases fairly abruptly from SW to NE (Figure 5). The shelf width also decreases from west to east (Figure 1a), and eventually disappears towards farther east from CC’. This also results in a change of shelf edge orientation, from W–E during T40-T27, to SW–NE orientation above T20. The average vertical sediment accumulation rate (T40-present) decreased from 366 m/My in AA’ to 223 m/My in BB’, to 151 m/My in
The margin progradation rate also decreased laterally to the east as the margin transformed from a strongly prograding to an aggrading margin (Figure 5, Table 1).

The late Miocene stage of QDNB margin growth (C1-4 in Line AA′, Figure 5) represents an Icehouse, moderate to high sediment supply system, and the regressive shelf deltas are likely to have been accommodation dominated deltas (sensu Porebski & Steel, 2006). A reversal in fault motion along the strike-slip Red River Fault (Location in Figure 1b) accompanied by rapid subsidence along this zone occurred at approximately 5.5 Ma (T30) at the boundary of Miocene and Pliocene (Li et al., 2017). It has been suggested that this seismicity may have triggered large-scale gravity currents to enter the QDNB to aid in the formation of the submarine canyon and basin-floor fans (Su et al., 2014; Zhao et al., 2016). The porosity loss from the mud occurred mainly above a depth of 2–2.5 km and was negligible below 4 km (Zhao et al., 2016).

4.5 Vertical changes in sediment accumulation rate along depositional strike

The Pliocene to Pleistocene stratigraphic boundary T27 (2.4 Ma), just above clinothem 6 in line AA′, clinothem 5 in line BB′ and clinothem 4 in CC′ (Figure 2, Figures A and B in Supplementary material), is of particular significance for the following reasons:

1. The time interval T40 (10.5 Ma) to T27 (2.4 Ma) in seismic line AA′ can be seen (Figure 6) to represent about one-third of the total thickness of the studied margin prism at this location, and has a time duration of about 8.1 My. The time interval T27 to present, however, represents about two-thirds of the sediment prism thickness at this location (Figure 2) but has a duration of only 2.4 My, taking into consideration the compaction-depth curve (Athy, 1930). The initial porosity of mud and sand were estimated to be 69% and 43%, respectively, with greater and faster compaction of mud than sand (Zhao et al., 2016). The porosity loss from the mud occurred mainly above a depth of 2–2.5 km and was negligible below 4 km (Zhao et al., 2016).

2. There was a demonstrably high sediment accumulation rate above T27 surface. The average sediment accumulation rate from surface T40 to T27 is calculated to be 216 m/My, 103 m/My and 104 m/My (Table 1) respectively along the location of seismic sections AA′, BB′ and CC′, whereas the accumulation rate from T27 to present rises to 875 m/My, 628 m/My and 310 m/My, respectively.

These calculations suggest that the QDNB margin accommodation and sediment accumulation rate increased by a factor of 5 in the western area and by a factor of more than 3 in the east, across this Pliocene-Pleistocene boundary (T27). Note in Figure 6 that this change in sediment accumulation rate corresponds with a significant increase in the cooling of global climate and with a significant increase in the amplitude and frequency of the rises and falls of eustatic sea level (Clift, 2006; Clift & Sun, 2006; Clift, Wan, & Blusztajn, 2014; Miller et al., 2005; Van Hoang et al., 2010; Zhao et al., 2015).

In addition, there is another significant parameter that changes through time on this margin, namely the dominant grain size of the sediment deposited on the outer shelf segment of the margin, during the Late Pleistocene as indicated by the well log of YC35-1-2 (depth 1,500 m-300 m on YC35-1-2 in Figure F and Figure 2). The integrated lithology column in Figure 6 shows that compared to the relatively sand-rich environment during the late Miocene (from about 10 Ma) there is a marked reduction of sand-sized sediment deposited, and this is especially prominent in the entirely mud prone Pleistocene succession from 2.4 Ma. We note that the lower half of the Pleistocene interval (in the well of Figure F) has penetrated the slope, which of course is likely to be muddy. However, the upper half of the Pleistocene thickness in the well (Figure F) is clearly penetrating the outer shelf, which is not normally entirely muddy (though possibly muddy locally between channels). This change of grain size is therefore considered to be real, as was also shown in more detail by Gong et al. (2016, their Figure 17).
In addition to the seismic and stratigraphic approaches to understanding the variability of the QDNB margin above, we also need a quantitative estimate of the sediment discharged onto the margin. We use two methods for sediment flux calculation to compare the results of estimated and predicted sediment flux; that of Petter et al. (2013) based on the use of a series of seismic lines, the other by Syvitski and Milliman (2007), the BQART model. The first is based on use of the kinematic wave equation and geometric parameters across a series of 2D cliniforms, with the result expressed as 2D sediment flux in a format of rate m²/year (Petter et al., 2013), whereas the second method is predictive, using the drainage basin parameters of catchment area, hinterland relief, climate and catchment rock types (BQART), with the term sediment discharge expressed as million tons per year (MT/year) (Syvitski & Milliman, 2007). The sediment volumes calculated from the seismic data have also been used to estimate the catchment area. The key geometric parameters for the cliniforms are listed in Figure G(D) in Supplementary material.

### FIGURE 5
The strong, along depositional strike, lateral change (SW to NE) in stratal architecture of the QDNB continental margin, as seen by the changes in architecture between seismic lines AA’, BB’ and CC’ (see seismic lines in Figure 2, Figures A and B in Supplementary material).

### TABLE 1
Progradation, aggradation and accumulation rate of the QDNB margin

<table>
<thead>
<tr>
<th>Clinoform Profiles</th>
<th>Progradation Rate $R_p$ (m/My)</th>
<th>Aggradation Rate $R_a$ (m/My)</th>
<th>Accumulation Rate $R_i$ (m/My)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>T40-T0</td>
<td>T40-T30</td>
<td>T30-T27</td>
</tr>
<tr>
<td>AA’</td>
<td>8973</td>
<td>9324</td>
<td>3418</td>
</tr>
<tr>
<td>BB’</td>
<td>3655</td>
<td>3773</td>
<td>650</td>
</tr>
<tr>
<td>CC’</td>
<td>92</td>
<td>92</td>
<td>92</td>
</tr>
</tbody>
</table>

| Clinoform Profiles | Maximum | Average | Maximum | Average | Maximum | Average | Maximum | Average | Maximum | Average | Maximum | Average | Maximum | Average | Average |
|--------------------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|
|                    | T40-T0  | T40-T30 | T30-T27 | T27-T20 | T20-T0  | T40-T0  | T40-T30 | T30-T27 | T27-T20 | T20-T0  | T40-T0  | T40-T30 | T30-T27 | T27-T20 | T20-T0  |
| AA’                | 514     | 366     | 285     | 216     | 1445    | 875     | 220     | 186     | 1490    | 691     | 514     | 366     | 285     | 216     | 1445    |
| BB’                | 390     | 223     | 156     | 103     | 1222    | 628     | 98      | 111     | 1160    | 492     | 390     | 223     | 156     | 103     | 1222    |
| CC’                | 280     | 151     | 175     | 104     | 652     | 310     | 95      | 121     | 713     | 216     | 280     | 151     | 175     | 104     | 652     |

5 | SEDIMENT DISCHARGE ESTIMATES FROM ANCIENT MARGINS

In addition to the seismic and stratigraphic approaches to understanding the variability of the QDNB margin above, we also need a quantitative estimate of the sediment discharged onto the margin. We use two methods for sediment flux calculation to compare the results of estimated and predicted sediment flux; that of Petter et al. (2013) based on the use of a series of seismic lines, the other by Syvitski and Milliman (2007), the BQART model. The first is based on use of the kinematic wave equation and geometric parameters across a series of 2D cliniforms, with the result expressed as 2D sediment flux in a format of rate m²/year (Petter et al., 2013), whereas the second method is predictive, using the drainage basin parameters of catchment area, hinterland relief, climate and catchment rock types (BQART), with the term sediment discharge expressed as million tons per year (MT/year) (Syvitski & Milliman, 2007). The sediment volumes calculated from the seismic data have also been used to estimate the catchment area. The key geometric parameters for the cliniforms are listed in Figure G(D) in Supplementary material.
5.1 Petter et al. (2013) workflow

1. Interpret the individual clinothems along each 2D seismic line. 2. Each clinothem is subdivided into shelf, slope and basin-floor segments, as already discussed above. 3. Perform a time-to-depth conversion for each seismic line, to obtain the clinothem framework in a depth domain. 4. Calculate the clinoform height along each of the seismic profiles. Note that decompaction is not applied in this study, the calculated result represents compacted thickness. 5. Calculate the 2D area of each clinothem as well as the areas of shelf, slope and basin floor of individual clinothems. 6. Calculate the sediment flux across each clinothem for each 2D line (Figure G in Supplementary material) by using the Petter et al. (2013) method. Using the kinematic wave equation, the changes in sediment flux in space (x) across the clinothems can be calculated using clinothem progradation and aggradation rate (P and A) in equations 1 and 2 (Figure G(C) in Supplementary material). Integral of equation 2 gives the sediment flux at any point x along the clinothem length L. The sediment flux at any point x can be obtained by equation 5, where t is interval time. The thickness trends of clinothems can be obtained from equation 6, and the pinchout distance can be resolved from equation 7. The corresponding clinothem height at the pinchout point can be obtained in equation 8. The horizontal distance from the shelf break to the clinothem pinchout point and the discrepancy in elevation of the clinothem can be seen from equations 9 and 10. Using equation 5, the sediment flux at the shelf edge can be solved. Similarly, the horizontal distance from the shelf edge to the inland pinchout point and the discrepancy in elevation of the clinothem can be solved by equations 11 and 12. 7. The 2D sediment flux (unit: m³/year) at the shelf edge can be calculated. According to the grain density (unit: g/cm³) and sedimentary system width (unit: km), the sediment supply and sand transport volume can be calculated (unit: MT/year).

5.2 Uncertainty and potential error in calculation

1. The use of the kinematic wave equation is based on the notion that the clinoform is in equilibrium at geological time scales. The hypothesis is that sediment is both preserved and notional that the clinoform is in equilibrium at geological time scales. The hypothesis is that sediment is both preserved and

FIGURE 6 (a) Time and clinoform framework of the stratal succession for QDNB margin (Shi et al., 2013). (b) Integrated lithology column of the QDNB margin (Shi et al., 2013), which is a regional composite vertical lithology. (c) and (d) The sedimentary succession is calibrated to the coeval global sea level changes (Miller et al., 2005) and temperature curve changes (Zachos et al., 2001) since Late Miocene. (e) Clastic mass accumulation rates of erosional flux from Pearl River Mouth Basin in northern SCS. MAR = mass accumulation rate (Clift, 2006; Clift et al., 2014). (f) Accumulation rate of QNDB, YGHB and Red River (Clift & Sun, 2006; Van Hoang et al., 2010; Zhao et al., 2015)
associated with the estimation of porosity, coefficient of consolidation, and the pinchout point trendline. For the burial and compaction errors, the present day clinothem thickness is decreased from its original thickness with a porosity decay that changes exponentially with depth due to compaction (Athy, 1930; Petter et al., 2013), which decreases about 10% (sandstones) and <5% (mudstones) from original porosities (40%–50%) (Petter et al., 2013; Sclater & Christie, 1980). Limited compaction of around 12% for the Plio-Pleistocene successions was also reported in Anell & Midtkandal, 2017. The influence of compaction has been partly counteracted by uplift and tilt of the strata. There are errors associated with erosion during deposition or post-depositionally, as well as with subsidence deformation, stratal tilting influenced by compaction and tectonic movements (Henriksen, Hampson, Helland-Hansen, Johannessen, & Steel, 2009; Zhang, Steel, & Ambrose, 2016).

5.3  | Sediment flux results using the Petter method
The sediment flux estimates, during margin growth in the past 10.5 My, changed through time and laterally between the seismic lines. The average 2D sediment flux across clinothems is estimated at 121 m²/year in Line AA’, 109 m²/year in Line BB’ and 93 m²/year in Line CC’, that is, a decreasing flux from the west to east (Table A in Supplementary material). Assuming the sedimentary system width is 80 km between each seismic line, and sediment density is 2.65 g/cm³, the average sediment discharge is estimated to be 26 Mt/year for clinothems in Line AA’, 23 Mt/year in Line BB’ and 20 Mt/year in Line CC’ (column sediment discharge in Table A in Supplementary material). Sediment flux percentages passing across the shelf edge are shown as Qs percentage in Table A and Figure E in Supplementary material, and this is defined as the percentage of the total sediment flux exported off the shelf.

5.4  | Calculation of sediment discharge using BQART Model: Methodology
Since the QDN Basin is located in a region of unglaciated river systems in the tropics and lower mid-latitude, the BQART equation (Syvitski & Milliman, 2007) was used to provide a second estimate of the sediment flux (see also Blum & Hattier-Womack, 2009).

The BQART method (temperature condition ≥2°C) allows evaluation of the relationship between the drainage area, sediment load and relief. For the BQART Equations (13 and 14) (Syvitski & Milliman, 2007; Zhang et al., 2018, 2016), \( Q_s \) represents for sediment load (Mt/year); \( A \) is basin catchment area (km²); \( R \) (km) represents maximum relief for drainage basin; \( T \) is basin average temperature in °C; constant \( \omega = 0.02 \) (kg/s) or 0.0006 (Mt/year); Mean water discharge \( Q \) is in km³/year; parameter \( B = 1.0 \) eliminates the influence of glaciers, basin wide lithology, lake and reservoir trapping, and human-influence soil erosion factor (Blum & Hattier-Womack, 2009; Somme, Piper, Deptuck, & Helland-Hansen, 2011; Syvitski & Milliman, 2007).

Other researchers have found that the BQART equation reasonably predicts Holocene sediment flux (Milliken, Anderson, Simms, & Blum, 2017), and is reliable for long-term average determinations (Milliken et al., 2017; Syvitski & Milliman, 2007) (Figure G(C&D) in Supplementary material). The possible errors of the BQART relationship applied to ancient systems, as has been discussed by Zhang et al. (2018), includes uncertainties in climate input and environmental conditions, and hierarchy of time scales (Zhang et al., 2018). There is also uncertainty from the original values of the parameter data collected from literature.

5.5  | Sediment flux results from BQART Model
Application of the BQART model results in a \( Q_s \) estimate of 6.4–14.1 Mt/year for temperature between 15–30 °C (Table B in Supplementary material). The estimated discharge based on sediment volume is about 7.1–18.2 Mt/year. The BQART estimates are less than those calculated by the Petter method (20–26 Mt/year) using three 2D seismic lines.

6  | DISCUSSION
6.1  | Global geological events and responses in the Qiongdongnan margin
Figure 6 provides an overview of global eustatic sea level, temperature changes and accumulation rate discussed for the T41-present QDNB shelf margin evolution (Clift, 2006; Clift & Sun, 2006; Clift et al., 2014; Miller et al., 2005; Van Hoang et al., 2010; Zachos, Pagani, Sloan, Thomas, & Billups, 2001; Zhao et al., 2015). The tectonic and sedimentary environment changes at the T40 surface reflect the beginning of rapid subsidence into the passive continental margin. Above surface T40, there was a significant increase in the amplitude of eustatic sea level rises and falls, from about 9.2 Ma. The large incised canyon at the T30 surface is possibly the result of several marked eustatic sea level falls just below the T30 surface. This was also a time of marked temperature decrease, falling below 0 degree centigrade at around 5 Ma. T28 (3.8 Ma) (Shi et al., 2013; Wang et al., 2015; Xie et al., 2008; Zhao et al., 2016) marks the initial point of increasing sediment accumulation rate in QDNB. Analog from late Pleistocene Po River shows that 350 m thick succession deposited within 17 ky driven by eustatic and climatic changes (Pellegrini et al., 2018, 2017).
6.2 | Along-strike change and asymmetry of the margin architecture

This along-strike change and asymmetry of the margin architecture is something fairly rarely reported, but is consistent with a strong sediment supply coming out across the Red River shelf. Interestingly, the asymmetry also implies a river domination for the protruding margin shape, as a strong wave climate in the region (e.g., seasonally varying wave condition linked with the monsoonal climate) with significant long-shore drift of sediment along the margin would have caused the above changes to be less dramatic.

6.3 | The cause of high accumulation rates and mudiness of the Pliocene-Pleistocene succession

The multiple factors which may have caused the high accumulation rates and mudiness towards the top of the study succession, include:

① Accelerated subsidence: In Asia, tectonic movements have been proposed as a dominant cause of the observed variations in Cenozoic erosional and depositional systems (Jiang et al., 2017; Métivier, Gaudemer, Tapponnier, & Klein, 1999). The accelerated subsidence of the QDN shelf margin may have been associated with isostatic uplift and increased erosion in the catchment areas, resulting in increased mud prone sediment flux. The uplift of the Tibetan Plateau possibly relates to the dextral strike-slip movement along the Red River fault zone since 5.5 Ma (Cao et al., 2015; Sun et al., 2014), which expedited the accelerated subsidence process in QDNB, and spread large volumes of sediment towards YGHB and QDNB (Li et al., 2017).

② Mud sources: Usually, the decrease in grain size (e.g., from 1.9 Ma) would link with reduced sediment supply, retreat of the coarser sediment and would indicate that the earlier topography was becoming worn down (i.e. the end of an earlier era of tectonic uplift). However, this is contradicted by the increased accumulation rates here in QDNB margin. Thus the increased mud requires an extraordinary source of large volumes of mud. We did already note that the lower half of the Pleistocene succession penetrated in well 35-1-2 was deepwater slope sediment and would automatically be mud prone, but this still leaves the anomalously muddy outer shelf in the Upper Pleistocene to be explained. The Late Pleistocene muds are unlikely to have accumulated by hyperpycnal and hypopycnal plumes from the adjacent Hainan river mouths because of the low discharges from these rivers and the very high accumulation rates of the mud. The small river discharge from Hainan Island, that is, 1.18 Mt/year from Changhuajiang (with area of 5.1 × 10^3 km^2 and 220 km in length) and Nandujiang (with area of 6.6 × 10^3 km^2) Rivers (Milliman & Farnsworth, 2011), is insufficient to support significant growth of the QDNB margin. On the contrary, a long-distance transport of mud by nearshore coastal currents is well known from many shelves, e.g., Amazon-Orinoco shelf or Adriatic shelf (Blum & Hattier-Womack, 2009; Nittouer, Kuehl, Sternberg, Figueiredo, & Faria, 1995; Pellegrini et al., 2015; Peng, Steel, Rossi, & Olariu, 2018; Ridente & Trincardi, 2005).

One possibility is a muddy supply from the Red River system, whose modern discharge is 133 Mt/year, and average sediment delivery since 1.8 Ma is 42 ± 8 Mt/year (Clift, 2006). Strong erosion within catchment areas would have increased sediment delivery to YGHB and raised the clastic influx into the basins since 4 Ma (Clift, 2006; Clift & Sun, 2006; Van Hoang et al., 2010). Muds exiting from large river mouths often follow a complex set of pathways (Wright & Nittouer, 1995). A buoyant plume of mud transport by shore-parallel currents could have inevitably brought large volumes of mud across the YGHB for hundreds of kilometres to deposit in the QDNB. During this transit, the sandy portion of the sediments would have been unloaded first in the YGHB. The depocenters in YGHB have migrated southeastward through time (Sun, Zhou, Zhong, Zeng, & Wu, 2003). Another possible analogue is the mud wedge that extends 800 km from the Yangtze River mouth southward along the East China Sea coast into Taiwan Strait and beyond (Liu et al., 2006, 2007).

Another possibility is that local mafic volcanism in the QDNB (Cao et al., 2015) could have provided sediments from submarine volcano and intrusive and extrusive igneous rocks dispersed from Hainan Island (Cao et al., 2015), especially the late Miocene-recent igneous rocks in northern Hainan island (Yan, Deng, Liu, Zhang, & Jiang, 2006). Regional tectonic events also show basaltic volcanism in central and southern Indochina since 11.6 Ma, in Leizhou Peninsula since 5.5 Ma, in northern Hainan since 1.8 Ma (Cao et al., 2015). The magmatism was vigorous in the Pliocene-Recent in the Leizhou Peninsula and the Quaternary in northern Hainan Island (Flower, Zhang, Chen, Tu, & Xie, 1992; Ho, Chen, & Jiang, 2000) but quiet or very weak before the Pliocene (6 Ma) (Yan et al., 2006). The Cenozoic volcanism on northern Hainan island was most extensive in the Pleistocene (Shi et al., 2011). Thin layers (less than 5 cm) of volcanic ash have been reported at ODP site 1,148 (600 km east of Hainan Island), which dated to younger than 1 Ma (Yan et al., 2006). Large amounts of volcanic ash could easily have exceeded the sediment discharge from the river-deltaic system. For example, the long-term, time averaged incision rates are estimated as 0.05–0.1 mm/year from volcanic ashes in the Colorado River (Blum & Hattier-Womack, 2009; Dethier, 2001).

③ Monsoonal driven fluid mud dispersal: A regional erosional response to Tibetan uplift and to strong East Asian monsoon rains during the global cooling stage may have caused increased sediment discharge (Clift et al., 2014;
Zhao et al., 2015) with high content of mud. Thick shelf mud deposits (Figure F in Supplementary material) are commonly associated with mud belts that are driven by littoral waves along the innermost Amazon shelf (Peng et al., 2018) or downdrift of Po Delta in the Adriatic Sea (Cattaneo, Trincardi, Asioli, & Correggiari, 2007). Summer monsoon intensification drove periods of faster erosion after 3–4 Ma (Van Hoang et al., 2010). The results of Late Quaternary East Asian monsoon intensity could have been expressed through current transport of fluid mud (Liu et al., 2010), especially from the Red River, by alongshore littoral or shelf currents. This hypothesis is discussed in more detail by Zhao et al. (2019).

7 CONCLUSIONS

We document the sediment flux, clinoform growth and the unusual lateral architectural change along a segment of the Late Cenozoic SCS continental margin. Key points include:

1. Lateral, along-strike change in architecture of the shelf margin: The architecture of Late Miocene-Pleistocene QDNB continental margin shows a dramatic growth change from west to east. There is a marked eastward decrease in margin progradation rate, in shelf width and in the thickness of individual clinothem increments of growth.

2. Time change of ramp to clinoform geometry: The early margin growth of the QDNB shows strong shoreline progradational patterns across a low-angle ramp morphology (i.e. no marked shelf slope break), but this became aggradational and a clinoforming shelf break margin developed, eventually with a late stage, strongly rising trajectory.

3. Accumulation rate: There was a dramatic increase in sediment accumulation rate after 2.4 Ma. Less than half the thickness of the total succession was deposited in the 8 My before the Pleistocene whereas more than half accumulated within the subsequent 2.4 My.

4. Grain size: During the earliest stages of QDNB margin growth there was a sand-rich sediment supply and highly progradational margin architecture, connecting directly with the central canyon transport system. During the Pleistocene there was a marked increase of mud content in the succession likely caused by long-distance, alongshore currents with high content of mud.

5. Sediment flux: Laterally, the total sediment flux onto the QDNB margin shows a dramatic decline from west to east, that is, away from the Red River depocenter. Vertically, the estimates of sediment flux across the eastern segments of the clinoformed margin indicate a decreasing percentage of sediment transported beyond the shelf edge through the time.

6. Sea level change: QDNB Late Cenozoic shelf margin growth and overall increased sediment flux responded to global, high-frequency transgressive-regressive climate cycles during an overall falling global sea level and decreasing temperature during this icehouse period.

ACKNOWLEDGEMENTS

This project was supported by National Natural Science Foundation of China (Grant No. 41702114) and Fundamental Research Funds for the Central Universities, China University of Geosciences (Wuhan) (No. CUG170616). We have strongly benefited from the constructive and thoughtful comments of reviewers David Hodgson, Stefano Patruno, Claudio Pellegrini and Changsong Lin.

DATA AVAILABILITY STATEMENT

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

ORCID

Si Chen [i] https://orcid.org/0000-0002-0305-8641
Ronald Steel [i] https://orcid.org/0000-0002-6846-7788
Hua Wang [i] https://orcid.org/0000-0003-2750-6258
Rui Zhao [i] https://orcid.org/0000-0003-2882-3810
Claudio Pellegrini and Changsong Lin.

REFERENCES


SUPPORTING INFORMATION

Additional supporting information may be found online in the Supporting Information section at the end of the article.

How to cite this article: Chen S, Steel R, Wang H, Zhao R, Olariu C. Clinoform growth and sediment flux into late Cenozoic Qiongdongnan shelf margin, South China Sea. Basin Res. 2019;00:1–18. https://doi.org/10.1111/bre.12400