A Geomorphological Investigation of Submerged Depositional Features within the Outer Silver Pit, Southern North Sea.

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Abstract
Two elongate ridges with significant bathymetric expression are observable at the eastern extremity of the Outer Silver Pit depression in the Southern North Sea. Using a 3D seismic dataset, supplemented with 2D seismic lines, bathymetry and Quaternary/Seabed geology maps, a comprehensive description of the features and their locality was gained enabling a broad geomorphic investigation of the ridges. It is suggested that the features are moribund, early Holocene, sand banks that were probably formed in an estuarine environment. It is consequently implied that the Outer Silver Pit had been a macro-tidal, tidally dominated estuary during the early Holocene which had relatively strong tidal currents operating in it.

Key Words
Outer Silver Pit, Geomorphology, sand banks, Sea-level change.

Introduction
A distinct E-W trending bathymetric deep, the Outer Silver Pit (OSP), lies at approximately 54N 2˚E on the bed of the North Sea (Figure 1). This deep is the largest of a series of offshore depressions in the Southern North Sea that are thought to have formed during Quaternary glaciations, either as the product of subglacial processes (e.g. Valentin, 1957; Robinson, 1968; Balson and Jeffery, 1991; Praeg, 2003) or catastrophic drainage events in an ice marginal environment (Wingfield, 1990). Alternatively, Donovan (1965) postulated that strong tidal currents in the Southern North Sea during the early Holocene marine transgression were responsible for eroding such deeps.

Whilst there has been considerable research into the formation of the Outer Silver Pit, the localised geomorphology within the depression has been largely ignored despite its potential to provide further insights into of the processes that led to the formation of the depression. The availability of 3D seismic data to this study provided an ideal dataset for the investigation of the morphological and stratigraphical characteristics of the OSP. This report will demonstrate that 3D seismic can be used to identify a number of distinctive geomorphological features within the Outer Silver Pit, the most prominent of these being two elongate ridges that also have bathymetric expression (Figure 2). These features will form the focus of this report. Through close examination of their morphology and locality, and by detailed comparison to modern bedforms with similar morphology, it is the aim of this study to classify the elongate ridge features and extract any information about the environments and conditions in which they were formed.

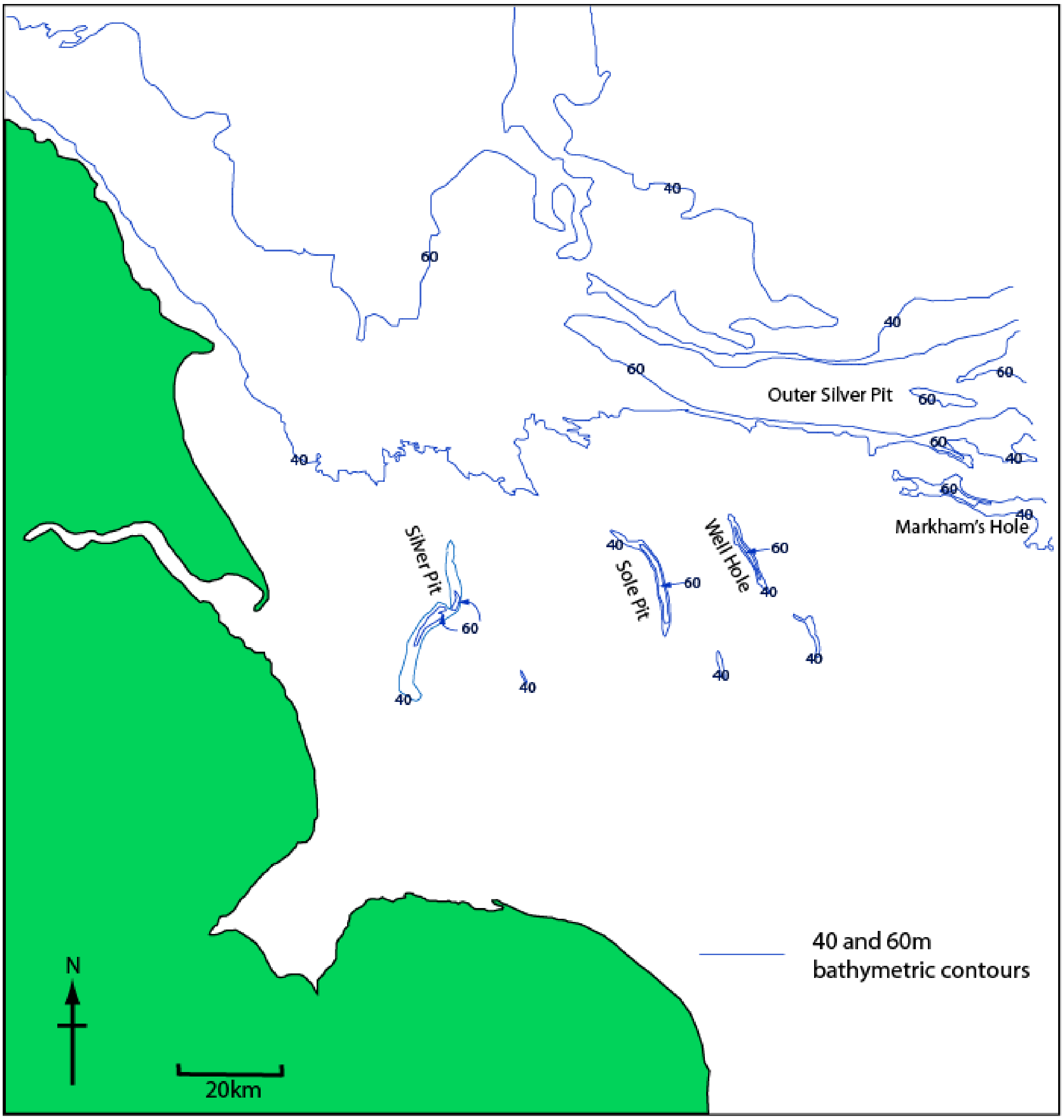

Figure 1: 40 and 60m bathymetric contours of the Southern North Sea. The major bathymetric depressions are labelled.

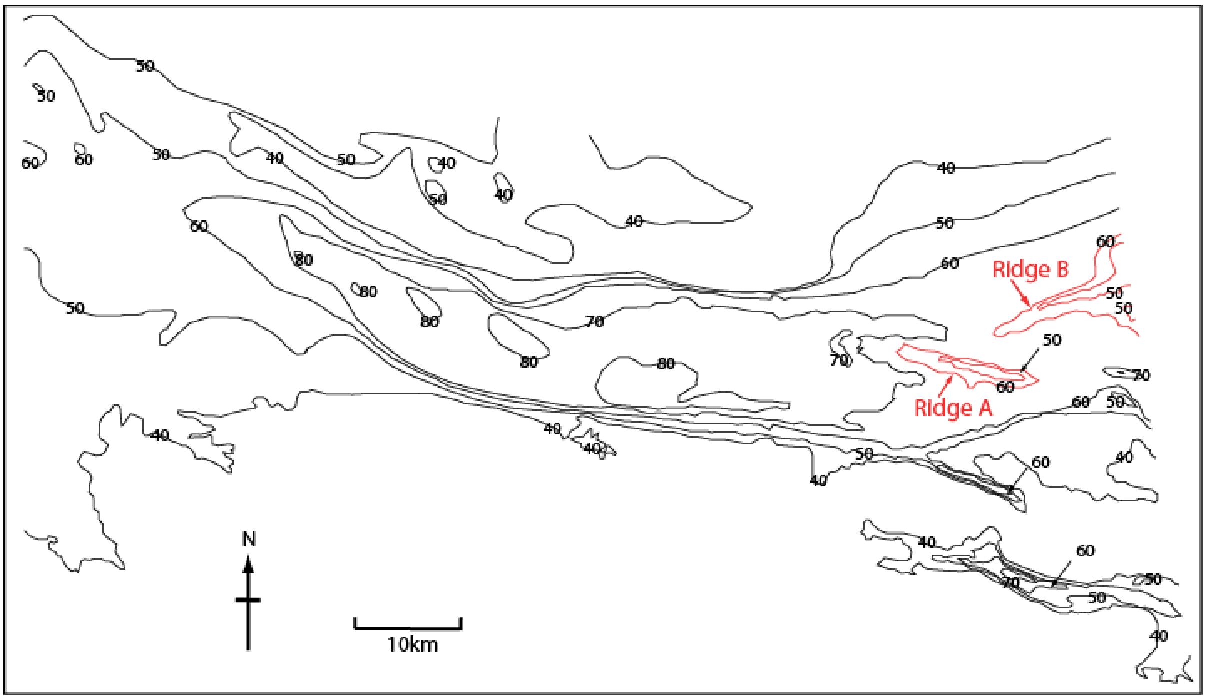

Figure 2: Bathymetric contours (40, 50, 60, 70 and 80m below sea level) of the Outer Silver Pit area; depicted in red are the elongate ridge features.

Feature descriptions

Figures 3 and 4 show the two elongate ridges within the OSP (Figure 2). Further details of the locations, dimensions and trends of the ridges are contained in Table 1 and Figure 5. The ridges are situated at the eastern limit of the Outer Silver Pit (Figure 2) and stratigraphically they lie at, or very close to, the seabed, approximately 40-50m below sea level at their crests (Figure 3). Ridge A is elongate and tapers towards the ESE but is rounded towards the WNW end. It is discretely located within the OSP, entirely disconnected from the banks of the depression. Although tapered at its WSW end, Ridge B broadens in an ENE direction until it connects to the bank of the OSP at its eastern extremity

	Ridge A	*Ridge B*
Length	~18km	~16.5km
Width	~2km	~3km
Height	~30m	~20m
Trend	ESE/WNW	WNW/ENE

Table 1: Position, dimensions and trends of Ridges A and B

In cross section, Ridge A displays an asymmetry along its major axis and both ridges are asymmetric along their minor axes (Figures 3 and 4); the asymmetry of the minor axis of Ridge A decreases towards its ESE end. The steeper slopes of the ridges are slightly concave whereas the shallower slopes are predominantly convex (Figure 3). The dip of the ridges' flanks are relatively shallow (Table 2 and Figure 5), reaching an approximate maximum of only 1.13on the steeper slope of Ridge B. Given that the trends of the ridges are offset the steeper slope of Ridge A faces ESE whereas the steep flank of Ridge B faces WSW (Figures 3 and 4). Cross sectional views of the features (Figure 3) also reveal that the crests of the ridges are relatively flat.

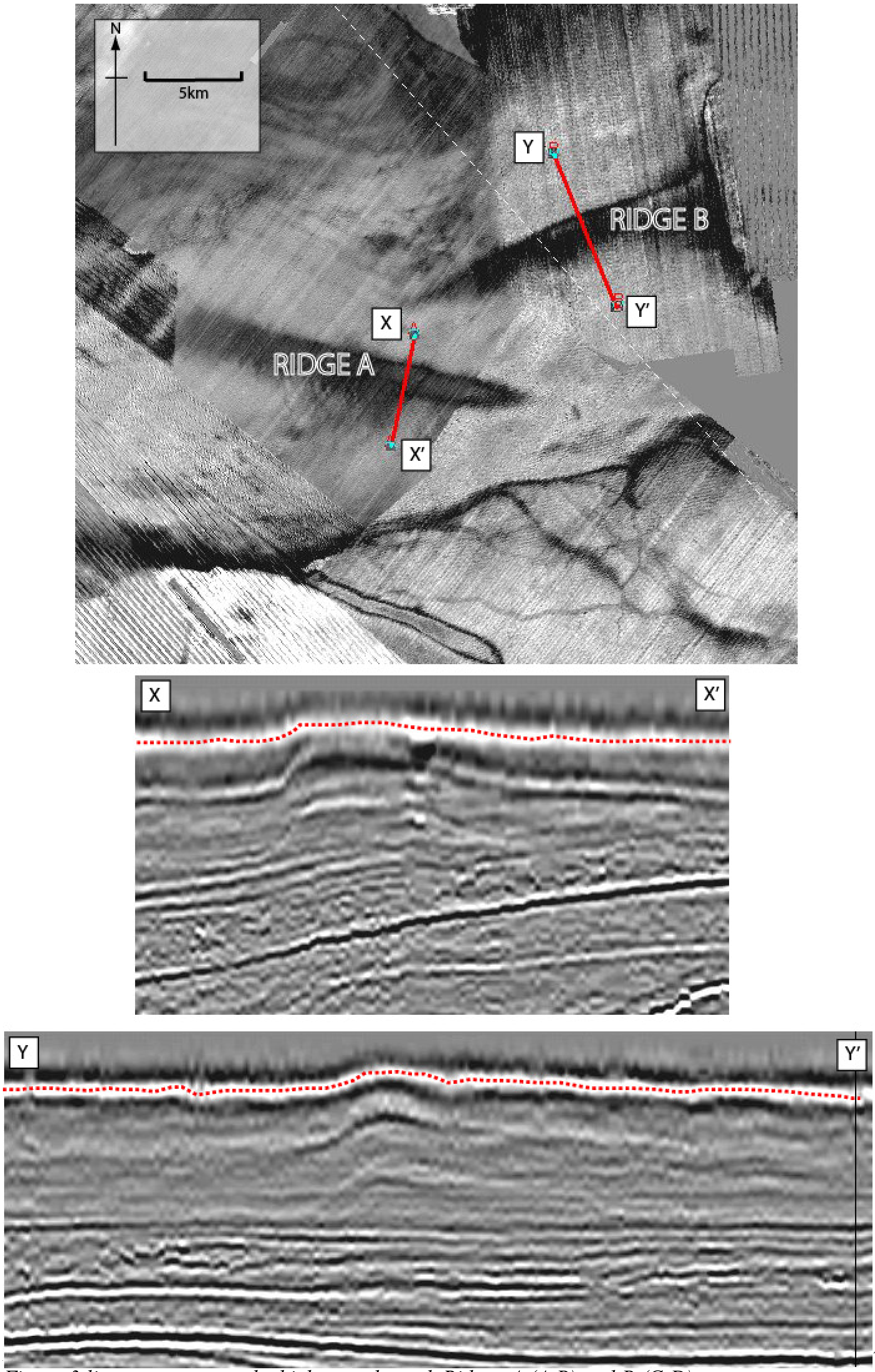

Figure 3 lines are annotated which pass through Ridges A (A-B) and B (C-D).

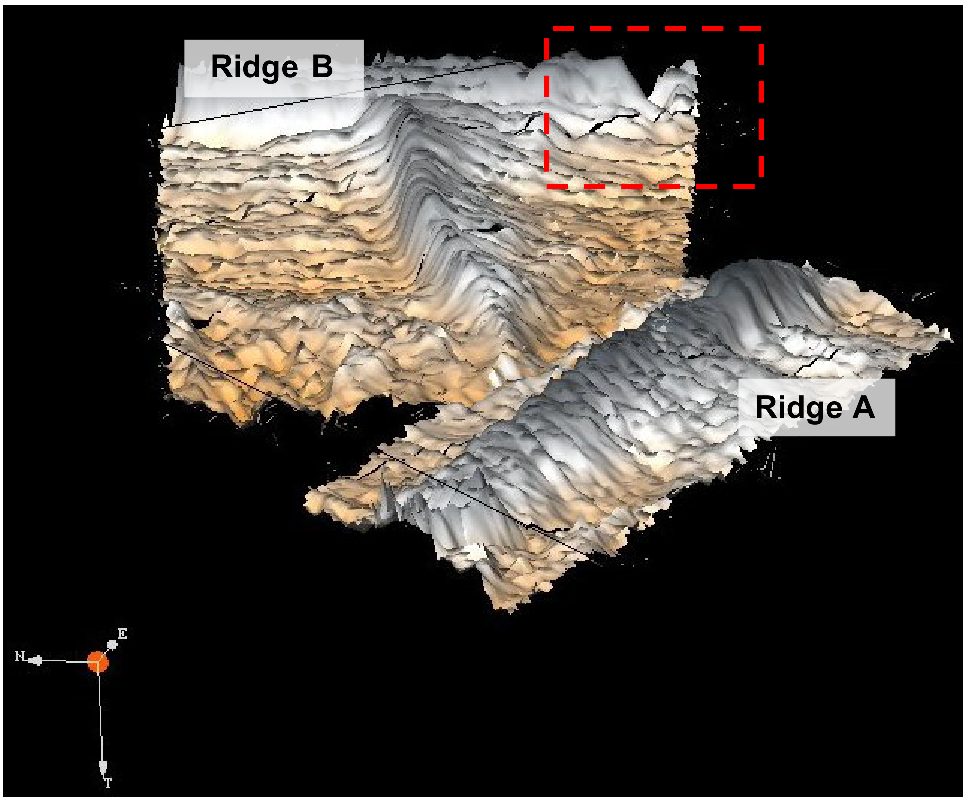

Figure 4: 3-D illuminated view of Ridges A and B, with a depression referred to in the text highlighted in red. The ridges are vertically exaggerated.

Table 2: Average dip of the flanks of Ridges A and B.

	Ridge A		*Ridge B*	
	Shallow Slope	*Steep Slope*	*Shallow Slope*	*Steep Slope*
Major Axis	0.07°	0.1°	0.09°	
Minor Axis	0.035°	0.84°	0.035°	1.13°

Two available 2D seismic lines (Figure 6), which pass through Ridge A, provide further details not apparent in the lower resolution 3D data. Firstly, the surface of ridge is smooth on both the steep and shallow slopes. Secondly, the ridge lies on a surface that undulates relatively sharply, truncates the strata below and is seen throughout most of the OSP, where it is covered by a relatively thin veneer of sediments. Finally, the 2D data shows that the internal structure of Ridge A is composed of a series of dipping internal reflectors (foresets).

On a broader scale, the ridges are located at the confluence of several features. At the eastern end of the OSP, and to the sides of Ridge B, the depression forks into two lesser branches (Figures 2, 5 and 7). The branch to the north of Ridge B trends in an ENE direction and is approximately 7.5km wide with a shallow northern flank. The second branch, to the south of Ridge B, trends in a SW direction and is approximately 5.5km wide with relatively steep banks on both sides. It is not possible to constrain the length of the either branch as they continue beyond the confines of the data. Also, to the immediate south of Ridge B lie two lesser depressions measuring approximately 2km and 0.75km in width (Figures 4 and 7) that emerge from the raised ground to the east of the OSP. The Quaternary Geology map of the area reveals that these small depressions correspond with outcrops of the Botney Cut Formation

(Figure 7), a system of partially or completely infilled subglacial valleys dated to the late Weichselian (Larminie, 1989).

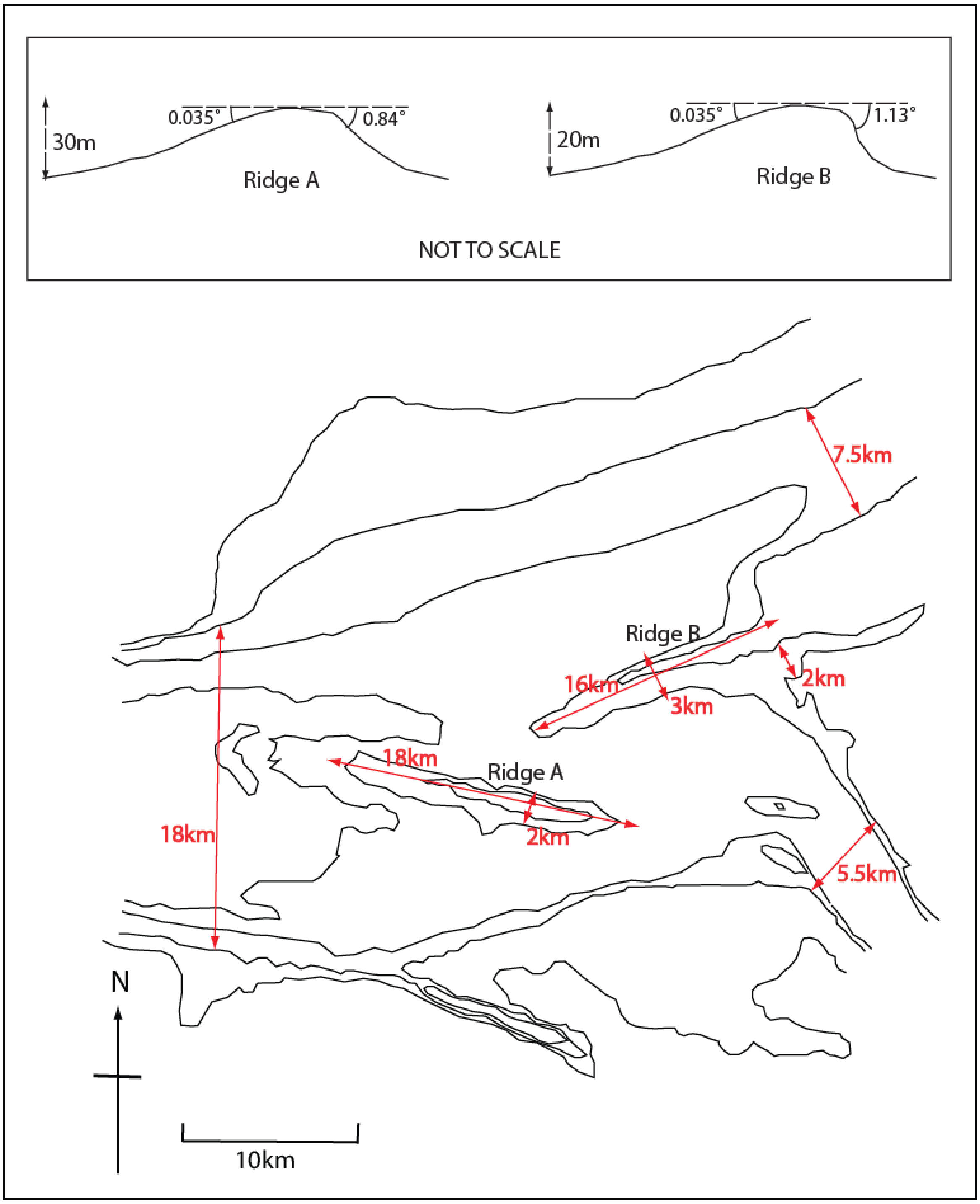

Figure 5: Dimensions of the OSP and Ridges A and B.

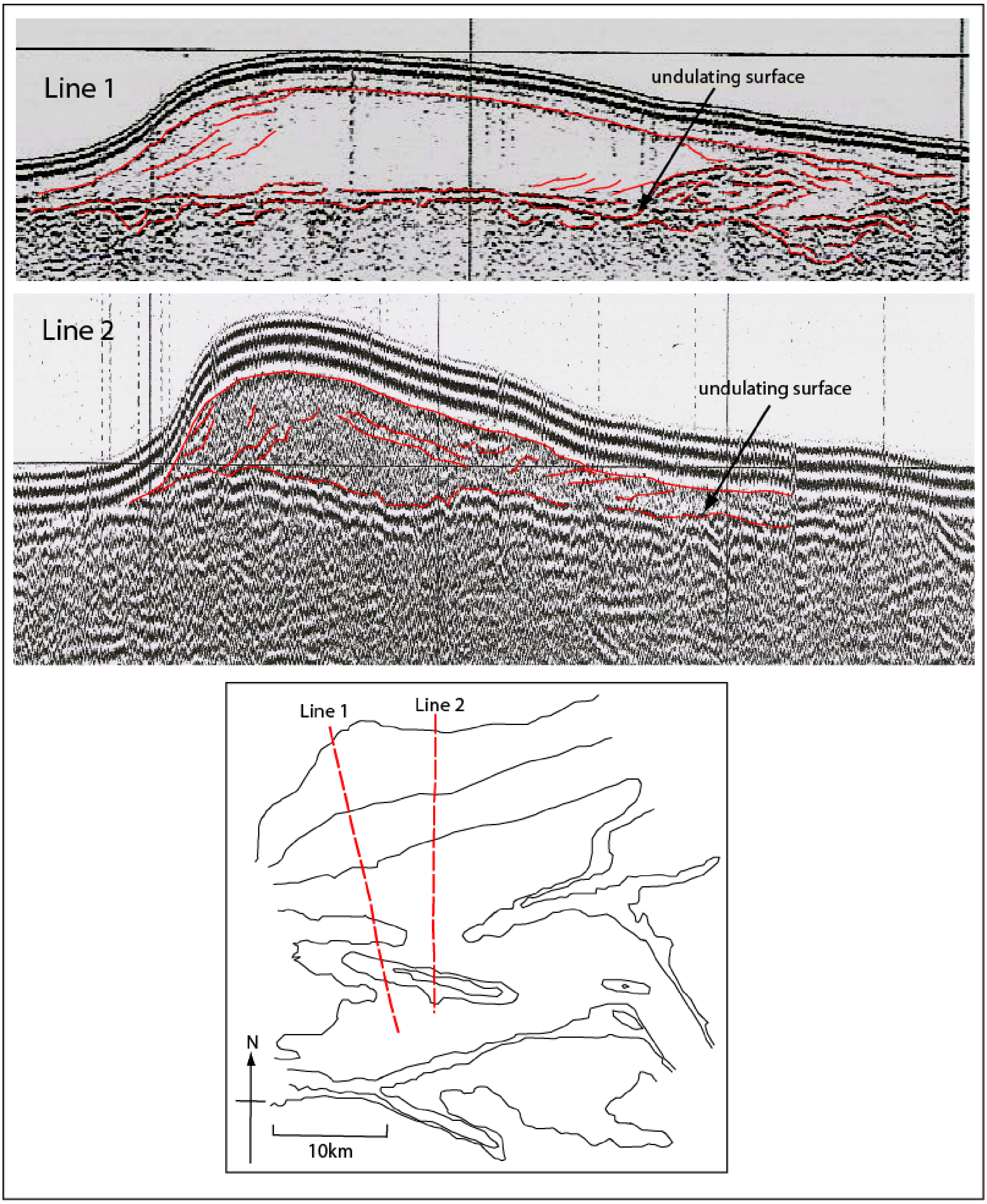

Figure 6: Two 2D seismic lines running through Ridge A and their location. Marked in red are the internal reflectors and at the base, the undulating surface upon which the ridge lies.

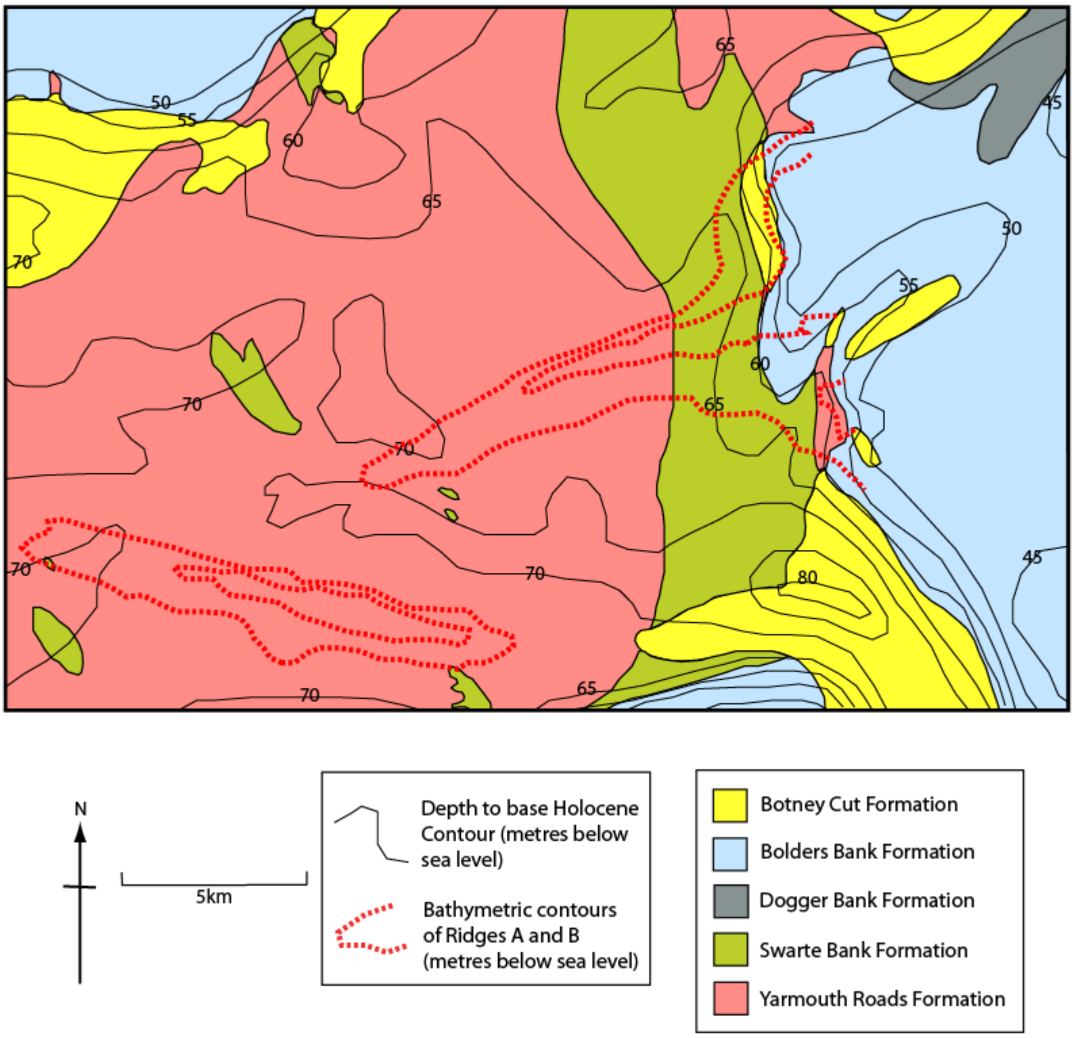

Figure 7: Quaternary Geology map of the eastern end of the OSP overlain with contours of the depth to the base of the Holocene sediments (black) and the bathymetry of Ridges A and B (dashed red). (adapted from the Larminie, 1988 and 1989).

The geology map of the Holocene and Sea Bed sediments (Figure 5) reveals many significant properties of the ridges. The contours of the depth to the base of the Holocene sediments in the Silverwell area shows an absence of the ridges in the topography, thus it can be concluded that the ridges are of Holocene age. The Holocene sediments of the ridges consist of the Tershellingerbank member (of the Nieuw Zeeland Gronden formation), which is defined as open marine sediments and are slightly muddy sands (mean grain size 120-300µm) that have been derived from Pleistocene glacial and periglacial deposits (Larminie, 1988). These sediments are overlain by sands at the seabed and in relation to the surrounding areas this is a larger sediment size than present at bed of the OSP depression but of the same size or smaller than the sea bed sediments on the nearby higher grounds. Furthermore, Figure 8, an RMS amplitude map of the ridge area, depicts the sand ridges as having a high amplitude signal commonly associated with sandy sediments.

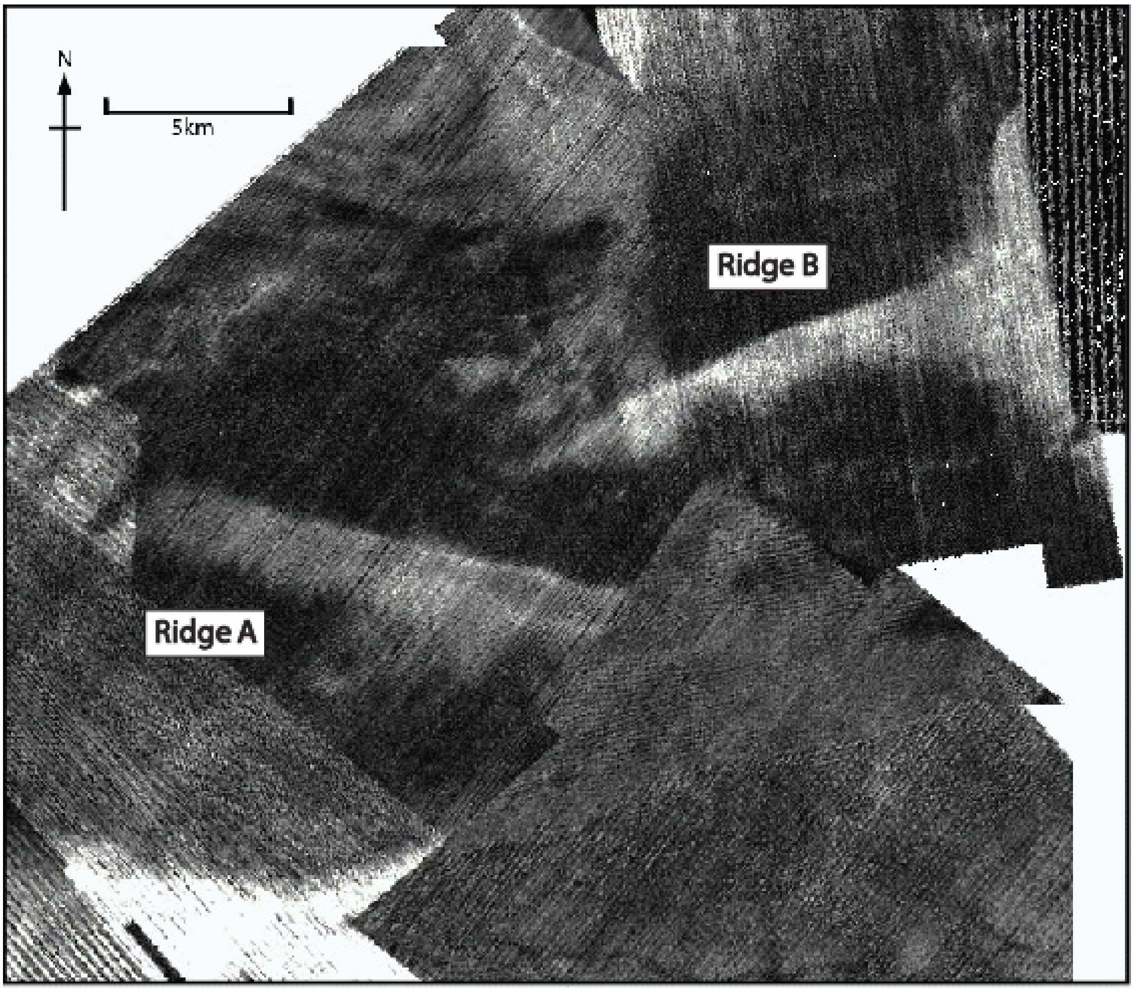

Figure 8: An RMS amplitude map from 0.004 to 0.05 seconds (TWT) of the eastern end of the OSP. N.B. the lighter colours depict high amplitude response areas.

Discussion/Feature classification

Deterioration of ice masses following the last glacial maximum (LGM) has led to a global sea level rise of approximately 120m (Shackleton, 1987) over the past ~20kyrs. In the British Isles a large proportion of the deglaciation had occurred by 13kyr BP (Lowe and Walker, 1997) uncovering vast tracts of the present North Sea Basin as a landsurface that was gradually inundated during the Holocene. Shennan *et al.,* (2000) reconstructed the pattern and timing of inundation in the North Sea basin during the Holocene using reliable indicators of past sea levels obtained from sediment core analysis and geophysical models that integrated ice sheet reconstructions, the Earth's rheology, eustasy and glacio- and hydro-isostasy. It was predicted that the OSP had become a shallow estuary by 9kyr BP, the Dogger Bank had become isolated from mainland Europe at high tide by 8kyr BP and by 6kyrBP the Dogger Bank had become completely submerged. Thus, as it is known from the Holocene geological maps and the bathymetry that Ridges A and B were formed at some time during the Holocene (10kyr BP to present), it is postulated that they are the product of either temperate terrestrial or marine processes.

Temperate terrestrial geomorphological processes

During the early Holocene, in the period prior to marine inundation (10kyr BP to 9kyr BP), a number of terrestrial landscape processes common to temperate environments are likely to have sculpted the landsurface of the Southern North Sea. There are two such classes of processes that could have led to the reworking and deposition of such volumes of sediment retained in Ridges A and B, namely mass movement and fluvial processes. Mass movement, the transfer of material down a slope under the influence of gravity (Summerfield, 1991), is capable of generating significant deposits of fine (colluvium) to large grained (talus) sediment. Inherently, sediments transported by mass movement are deposited at the base of the slope from which they originated. Thus, the situation of Ridge A at 4-5km distance from the proximal slope eliminates the possibility of it being deposited by such a gravitational process. Although Ridge B is attached to a slope of the OSP depression, it too can be disregarded as a mass movement deposit as the feature is approximately the same height as the adjacent slope and is thus by no means a mass of material deposited at the base of it.

It is more reasonable to assume that fluvial processes contributed to shaping of the landscape in the OSP during the very early Holocene as runoff would have been routed to, and exploited, any local depressions. Thus it is a distinct possibility that Ridges A and B were formed in a fluvial environment. Fluvial bedforms, specifically bars, can achieve significant dimensions, developing lengths that are comparable to the width of the channel in which they form (Knighton, 1998). They can take on a variety of shapes and occur in a range of conditions (Knighton, 1998). The point of branching in the OSP channel at its eastern end (in the vicinity of the two ridges-see feature descriptions), may have hosted a confluence of two rivers under fluvial conditions; several observations exist of bars that have formed at confluences in modern fluvial systems (e.g. Melis *et al.,* 1994; Rhoads and Kentworthy, 1995; De Serres *et al.,* 1999). It is therefore possible that features similar to Ridges A and B could have formed in the position in which they currently exist. However, without going into further details of the morphology and formation of such bars it is possible to discount, with some confidence, Ridges A and B as being fluvial bedforms from the very early Holocene. This is because significant evidence exists that is suggestive of at least one major erosional event in the OSP during the early Holocene.

The Quaternary Geology map of the OSP area reveals that, aside from the occasional patches, there is a general absence of late Pleistocene/LGM sediments in the depression which makes it somewhat distinctive from the continuous cover in the surrounding area. Although the absence of these sediments can not be taken as sole evidence of their erosion, it is a notable spatial correlation and reasoning would suggest that it is highly improbable that the ice sheet that covered the OSP area during the LGM (Carr *et al.,* 2006) would not have deposited material either directly, or in the form of proglacial outwash, in such a selective manner in this region[1]. Evidence of erosion in the OSP during the early Holocene is reported by Larminie (1988), who suggests that where the Holocene sediments overlie the Botney Cut Formation (the LGM tunnel valley infill), they are separated by an erosional contact. This is supported by 2D lines crossing the OSP. Figure 10, a 2D line situated to the north of Ridges A and B (Figure 10), shows the base Holocene reflector to truncate the strata below (the Botney Cut Formation). The sediments recorded as overlying the erosional surface in the Outer Silver Pit are described as fully marine (Larminie, 1988);

[1] The Holocene map also shows that the early Holocene, Brackish marine sediments of the Elbow Formation, that exist in the Southern North Sea and in the locality of the OSP are also absent from the depression. Because the Outer Silver Pit is a closed depression it is viable that there were no Holocene Intertidal deposits; when sea level rose sufficiently to cross the threshold of the depression water would have flooded the pit thus the inundation would have been relatively sudden and not a gradual incursion.

consequently, it is plausible to postulate that it is possible that marine processes led to the erosion of late Pleistocene sediments and any early Holocene terrestrial sediments that may once have been present in the Outer Silver Pit. It is also probable that the strong erosional processes capable of removing these sediments would also have reworked Ridges A and B had they been present. As the 2D lines in Figure 6 show, the ridges overly the truncated surface and thus it is likely that they are a product of marine processes that operated subsequent to the recorded erosional event.

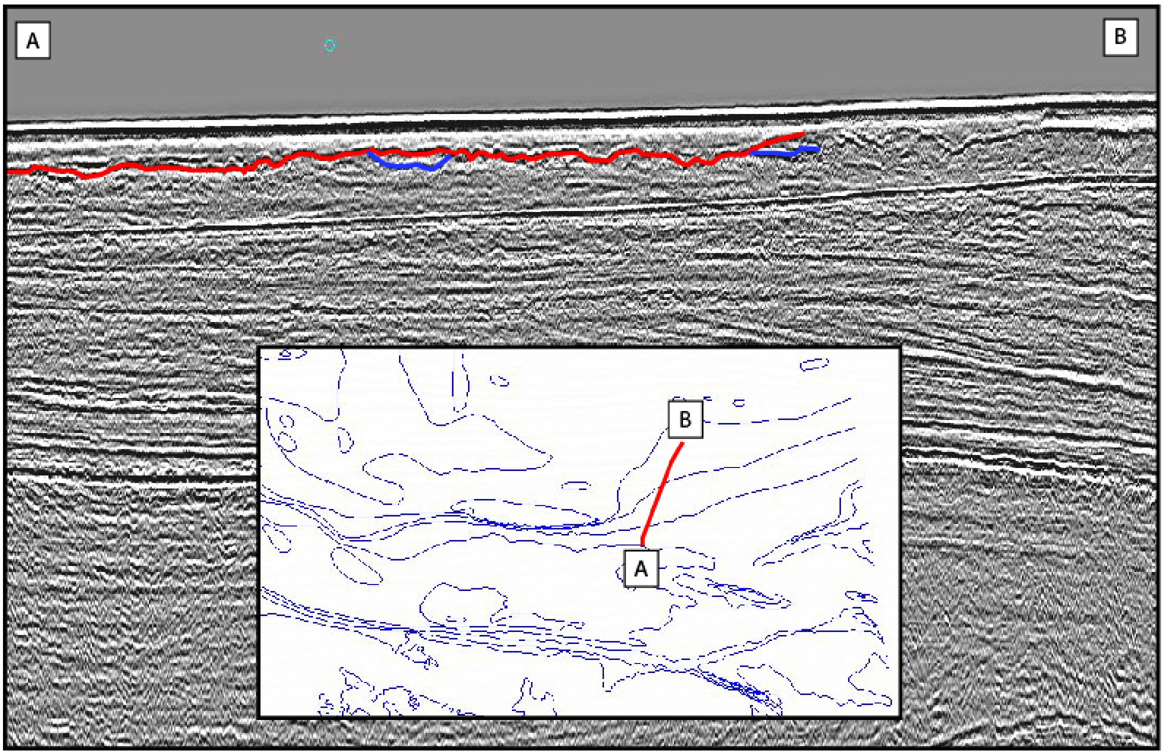

Figure 9: Truncated strata on the bed of the Outer Silver Pit (location depicted in the insert). The red horizon is the Holocene Base reflector (Hebbeln and Meggers, 1999); the blue horizons show the strata that have been truncated.

Marine geomorphological processes

The entrainment and subsequent transportation of the large volumes of sand sized sediments that comprise Ridges A and B would have required relatively strong currents over a sustained period of time. At present the Outer Silver Pit is a low energy environment thought to be sheltered from the high tidal current velocities, that convey from the North, by the shallow Dogger Bank (Eisma, 1975); it is also understood that the OSP is void from the influence of surface waves that are not considered to penetrate beyond 30m in depth in the Southern North Sea (McCave, 1971). Such low hydraulic energy has led to sedimentation on the bed of the OSP[2] of fine sands and silts (Veenstra, 1965; Larminie, 1988) which are locally laminated[3] (Larminie, 1988). Indeed, the late Holocene sedimentary record of the OSP is similar to that of present (Larminie, 1988) suggesting that the depositional environment was

[2] As mentioned in the previous section the sediments on the top of Ridges A and B are of a larger grain size than those on the bed of the OSP but similar to those at similar depth below sea level. This is suggestive of the action of currents and/ or wave action on the tops of the ridges.

[3] The laminations in the OSP form as a consequence of the yearly generation and deterioration of a thermocline at 20-30m depth which is able to develop as a consequence of the minimal impact of tidal action (Eisma, 1975).

also similarly low energy. Furthermore, Shennan *et al.,* (2000) propose that by 6kyr BP the coastline at the margins of the North Sea was comparable to that at present, and therefore it is feasible to suggest that the currents and water depths around the OSP were similar. Consequently, it is probable that Ridges A and B were formed in the early Holocene when the water depth in the OSP was relatively shallow and the currents relatively strong. However, insufficient evidence can be derived from the available sediment and sea level records to further constrain the precise marine influenced environment the ridges were formed in, i.e. estuarine, strait or shallow clastic shelf environment. Therefore, in order to gain such palaeoenvironmental insight the morphology of the features must be examined in respect to modern features of similar form.

The distinctive elongate morphologies of the ridges are similar to a number of landforms that develop subaerially along coasts and on the bed of continental shelves or tidal inlets.

Subaerial coastal landforms

Of the abundance of subaerial coastal landforms that exist globally, Spits and Barrier Islands are identifiable as being depositional elongate ridges. *Spits* protrude into estuary/bay mouths or the sea whilst attached to the mainland at one end (Masselink and Hughes, 2003). They generally have one or more landward pointing recurves at their distal ends, although they can be entirely linear (Bird, 2000). Spits form when there is a sudden break in the direction of the coastline but sediment transport and deposition continues along the original pathway (Haslett, 2000). They are built up above high tide level and lagoons and marshes develop on the sheltered landward side.

There are several aspects of Ridges A and B and their context which suggest it is unlikely that they are spits. Although Ridge B protrudes from 'land' at a change in direction of that land body, as is characteristic of spits, Ridge A is positioned independently at a point where there is no directional change in the 'coast' and thus suggests that Ridge A was not formed by processes of continued longshore drift of sediment along a pathway not contiguous with the coastline. If these features were spits, it would be likely that the amplitude signal from the 'sheltered' area on the landward side would be noticeably lower as a result of the fine sediments that accumulate in low energy environments of lagoons or marshes. Figure 8 shows that there is no notable difference in the amplitude signal between the land adjacent to the ridges on either side, suggesting the absence of a lagoon or marsh. However, it must be noted that it is possible that these sediments were eroded during the Holocene transgression. The longshore drift processes that are responsible for the formation of spits are likely to be minimal within an estuary or strait environment where the flow dynamics are likely to be dominated by the flood and ebb currents of the tide. It may have been possible for a spit to form at a point during transgression if the eastern most land body in the OSP had been an open coastline. Figure 10 however, shows that it was not and that it was inundated at an early point relative to other local land bodies. Finally, although in certain circumstances it is possible that spits can be linear, Ridges A and B lack the recurved end that is frequently characteristic of spits.

Barrier Islands are linear, shore parallel sand bodies that, like spits, extend above sea-level (Masselink and Hughes, 2003). They vary greatly in size and can be up to 100m high, 100's of metres wide and 1000's of metres long (Haslett, 2000). Further similarity to spits extends to the low energy lagoons that are often situated between the barrier and the mainland (Masselink and Hughes, 2003). Again, there are several features of Ridges A and B and their locality which suggests these features are not barrier islands. Firstly, neither of the ridges lies parallel to a potential shoreline. Secondly, at a time when these features would still be in part above sea level, they would have been situated in an inlet (estuary/strait) environment and not on the open coast where Barrier Islands are observed to form (Figure 10). Finally, as previously stated there is no suggestion in the amplitude data of the presence of fine lagoonal sediments in a possible lee area of either ridge, thus suggesting the (possible) absence of the sheltered environment that exists between barrier islands and the mainland today.

Marine bedforms

The possibility that Ridges A and B are either submerged Spits or Barrier Islands is rejected and alternatively a suite of marine bedforms are considered. There are three main elongate sandy bedforms which are commonly identified in marine or tidally influenced environments, these are sand ribbons, sand waves and sand banks/ridges. Sand ribbons are longitudinal bedforms that develop parallel or sub-parallel to the dominant tidal flow current (Cameron *et al.,* 1992). They occur in areas with relatively high surface current velocities, often in excess of 100cm/s (Johnson and Baldwin, 1996). Although sand ribbons vary greatly in size they are generally less than 15km in length, 200m wide and up to 1m thick (Kenyon, 1970), however, as a result of their diminutive thickness they tend not to be identifiable on seismic images. Sand waves, or small/medium subaqueous dunes or megaripples as they are sometimes referred to, are flow transverse features, i.e. their crests lie approximately perpendicular to the direction of the main current (Cameron *et al.,* 1992; Blondeaux, 2001).They are greater in height than sand ribbons, generally falling between 1.5m and 10m in thickness (Ashley, 1990; Johnson and Baldwin, 1996) and they generally occur in areas where current velocities exceed 65cm/s (Johnson and Baldwin, 1996). Sand waves can be both symmetrical and asymmetrical in form (Blondeaux, 2001) with maximum slope angles of 10-12° (Cameron *et al.,* 1992).

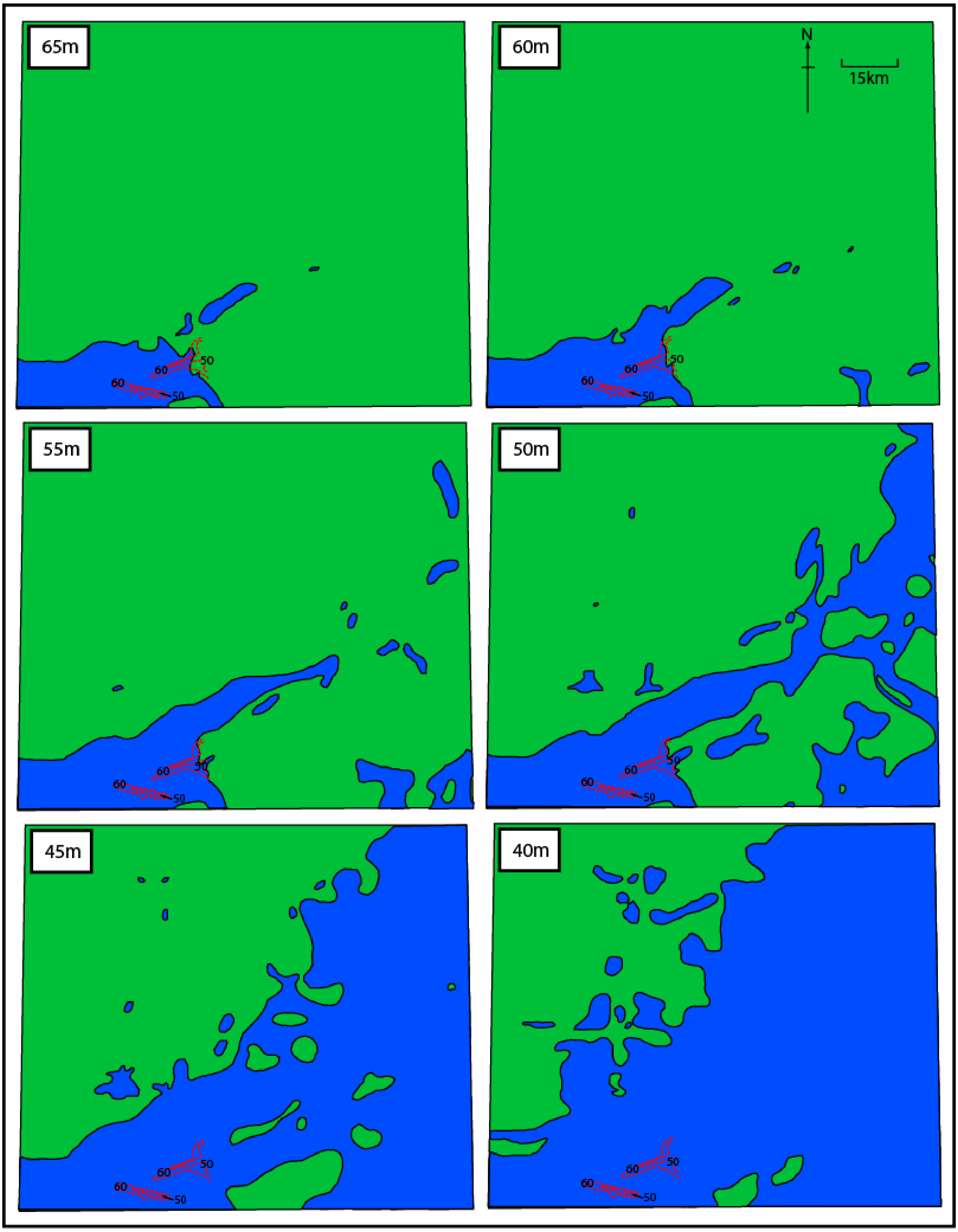

Figure 10: Changes in land area with rising sea level based upon the depth to base Holocene map (Larminie, 1989). Ridges A and B are depicted in red.

Ridges A and B are both significantly larger (table 1) than the reported dimensions of sand ribbons or sand waves and thus it is concluded that they are neither feature. Conversely, sand banks/ ridges are much larger features commonly measuring up to 80km long, 1-3km wide and 10-50m high (Johnson and Baldwin, 1996; Dyer and

Huntley, 1999) and are of similar dimensions to Ridges A and B. Furthermore, in agreement with the morphology of Ridges A and B, Johnson and Baldwin (1996) describe sand banks/ ridges as linear bedforms that are asymmetrical in cross section and composed of medium to fine sands.

The morphological evidence is strongly suggestive that Ridges A and B are indeed sand banks/ridges in terms of both the similarity which exists between them and also the elimination of a range of other possible features based on the available data. However, this study is lacking in the provision of data which may provide validation for such claims. For example, higher resolution seismic data and sediment cores may be able to provide greater detail of the internal structure of the features and thus reveal the presence/absence of certain sedimentary structures characteristic of sand banks/ ridges[4]. Also access to high density sediment analysis would provide information of the existence/ absence of any sediment grain size gradations which occur across the features[5]. However, the body of evidence available is considered to be sufficient to classify, albeit tentatively, the features in the OSP as sand banks/ridges. Thus it is with a degree of uncertainty that the following inferences based on this classification are made.

It is possible to make the distinction between sand ridges and sand banks based upon the criteria that ridges have a length/width ratio which exceeds 40 (Amos and King, 1984). The length/width ratios of Ridges A and B are just 9 and 5.5 respectively and so are herein referred to as 'sand banks'.

Sand banks may exist in either an active or moribund state. Active sand banks are present in areas where the tidal currents are relatively strong (>50cm/s). Their crests are shallow in the water and are generally quite sharp, except from when they approach sea level at which point they display flattened tops. The steep slopes of active ridges are relatively steeply inclined at ~6° and they are flanked with sand waves. Conversely, moribund sand banks are situated in relatively deep water where currents have diminished to <50cm/s and are thus insufficient to transport sand on the sea bed (Johnson and Baldwin, 1996; Dyer and Huntley, 1999). They have rounded profiles, shallow slopes of <1° and an absence of sand waves on their flanks (Johnson and Baldwin, 1996; Dyer and Huntley, 1999). The surrounding sea floor is covered by sandy or muddy sediments as opposed to larger gravel sized sediments found in the vicinity of active banks (Johnson and Baldwin, 1996; Dyer and Huntley, 1999). As previously stated it is known that the currents in the OSP are at present very weak, the crests of the sand banks are flattened and at a depth of approximately 40-50m below sea level and the steep slopes have mean angles of only 0.84° and 1.13°. The 2D seismic data depicts the surfaces of the sand banks as smooth and thus void of sand waves. Thus on the basis of the geomorphological observations the sandbanks in the OSP are interpreted as being moribund.
The moribund sand banks within the Outer Silver Pit are thus relict features formed in conditions that no longer exist at this location. Such features can provide insight into past environments through the understanding of processes which operate to form analogous landforms in the present day. Thus it is the purpose of the following section

[4] The internal structure of linear sandbanks within estuaries is likely to vary according to the precise controls upon its formation and maintenance. However, one may expect to observe such structures as clay drapes, rippled beds, 'ebb foreset packages' (Fenies *et al.,* 1999).
[5] Larger particle sizes are likely to be observed on the flank where the dominant currents operated.

to utilise the available data in an attempt to reconstruct the processes that operated in the OSP when the sand banks were last active, i.e. the Early Holocene[6]. Due to the rapid and significant extent of sea level rise in the North Sea during the early Holocene, marine conditions within the OSP would have undergone significant transformations e.g. from estuarine, to strait, to open sea. Therefore the suite of processes which initially led to the formation of the sand banks may have been different to those that ultimately shaped and maintained them prior to their becoming moribund.

Environmental Interpretation

Sand Bank classification

Sand banks develop in a number of marine settings where there is an abundant supply of sand and currents strong enough to transport the material. Dyer and Huntley (1999) produced a classification system of the various types of sand banks resulting in their following subdivision; i) offshore banks, ii) estuary mouth banks (including those formed on tidal deltas and those formed in wide estuary mouths where there is not a delta) and iii) headland associated banks (separating those banks formed around stable and recessional headlands).

Offshore sand banks are described by Blondeaux (2001) as rhythmical features that generally have a crest spacing of a few kilometres. They form with their long axes orientated at a small oblique angle to the peak tidal flow direction, where there is a convergence in bed load transport paths (Dyer and Huntley, 1999). The crests of the banks are often at or only a few metres below the sea surface at low tide. Although there is more than one 'ridge' within the OSP, sets of offshore sand banks/ridges that have been identified in the Southern North Sea commonly occur in groups of 10 or more (Cameron *et al.,* 1992). Furthermore at a point when the crests of the banks would have been at or close to sea level at low tide (50-40m below present) the OSP would have been an enclosed sea way (i.e. an estuary or a strait)[7] not an offshore environment. Thus it is concluded that the sand banks within the Outer Silver Pit can not be classified as 'offshore banks’.

It is also possible to conclude that Ridges A and B are not headland associated banks. The crests of the sand banks in the Outer Silver Pit are at approximately the same depth below sea level (~50m) as the surrounding 'land/potential headland'. Thus at the point when the water depth in the OSP was sufficient to submerge the sandbanks, the adjacent ‘land’ would also have been submerged and thus could not have been a headland. However, it is possible that at some time in their development the ridges were a lesser height in relation to the 'land' but subsequently gained height during inundation and landform evolution. Nonetheless, regardless of the stage of inundation, at a time when these sandbanks could have been active, the OSP would have been a seaway, not an open coastline with significant longshore processes to generate headland banks. Furthermore, there remains several components of the sandbanks' morphology which suggests they are not headland associated banks. Headland

[6] As previously noted in this section, by 6kyr BP the North Sea sea levels, and consequently the conditions in the OSP, were similar to those at present.

[7] Assuming a tidal range of less than 10 metres based on the current (~8m) and Holocene tidal ranges (<8m) of the Humber estuary on the east coast of England (Shennan and Horton, 2002). Data for Holocene tidal ranges of the OSP is not currently available.

associated banks occur in zones of littoral sediment transport convergence where there is an acute change in the direction of the coastline (Dyer and Huntley, 1999). Sandbanks form on one or both sides of the headland and are separated from the headland by a deep narrow channel; this precludes Ridge B from being a headland associated ridge as it protrudes from the cusp of the 'land' to which it is attached. Headland associated banks are generally only a few kilometres in length and thus are much shorter than the sand banks in the OSP (table 1). The sand banks in the OSP also do not portray the pear shaped form commonly associated with headland banks (Dyer and Huntley, 1999).

On the premise that, when last active, the sand banks within the OSP would have been close to the water surface or exposed at low tide it is possible to infer, based on the submerged topographical data, that the sand banks were probably formed in a strait or estuary environment. Figure 11 is a generalised depiction of the possible landsurface, intertidal areas and inundated areas using bathymetric data and based on a series of assumptions which, at present, can not be further constrained:

1. It is assumed that the crests of the sand banks within the OSP were at, what is now, 50m below sea level when they were last active. However, this depth is derived from the bathymetric data which has a 10m contour interval and so theoretically the crest of the bank could lie at a lesser depth, up to ~41m below present sea level. Also, the height of the sand bank may have changed from when it was last active as a result of post inundation deposition. Larminie (1988) suggests a generalised range of only 1-5m of Holocene sedimentation in the locality of the sand banks. It is not thought that erosion would have had a significant impact upon the elevation of the sand banks since having been moribund, because for them to be moribund the local currents operating must be so weak as to not transport sand sized material.
2. It is assumed that the topographic expression of the sea bed surface represented in the bathymetric data approximates to that of the landsurface at the time the sand banks were last active. For the same reasons as stated above in point 1, this may not have been so. The topographic data in the depth to the base of the Holocene map may have presented a better representation of this landsurface[8], however, because the sand banks are Holocene features they are not depicted. In using the bathymetric data it is argued that a more realistic comparison of the sand banks' elevation in relation to that of the surrounding land is acquired as late Holocene deposition is compensated for; this is of course making the further assumption that the sedimentation rate was spatially continuous.
3. The tidal range is assumed to be 10m, with low tide being at, what is today, 50m below sea level. The level of low tide is based on the further assumption that the crest of the active sand banks became exposed at low tide. This is a somewhat robust assumption based on observations of present day sand banks in estuaries (e.g. Wright, *et al.,* 1975). However, in constraining low tide to 50m, i.e. the height of the crest, all of the above errors outlined in point 1 are subsumed. The assumption of a 10m tidal range is an unavoidable gross assumption, as there is no data available for Holocene tidal ranges/prisms within the OSP. Nonetheless the assumption is founded in the knowledge of

[8] Although, nonetheless problems exist with quantifying and identifying the spatial and temporal distribution of erosion, and subsequently inferring the effects the change in topography may have had on hydraulic processes relative to the stages of sandbank formation.

present day tidal ranges of the nearby east coast of England and the current understanding of the effects of such a land configuration on tidal prisms. The exact range of 10m was taken for the purpose of simplicity in the production of Figure 11 as the contour spacing of the bathymetry is 10m. This assumption is very limiting for this study, as the land surrounding the OSP is very low relief and therefore a small difference in the tidal range could produce very different marine conditions from high to low tide.

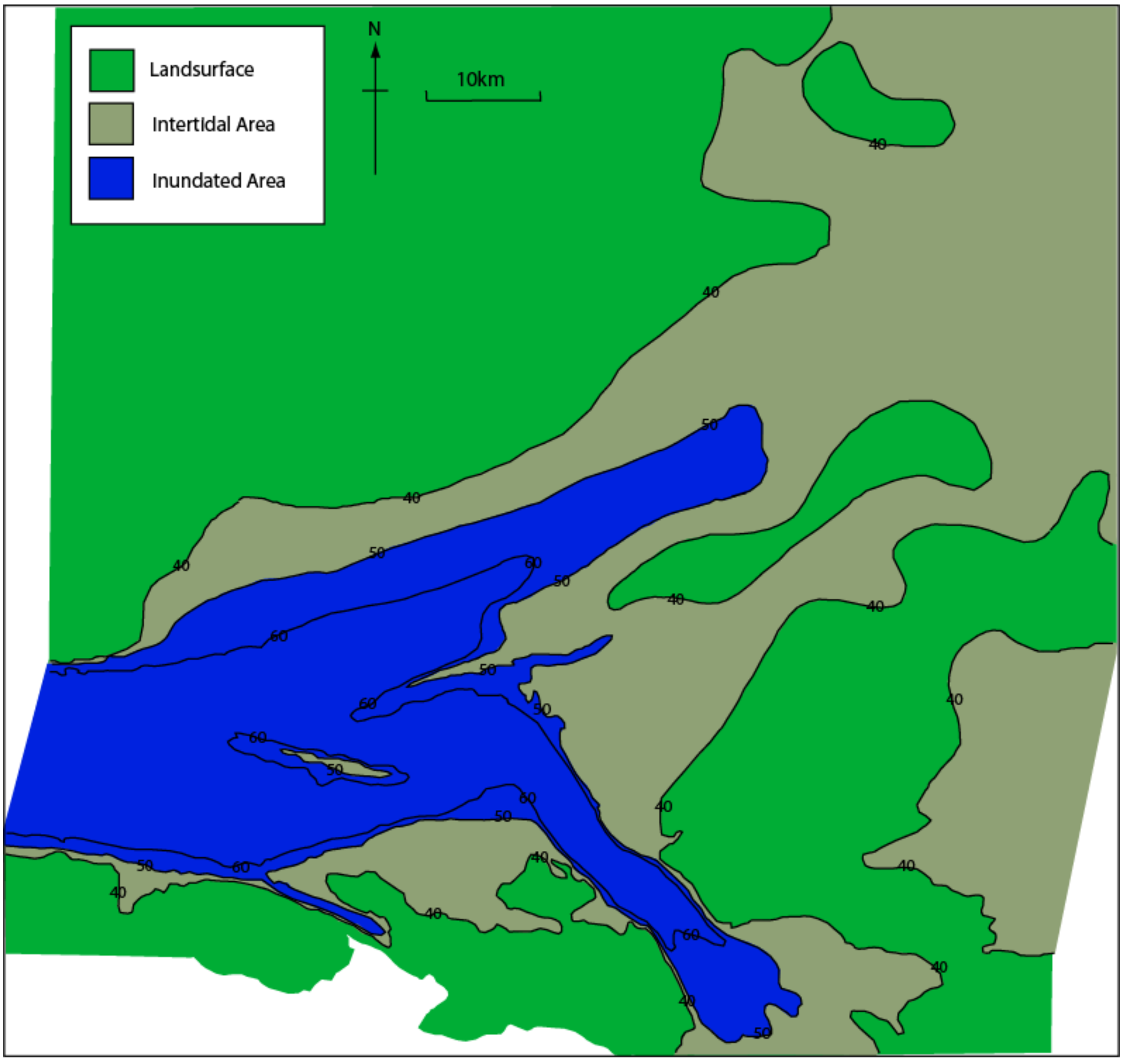

Figure 11 : Postulated land configuration at the time when the sand banks were last active. The contours are the bathymetric contours from Larminie (1988). The map is based on a number of assumptions highlighted in the text.

Estuaries and Sand Bank Formation

Estuaries are varied and complex environments; the unique suite of processes which operate in a given estuary are a function of several individual factors (e.g. tides, waves, fluvial input and morphology) which are temporally dynamic over the short- and long-term. In order to gain insight and aid interpretation of estuaries and their many facets, several classification schemes of various estuarine attributes have been proposed. Such classifications are also valuable in the interpretation of past estuarine conditions based on relict features, thus potentially aiding an interpretation of past conditions within the OSP founded on the available information of the moribund banks and their surroundings.

Hayes (1975) proposed that estuaries could be grouped according to tidal range. He applied Davies' (1964) scheme of tidal classification which identified three classes of tidal range, micro-tidal (0-2 m), meso-tidal (2-4m) and macro-tidal (>4m). Based on extensive observations of shorelines from around the world, Hayes (1964) concluded that distinctions could be made between tidal ranges and their associated suite of depositional landforms. Most notably he observed that linear sand banks were widely associated with macrotidal estuaries. The currents in macro-tidal estuaries are strong and capable of transporting the relatively coarse grain sediments which form sand banks.

In macro-tidal/tidally dominated[9] estuaries independent and mutually evasive flood- and ebb-dominated channels are widely observed (e.g. Robinson, 1960; Price, 1963; Ludwick, 1975; Wright *et al.,* 1975). These mutually evasive channels develop as a result of the tidal asymmetry which occurs in such estuaries. Tidal asymmetry refers to the asymmetry in magnitude, velocity and duration between the flood and ebb tides in a given estuary (Masselink and Hughes, 2003). The asymmetry is generated as a result of differences in the magnitude of influence exerted by friction upon the flow of each tide. In flow contained within the channel a lesser proportion of the volume of the flood tide (i.e. the crest of the tidal wave) is in contact with the channel surface (and therefore friction) than that of the shallow ebb tide. Thus the velocity of in channel flow of the flood tide is greater than that in the ebb. Given that the discharge volume through the channel of the flood tide is similar to that of the ebb tide, it follows that the duration of the ebb tide is greater in order to compensate for the reduction in velocity.

In reality the tides in many estuaries are not fully contained within the main channel along its entire length or throughout the full tidal cycle, frequently, broad zones of intertidal land flank the channel. Friedrichs and Aubrey (1988) examined the relationship between channel shape and tidal asymmetry. They suggested that where large intertidal areas become submerged during the flood tide, the proportion of the water volume exposed to surface friction effects, and indeed the severity of friction[10], is increased and thus the efficiency of the conveyance of the tide up-estuary is reduced. Conversely, during the ebb tide, when water levels are lower, flow is conveyed within the channel where there is a lesser frictional effect. Consequently, the ebb tide has a greater velocity than the flood tide in areas with significant tracts of intertidal land, leading to an overall ebb-dominance. In areas of ebb-dominance, ebb-dominated currents/channels exist within the main channel and the flood dominant currents prevail over the shallow, intertidal areas. To the contrary, in areas of flood dominance, it is the flood currents that dominate in the main channel and the ebb currents that are prevalent along the shallow margins of the channel. It is not uncommon for estuaries to exhibit portions of both flood- and ebb-dominance as a result of the changing morphology along their lengths. Ebb-dominance more frequently occurs towards the head of estuaries where there are larger intertidal zones and flood-dominance mostly occurs towards the mouth. At a given point these regions

[9] As a consequence of large tidal ranges, the driving morphological processes in such estuaries are tidally dominated.
[10] The roughness coefficient (measure of friction) is generally greater over intertidal areas than within channel as a result of the vegetation which colonises the periodically emergent land.

will interdigitate (Harris, 1988) and, as previously mentioned, the flood and ebb channels tend to be mutually evasive.

It is between the mutually evasive ebb and flood channels within estuaries, where bed loads converge and sand banks occur (e.g. Wright *et al.,* 1975; Harris, 1988; Harris *et al.* 1992; Dyer and Huntley, 1999). The sand waves which exist on the sides of sand banks are commonly observed to be ebb or flood orientated on opposite sides (e.g. Wright *et al.,* 1975; Harris, 1988). Harris (1988) suggests that this is indicative of a 'circulatory pattern around the sandbank crest'. In most cases the tidal flow in one direction will dominate; it is this that causes the sand bank cross-sectional asymmetry (Kenyon *et al.,* 1981). In estuaries sand banks migrate away from their steep slopes, as it is the steep slope that is actively eroded (Dyer and Huntley, 1999). Also sand banks are commonly orientated obliquely to the direction of peak tidal flow (Kenyon *et al.,* 1981).

Only a restricted volume of quantitative studies on sand banks in estuaries exists. Such quantitative study would broaden and improve the depth of understanding of the processes operating to build and maintain the banks. This therefore limits the level of environmental interpretation that can be derived from the sand banks within the Outer Silver Pit. Nevertheless it is possible to derive some basic understanding of the processes in the OSP during the early Holocene based on the sand banks.

The presence of the sand banks suggests that the OSP was macrotidal and therefore tidally dominated. The tidal currents would have been strong and capable of transporting the sand sized sediments in the sand banks. Tidal asymmetry is likely to have led to the formation of ebb- and flood-dominated channels between which the sand banks may have formed. It is possible that 'Ridge A' was formed in such a way, as it lies independently in the 'main channel' and trends in such a manner relative to the surrounding topography that it is plausible to suggest it would have formed at an oblique angle relative to the direction of the dominant current. However, the positioning of 'Ridge B', attached to a mass of land and at the confluence of the branch in the OSP depression, may suggest an alternative mode of formation. The convergence or divergence of water moving downstream or upstream along the branches of the OSP most probably resulted in certain hydraulic processes that would have led to a loss of fluid energy and the subsequent deposition of the sand bank sediments. Furthermore, the two smaller depressions directly to the south of Ridge B (see feature descriptions) may have accommodated fluvial and/or tidal channels that are also likely to have influenced the hydrodynamics controlling the processes forming/maintaining the sandbanks. However, it is beyond the scope of this study to examine in further detail the precise fluid mechanics which may have operated in this area which resulted in the deposition of 'Ridge B'. It is however, important to suggest that sedimentological analysis and 2D seismic data depicting the internal structure of Ridge B has the potential to provide clarification of the precise processes which operated to form the sandbanks.

Conclusions

The two elongate ridge features identified in the Outer Silver Pit in the 3D seismic data are interpreted as moribund sand banks which formed in as estuarine environment during the early Holocene marine transgression between approximately 10k and 6kyr BP. From this interpretation and knowledge of modern analogues, it is inferred that the tidal range of the Outer Silver Pit was probably macro-tidal and the tidal currents conveyed in the estuary were relatively strong. The identification of a strongly undulating truncation surface in the 2D seismic data suggestive of a major erosional event is supportive of the theories of Donovan (1965; 1975) which suggest that strong marine currents were, at least, in part responsible in the formation of the OSP depression.

It is of paramount importance to highlight that this study and its findings are fundamentally restricted by the limitations of both the available data and the level of current understanding of the modern analogous systems. In terms of the available data, limitations arising from the relatively low resolution of the 3D seismic images were, in part, overcome by use of supplementary 2D seismic data. However the data quality remained insufficient to examine/identify very fine scale sedimentary structures, for example clay drapes and rippled beds, which have the potential to provide validation of the geomorphological classification of the sand banks. Superior resolution 2D line data (e.g. high frequency sonar source) through the sand banks may have allowed for examination of such structures, however this data was not available. A lack of physical evidence in the way of sedimentary data from cores was unavailable for this study, indeed such data could have provided the means for ground truthing and validation of the interpretations made by providing, for example, information of grain size distribution and detailed internal structure. Furthermore, sediment cores could provide the opportunity for dating and therefore constraint of the timing when the ridges were last active. Finally, current limits in quantitative understanding of the formation and dynamics of sand banks in estuaries, as previously mentioned, render the prediction of the processes that led to the formation of the sandbanks in the OSP problematic. As Knighton (1998) suggests 'ancient deposits and bed forms are a key element in palaeohydraulic reconstruction'; however, 'the reliability of such reconstructions depends on an adequate understanding of the formative processes operating within the present-day environment'.

Despite the limitations and restrictions faced by this study it has highlighted the suitability of 3D seismic data for the identification and broad interpretation of submerged, large scale, geomorphological features and their surroundings. Furthermore this study has reflected the vital and significant role that relict landforms play in the reconstruction of past environments and specifically the palaeohydraulic processes. Although nonetheless, fervently demonstrating the importance of the reliability and wealth of data in producing reconstructions which are robust and of scientific value.

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