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Middle–Late Pleistocene landscape evolution of the Dover Strait inferred from buried and submerged erosional landforms

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ABSTRACT

Prominent landforms, either buried or preserved at the seafloor, provide important constraints on the processes that led to the opening and present-day configuration of the Dover Strait. Here, we extend previous investigations on two distinct landform features, the Fosse Dangeard and Lobourg Channel, to better understand the poly-phase history of their formation and inferences for the opening and Pleistocene evolution of the Dover Strait.

The Fosse Dangeard consist of several interconnected palaeo-depressions. Their morphology and spatial distribution are interpreted to be the result of plunge-pool erosion generated at the base of north-eastward retreating waterfalls. Their infills comprise internal erosional surfaces that provide evidence for the occurrence of several erosional episodes following their initial incision.

The Lobourg Channel comprises various sets of erosional features, attesting to the occurrence of several phases of intense fluvial and/or flood erosion. The last one of these carved a prominent inner channel, which truncates the uppermost infill of the Fosse Dangeard. The morphology of the Lobourg inner channel and the erosional features associated with its incision strongly resemble landforms found in megaflood-eroded terrains, indicating that this valley was likely eroded by one or several megafloods.

Our study therefore corroborates the existence of waterfalls in the Dover Strait at least once during the Pleistocene Epoch. It also provides evidence of the occurrence of multiple episodes of fluvial and flood erosion, including megafloods. Finally, this study allows us to establish a relative chronology of the erosional/depositional episodes that resulted in the present-day morphology of this region.

1. Introduction

The opening of the Dover Strait is widely considered to be one of the most important events in the Pleistocene evolution of northwestern Europe (Smith, 1985; Gibbard, 1988, 1995; Gupta et al., 2007; Gibbard and Cohen, 2015). It had a significant impact on the organization of northwestern European river systems at continental scale, as well as on biogeographic distributions and the pattern of early human colonization of Britain (Preece, 1995; Sutcliffe, 1995; Stuart, 1995; Meijer and Preece, 1995; Ashton and Lewis, 2002; Ashton and Hosfield, 2010; Ashton et al., 2011). Whilst the importance of the opening of the Strait has long been recognized the timing and processes responsible have remained unclear. Understanding these factors is key to improve reconstruction of the Quaternary history of northwestern Europe.

According to palaeogeographic reconstructions, northwestern France and south-eastern Britain were connected by a land-bridge, which was formed by the northern limb of the Weald–Artois anticline (Figs. 1 and 2; Gibbard and Cohen, 2015; Gibbard and

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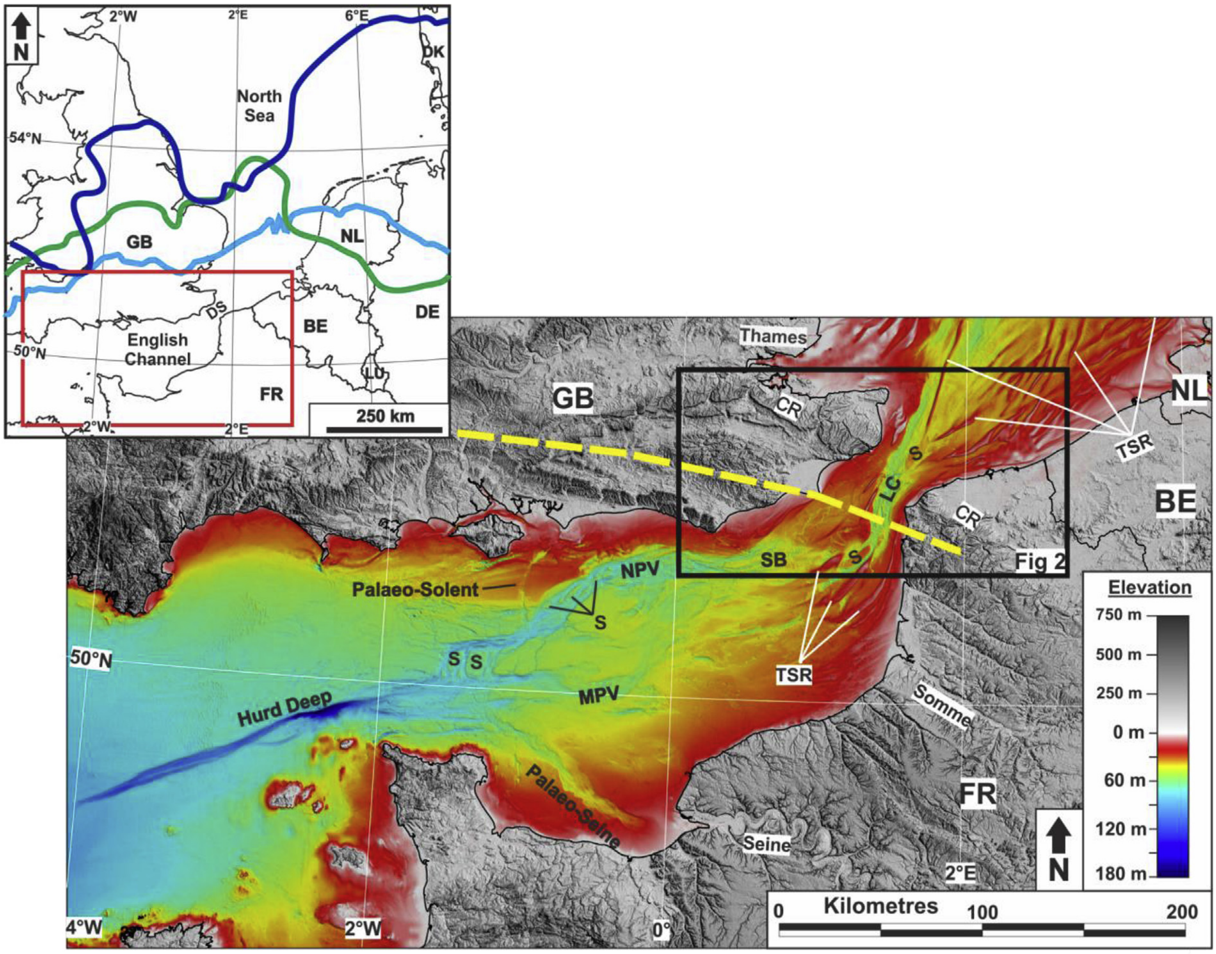

Fig. 1. Topographic–bathymetric map of the English Channel area derived from SRTM elevation and EMODnet bathymetric data. The map is illuminated (light direction altitude: 60°, azimuth: 60°) and has a vertical exaggeration of 9.5. Digital data from www.emodnet-bathymetry.eu and www.cgiar-csi.org/data/srtm-90m-digital-elevation-database-v4-1. For this and following figures, horizontal datum: WGS84; SRTM vertical datum: EGM96; bathymetric vertical datum: lowest astronomic tide (LAT). The dashed yellow line marks the axis of Weald–Artois anticline and the black rectangle the area shown in Fig. 2. Annotation is as follows. Channel palaeovalleys: LC: Lobourg Channel; SB: South Basserelle palaeovalley; NPV: Northern palaeovalleys; MPV: Median palaeovalley. TSR: tidal sand ridges; S: streamlined islands; CR: chalk ridge. Inset: Political map of northwestern Europe with maximum southern extent of ice sheets during Elsterian (light blue), Saalian (green) and Weichselian (dark blue) glaciations according to Ehlers and Gibbard (2004), Lee et al. (2012) and Sejrup et al. (2016). The red rectangle marks the area shown in main figure. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

Lewin, 2016). It has been proposed that the land-bridge (hereafter referred to as "the Weald–Artois ridge"), separated the North Sea basin from the English Channel until ~450 ka ago (Gibbard and Cohen, 2015). At that time, northwestern Europe was in the middle of the Elsterian glaciation, during which the Irish–British and Fennoscandian ice sheets merged for the first time across the central North Sea region (Ehlers and Gibbard, 2004). The coalescence of the ice sheets caused the isolation of the southern North Sea basin from the North Atlantic Ocean, and blocked the northern drainage routes of palaeo-rivers (Gibbard, 1995). This geographic setting, complemented by flexural loading due to the load of the ice sheet, is thought to have provided the conditions for the formation of an extensive proglacial lake in the southern North Sea region, impounded by the merged ice sheets to the north and the Weald–Artois ridge to the southwest (Smith, 1985; Murton and Murton, 2012; Gibbard and Cohen, 2015; Gibbard and Lewin, 2016).

Several authors have proposed that at some point during the Elsterian glacial maximum, the southern North Sea lake overtopped the Weald–Artois ridge, resulting in breaching of the rock ridge and eventual creation of the Dover Strait (Smith, 1985; Gibbard and Cohen, 2015; Gupta et al., 2017). Testing this hypothesis requires knowledge of palaeo-landscapes preserved on the seafloor and in the shallow subsurface of the Dover Strait.

Two striking large-scale erosional features are present in the Dover Strait area: the tens-of-metre deep sediment-infilled palaeo-depressions known as the Fosse Dangeard (Destombes et al., 1975; Smith, 1985; Gupta et al., 2017), and a prominent palaeovalley known as the Lobourg Channel, which connects downstream to the Channel palaeovalley system (Figs. 1 and 2; e.g., Mellett et al., 2013; Collier et al., 2015).

In a recent study, Gupta et al. (2017), following on an earlier suggestion of Smith (1985), presented initial evidence indicating that the Fosse Dangeard palaeo-depressions represent fossilized plunge pools formed by waterfalls plunging over the Weald-Artois ridge. The presence of these palaeo-plunge pools lends credence to the model of a proglacial lake having existed during the Elsterian glaciation in the southern North Sea basin. That study proposed that waterfall recession eventually led to the breaching of the Weald–Artois ridge, thus opening the Dover Strait.

The opening of the Strait during the Elsterian glaciation is consistent with sediments sampled in the Bay of Biscay, which attest to the occurrence of fluvial exchange (including massive

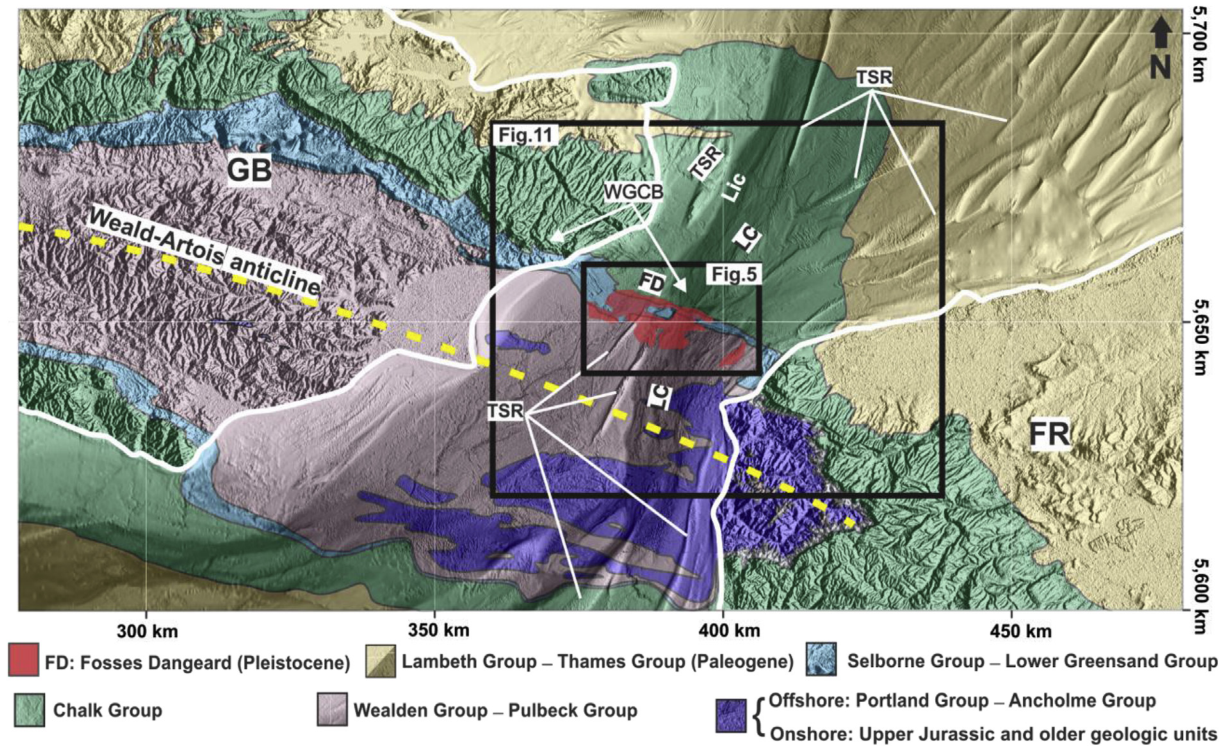

Fig. 2. Bedrock geology (modified from Balson and D'Olier. (1989) and Crosby et al. (1988)) and Fosse Dangeard location (this study) superimposed on a bathymetric–topographic map of the Dover Strait region. Note that the Fosse Dangeard are carved along the northern limb of the Weald-Artois anticline. The dashed yellow line marks the axis of Weald–Artois anticline. LC: Lobourg Channel; FD: Fosse Dangeard (red areas); Lic: Lobourg inner channel; TSR: tidal sand ridges; WGCB: boundary between White Chalk and Grey Chalk subgroups; thick white line: coastlines. Projection: UTM, WGS84, zone 31N.

fluvial discharges) between the southern North Sea basin and the English Channel during the Elsterian glacial maximum and subsequent Pleistocene marine lowstands (Toucanne et al., 2009a, 2009b, 2010, 2015). Fluvial exchange between these regions during Middle–Late Pleistocene marine lowstands is also attested by the incision of the Channel palaeovalleys and their continuation into the southern North Sea across the Dover Strait (Antoine et al., 2003; Lericolais et al., 2003; Hijma et al., 2012; Mellett et al., 2013; Collier et al., 2015). Importantly, the morphology and erosional pattern of some of these palaeovalleys (e.g., the Northern Palaeovalley and the Lobourg Channel) strongly suggest the occurrence of high discharge flood events – megafloods – similar to those resulting from the breaching of the rock dams and overspill of major lakes (Gupta et al., 2007, 2017; Collier et al., 2015).

The landscape evolution of the Dover Strait since its formation remains however largely unknown. Here, we extend our earlier investigations by describing and analysing in detail the subsurface and seafloor palaeo-landscapes preserved in the Dover Strait and establish their possible cause and relative chronology. To do so, we firstly describe and interpret the 3-dimensional geometry and spatial distribution of the various palaeo-depressions that comprise the Fosse Dangeard. Secondly, we analyse the various seismic units composing their infills, including seismic facies and internal erosional surfaces. Finally, we characterise the morphology of the Lobourg Channel, establishing the poly-phase history of its formation and its relationship with the formation of the Fosse Dangeard and other erosional/depositional landscape features.

The combined analysis and interpretation of the erosional/depositional landforms preserved in the submarine Dover Strait provides invaluable constraints toward understanding the evolution of buried and submarine landscapes of the English Channel and southern North Sea, as well as their relationship with Quaternary glacial/interglacial cycles.

2. Geological and palaeogeographic setting

A series of WNW-striking Cretaceous and Jurassic sedimentary formations outcrop on the seafloor of the Dover Strait (Fig. 2). These geological formations are folded into the regional Weald–Artois anticline (Fig. 2) and locally deformed and offset by the North Artois Shear Zone (Figs. 4 and 5; García-Moreno et al., 2015). In this paper, we refer to pre-Quaternary formations as bedrock.

Bedrock formations are locally covered by sand dunes and tens-of-metres to kilometre-long sand ridges (Fig. 2; see also Dyer and Huntley, 1999; James et al., 2002; Reynaud et al., 2003). Both the Fosse Dangeard and the Lobourg Channel are carved predominantly into bedrock (García-Moreno et al., 2015).

The onshore bedrock geology is well-known, and is classified into 7 Groups and 24 Formations (see Table 1 for details). Recognising and mapping these units across the Strait itself relies largely on their geophysical expression (García-Moreno et al., 2015). For example, the base of the Holywell Nodular Chalk Formation and the Hythe Formation are well marked in the bathymetry, forming, respectively, a 1–3 m high scarp and 5–10 m high ridge (Fig. 5b). Other bedrock formations exhibit little to no distinct geomorphologic expression across the Strait. The Hythe Formation and the Holywell Nodular Chalk Formation appear thus to be more resistant to erosion than other neighbouring bedrock units.

In seismic-reflection data, most bedrock formations are easily recognized by comparing with previous interpretations of offshore seismic-reflection data and boreholes, and with the thickness of these formations measured in onshore outcrops and boreholes

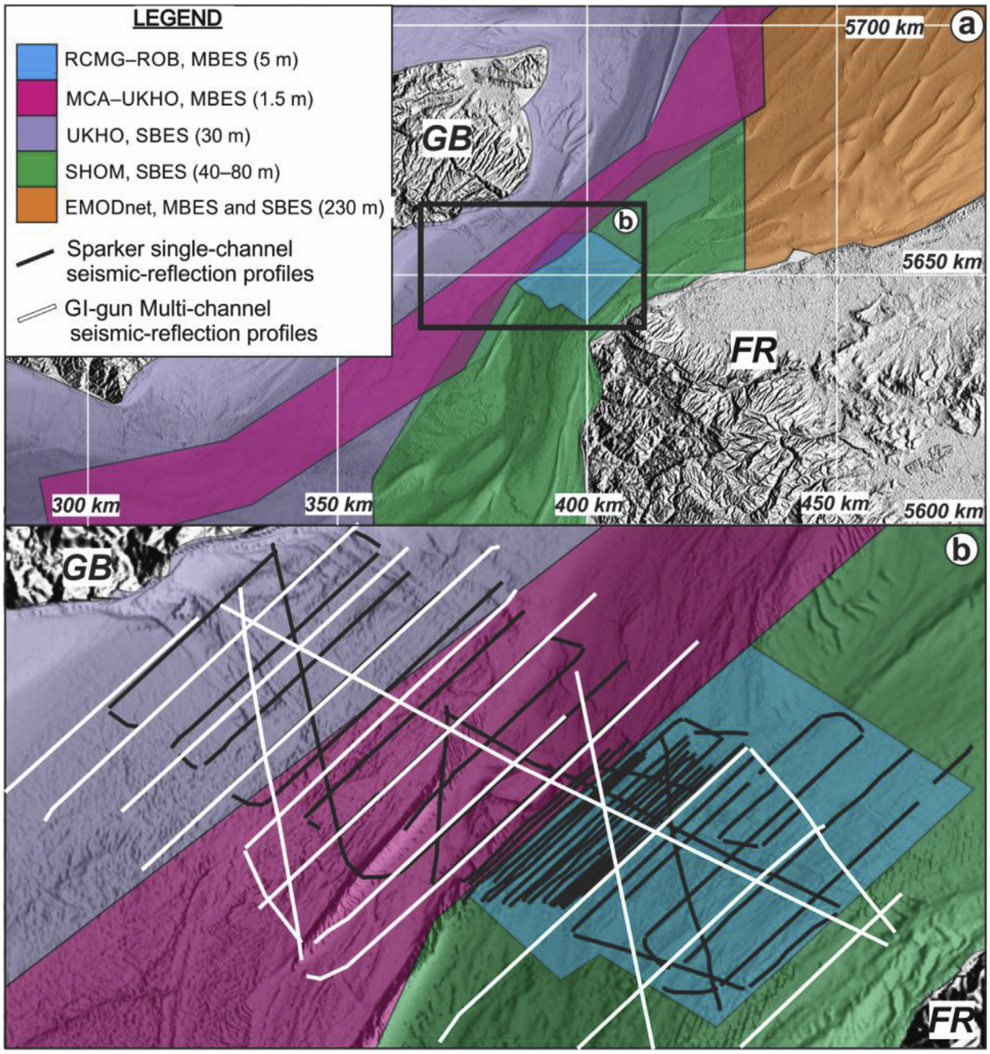

Fig. 3. Coverage of bathymetric and seismic-reflection datasets (see Table 3). After García-Moreno et al. (2015) and Gupta et al. (2017). Projection: UTM, WGS84, zone 31N.

(e.g., Hamblin et al., 1992; Aldiss et al., 2012; García-Moreno et al., 2015; Mortimore and James, 2015). We have however not been able to map the offshore extents of the outcrops of individual formations of the Wealden Group and White Chalk Subgroup because our seismic dataset does not cover their outcrops (see Fig. 3). In addition, most of the formations composing these lithostratigraphic groups have never been properly mapped offshore (see Table 1).

Quaternary deposits are rather localized, being mostly restricted to a few Holocene sand ridges, recent mobile dunes and the infill of the Fosse Dangeard (Figs. 1 and 2; Destombes et al., 1975; Hamblin et al., 1992; James et al., 2002; Reynaud et al., 2003). The nature and absolute age of the sediments infilling the Fosse Dangeard are largely unknown with only one borehole described in the literature (Destombes et al., 1975). This borehole sampled the uppermost 50 m of the sediments infilling the northernmost palaeo-depression forming the Fosse Dangeard (see Fig. 5 for location). It consisted of, from bottom to top: basal conglomerate, alternation of silty clay and fine–medium sand with traces of travertine, and marine sands with pebbles. Based on palynological analyses, Destombes et al. (1975) inferred that these sediments were deposited during the Brørup interstadial (Marine Isotope Stages/MIS 5c; 87–109 ka BP) of the Weichselian glaciation (see Table 2 for equivalences between glacial/interglacial stages, MIS and geological time). However, the lack of absolute dating and additional boreholes means that the age and spatial distribution of this facies are unknown. Moreover, the sediments sampled by Destombes et al. (1975) may only represent the last phase/s of infilling of that palaeo-depression, since the core only penetrated half of the infill.

Initially, the Fosse Dangeard was interpreted as palaeo-depressions formed by glacial erosion during the Saalian glaciation (Destombes et al., 1975). However, this interpretation is not widely accepted today, as more recent studies have demonstrated that the Irish–British ice sheet did not reach this far south during any of the Middle–Late Pleistocene glaciations (e.g., Ehlers and Gibbard, 2004; Lee et al., 2012; Sejrup et al., 2016; Carr et al., 2006).

The formation of the Fosse Dangeard is currently linked to plunge pool erosion generated at the base of waterfalls spilling over the Weald–Artois ridge from a proglacial lake, which purportedly formed in the southern North Sea basin during a marine lowstand (see previous section of this paper; Smith, 1985; Gupta et al., 2017). According to recent investigations, waterfall recession eventually led to the breaching of the Weald–Artois ridge, thus opening the Dover Strait (Gupta et al., 2017). Based on indirect sedimentary data and palaeogeographic reconstructions of northwestern Europe, present consensus holds that this occurred during the Elsterian

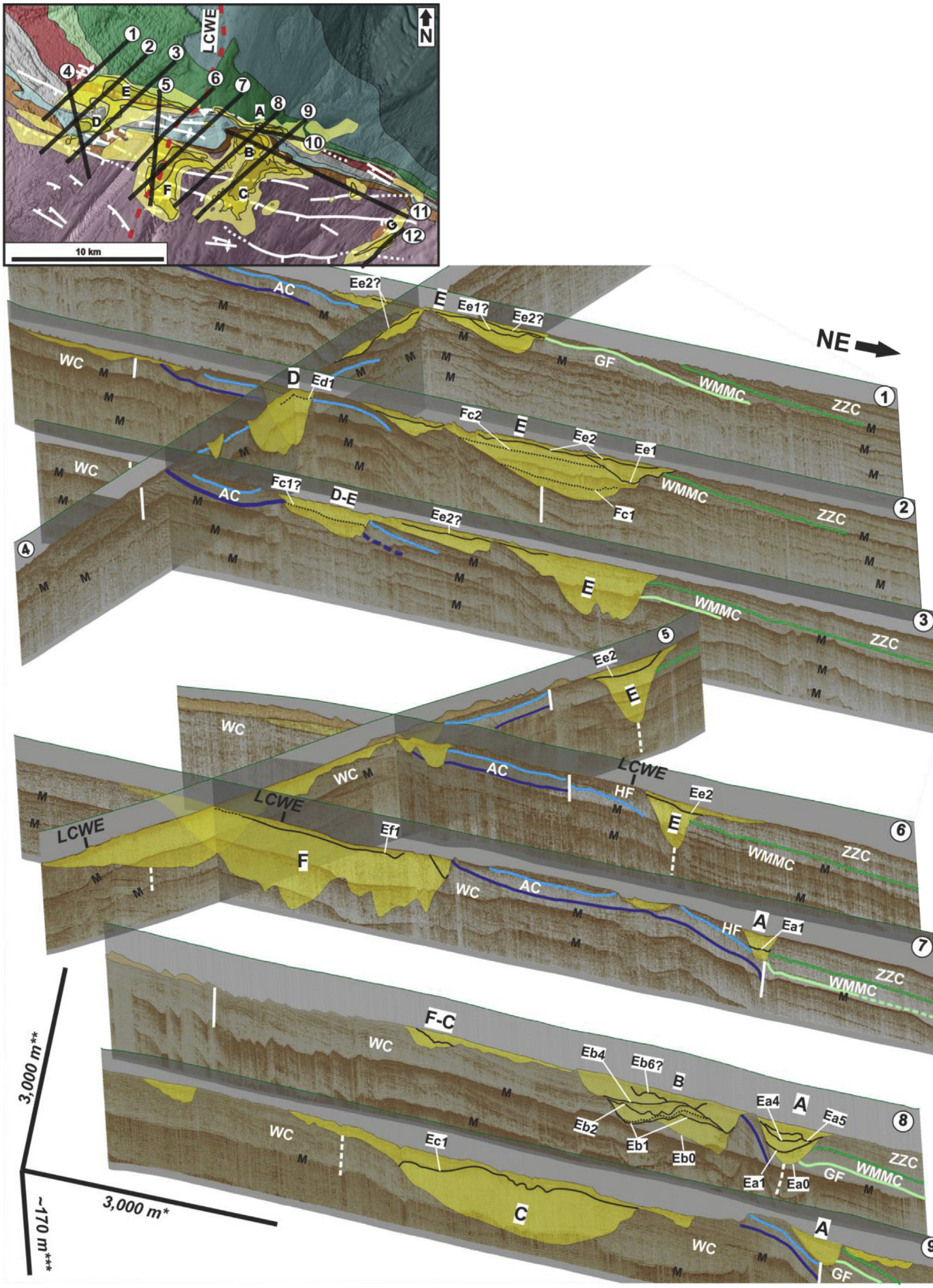

Fig. 4. 3D plot of nine interpreted seismic reflection profiles traversing the Fosse Dangeard (see location in inset map). Inset map: Geologic/structural map of the central Dover Strait (see Fig. 5). Yellow colour: Fosse Dangeard infill; black continuous lines: internal erosional surfaces (Eb1, etc.); black dotted lines: seismic horizons marking changes in seismic facies (e.g. Fc1 and Fc2); dark blue line: base Atherfield Clay Formation; light blue line: base Hythe Formation; light green line: base West Melbury Marly Chalk Formation; dark green line: base Zig Zag Chalk Formation. For other stratigraphic annotations see Table 1. M: seabed multiples; thick white lines: faults offsetting bedrock units and inferred from strata geometry (dashed lines); Dashed red line in inset: western edge of the Lobourg Channel (LCWE).*Distance measured along the orientation of the seismic profiles; ** distance measured across the orientation of the SW–NE seismic profiles; ***depth below the seafloor surface calculated assuming a mean seismic velocity of 2000 ms^{-1}.

glacial maximum, i.e., ~450 ka (Gibbard, 1995; Toucanne et al., 2009a; Gibbard and Cohen, 2015; Gibbard and Lewin, 2016; Gupta et al., 2017).

Since the opening of the Dover Strait, the study area has been subjected to several episodes of intense submarine and subaerial erosion (Hamblin et al., 1992; Gibbard, 1995; Ehlers and Gibbard, 2004; Gibbard and Cohen, 2015). Erosion appears to have been especially intense during Saalian and Weichselian glacial maxima (e.g., Gibbard and Cohen, 2015), when the Irish–British and Fennoscandian ice sheets once again merged across the central and northern North Sea (Ehlers and Gibbard, 2004; Graham et al., 2007; Lee et al., 2012; Sejrup et al., 2016; Carr et al., 2006). The merged ice sheet blocked the northern routes of several European drainage systems (e.g., the Rhine, Meuse, Thames, Weser, etc.), diverting them toward the Dover Strait (Gibbard, 1995). As well as river waters, this drainage system also acted as the conduit for meltwater runoff from a significant part of the ice complex (Gibbard, 1995; Gibbard and Cohen, 2015; Toucanne et al., 2015; Sejrup et al., 2016; Patton et al., 2017).

Several studies have suggested that the southern North Sea – English Channel drainage system that formed during Saalian and Weichselian glacial maxima also comprised major lakes in its catchment, although more localized than the one formed during the Elsterian glaciation (e.g., Busschers et al., 2008; Gibbard and Cohen, 2015; Sejrup et al., 2016). The discharges of such drainage systems, sensitive to periodic meltwater injections triggered by ice sheet fluctuations, lake-outburst floods, etc. is believed to have induced intense fluvial and flood erosion in the Dover Strait (e.g. Gupta et al., 2007; Collier et al., 2015; Gibbard and Cohen, 2015; Gupta et al., 2017). The occurrence of high-magnitude flood flows and/or massive fluvial discharges in the English Channel during the Elsterian, Saalian and Weichselian glacial maxima is consistent with sedimentary data collected from the Bay of Biscay (Toucanne et al., 2009a, 2009b, 2010, 2015). The opening of the Dover Strait and subsequent fluvial and flood erosion during Middle–Late Pleistocene lowstands are currently considered as the main erosional processes that carved the Lobourg Channel (Smith, 1985; Gupta et al., 2007, 2017; Mellett et al., 2013; Collier et al., 2015).

Each Middle–Late Pleistocene glacial stage was followed by an interglacial (see Table 2), during which sea level reached highstands similar or slightly higher than present (e.g. Meijer and Preece, 1995; Turner, 2000; Cohen et al., 2014). The erosional/depositional settings during pre-Holocene interglacial stages are largely unknown. Nevertheless, tidal and coastal erosion during the

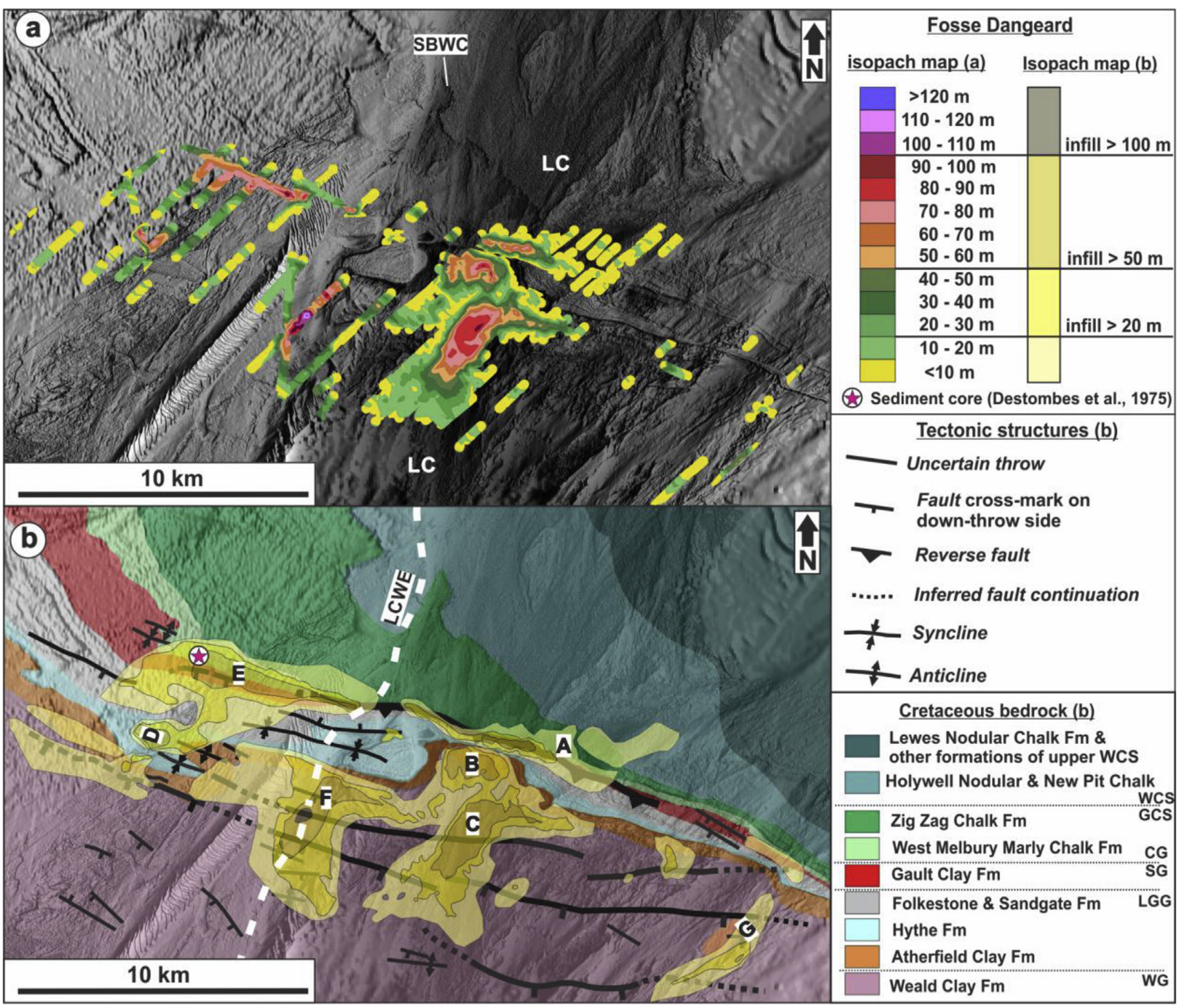

Fig. 5. Relationship between the Fosse Dangeard and bedrock formations. a) Isopach map of the Fosse Dangeard based on seismic reflection data superimposed on a bathymetric map. b) simplified version of isopach map shown in (a) superimposed on the geological/structural map (modified from García-Moreno et al., 2015). Labels A-G mark individual palaeo-depressions. Note that Fosse E, Fosse D, and part of Fosse F are carved outside the Lobourg Channel. LCWE (white dashed line): Lobourg Channel western edge; LC: Lobourg Channel; SBWC: scarp formed at the base of the Holywell Nodular Chalk Formation (White Chalk Subgroup). For other stratigraphic annotations see Table 1.

various highstands that followed the initial opening of the Dover Strait may have significantly contributed to widening the Strait and shaping the morphology of its seafloor. Currently, the Dover Strait is subject to coastal erosion and strong NE–SW-oriented tidal currents along its seafloor. These currents appear to have washed out sediments deposited during previous glacial/interglacial stages and have precluded deposition in most of this area (Hamblin et al., 1992; Reynaud et al., 2003).

3. Data and methods

In this study, we combine 2D seismic-reflection data with

Table 1
Comparison between onshore and offshore recognition of bedrock formations outcropping in the Dover Strait region. Modified from Hamblin et al. (1992)[1]; Hopson (2005)[2]; Hopson et al. (2008)[3]; Aldiss et al. (2012)[4]; Radley and Allen (2012)[5]; García–Moreno et al. (2015)[6]; and Mortimore and James (2015)[7].

Group	Sub-group	Formation	Composition	Age	Submarine Dover Strait
Chalk (CG)	White Chalk (WCS)	Portsdown Chalk	Chalk with marl seams and flint bands	Late Cretaceous	
		Culver Chalk	Soft chalk with flint seams		
		Newhaven Chalk	Soft to medium hard chalk with marl seams, flint bands and phosphatic chalk		Previously known as Upper Chalk Formation (Fm)[2]. The extents of the outcrops of individual formations are not defined offshore.
		Seaford Chalk	Hard chalk with nodular and tabular flint seams		
		Lewes Nodular Chalk	Hard to very hard nodular chalk with interbedded soft to medium hard chalk and marls		
		New Pit Chalk (NPC)	Moderately hard chalk with marls and flints (upper part)		Previously known as Middle Chalk Fm. Boundary NPC–HNC not defined offshore. Base of HNC forms a scarp in seafloor[6].
		Holywell Nodular Chalk (HNC)	Hard nodular chalk with some marls and shell debris. Base: thin marl-chalk succession (Plenus Marls Mm) overlain by very hard chalk/limestone (Melbourn Rock).[4]		
	Grey Chalk (GCS)	Zig Zag Chalk (ZZC)	Pale-grey blocky chalk with marls and marly chalk at its lower part		
		West Melbury Marly Chalk (WMMC)	Dark grey, chalky marl. Base: thin sandy marl unit with glauconite (Glauconitic marl Mm)		Previously known as Lower Chalk Fm[2]. Base of Zig Zag Chalk: seismic marker in seismic reflection data[1,7]
Selborne (SG)		Upper Greensand	Sand and sandstone, silty, glauconitic and shelly	Early Cretaceous	Upper Greensand units are undistinguishable from Gault Formation[1]
		Gault (GF)	Clay or mudstone, glauconitic in part, with a sandy base		
Lower Greensand (LGG)		Folkestone (FF)	Cross-bedded sands and weakly cemented sandstones		
		Sandgate (SF)	Sands, silts, silty clays and soft sandstones		
		Hythe (HF)	Alternating sandy limestones and glauconitic sandy mudstones		Distinct facies in seismic reflection data. Hythe Formation forms a prominent ridge in the seafloor across the Strait[6]
		Atherfield Clay (AC)	Sandy mudstone		
Wealden (WG)		Weald Clay (WC)	Mudstones and shales		
		Turbridge Wells Sand	Interbedded sandstones, siltstones and shales		
		Wadhurs Clay	Shales and mudstones		Significant lateral facies and thickness variations[1,3,5]. Apart from the WC, the extents of the outcrops of these formations are not defined offshore.
		Ashdown	Sandstones and siltstones		
Purbeck (PuG)		Durlston	Interbedded, shelly limestones and mudstones		
		Lulworth	Limestones, marls, evaporites and some sands.	Late Jurassic	
Portland (PoG)		Portland Sand	Siltstones, sandstones and mudstones with calcite and dolomite		
		Portland Stone	Calcareous and glauconitic sandstone		Portland formations more clayey than onshore[1,3]; undistinguishable from the Kimmeridge Clay Formation[1]
Ancholme (AG)		Kimmeridge Clay	Organic rich mudstone with occasional hard, thin carbonate-cemented horizons		

Table 2
Land-based chronostratigraphic terminology used for major Quaternary glacial and interglacial stages in northwestern Europe and their correlation with Marine Isotope Stages (MIS), absolute timing and geological timescale (Gibbard and Cohen, 2008; Sejrup et al., 2016*). Time intervals during which Fennoscandian and Irish–British ice sheets merged across the North Sea (i.e. glacial maxima) are indicated.

Glacial/interglacial Stages		Inter/Glacial	MIS	ka BP	Ice sheets merged across North Sea	Series	Subseries
Northern Europe	Great Britain						
	Flandrian	Intergl.	1	Present – 11.7		Holocene	
Weichselian	Devensian	Glacial	2 – 5d	11.7–110	30–19* ka BP (MIS 2); ~70 ka BP (MIS 4)?		Late
Eemian	Ipswichian	Intergl.	5e	110–130			
Saalian	Wolstonian	Glacial	6–10	130–374	175–155 ka BP; 150–140 ka BP (both MIS 6)	Pleistocene	
Holsteinian	Hoxnian	Intergl.	11	374–424			Middle
Elsterian	Anglian	Glacial	12	424–478	~450 ka BP		

single- and multi-beam bathymetric data to interpret the 3D morphologies, infills and interrelationship of the Fosse Dangeard and Lobourg Channel. The various datasets used in this study are the same that were used in García-Moreno et al. (2015) and Gupta et al. (2017). The novelty of the present study is not the database, but its more detailed interpretation.

Technical details about the acquisition and processing of the various datasets used are given in the Method sections and supplementary documents of García-Moreno et al. (2015) and Gupta et al. (2017). In this section, we will therefore only summarize their main characteristics, which are listed in Table 3. The coverage of the various bathymetric datasets and seismic reflection surveys are shown in Fig. 3.

The analysis was carried out by combining 2D and 3D seismic interpretation using OpendTect and IHS Kingdom software, with 3D geomorphologic analysis performed using Global Mapper and ArcMap. The combined interpretation has resulted in detailed bathymetric maps of the Dover Strait's seafloor and isopach maps of the Fosse Dangeard. These maps are more complete and accurate than those published in previous studies. Isopach maps were built by assuming a mean seismic velocity through the palaeo-depressions' infills of 1800 m s^{-1} (see Arthur et al., 1997) and subtracting any contributions from Holocene sandbanks and dunes. Note that dip angles provided in this paper refer to average apparent dips measured from the seismic reflection data assuming this same average velocity.

4. The Fosse Dangeard: geometry and seismic stratigraphy

The Fosse Dangeard is a set of sediment-infilled depressions that are eroded into bedrock in the central part of the Dover Strait along a WNW–ESE-elongated area of ~312 km^2 (Fig. 5b). It comprises 7 major interconnected palaeo-depressions with maximum depths ranging between 50 m and 140 m, and several scattered minor sediment-filled erosional depressions with depths ≤20 m (Fig. 5b). For ease of reference we name individual palaeo-depressions Fosse A–G. Details about the morphologies and infills of individual palaeo-depressions are listed in Tables 4 and 5.

The basal erosional surfaces of individual palaeo-depressions are easily recognisable in the seismic reflection data because of the angular unconformity that they form with sub-cropping bedrock (Fig. 4). Their morphology and seismic facies are also strikingly different from those of the bedrock formations. The isopach map resulting from the mapping of the basal erosional

Table 3
Geophysical datasets available for the present study. See García-Moreno et al. (2015) and Gupta et al. (2017) for details on their acquisition and processing. Lille U.: Lille University; RCMG: Renard Centre of Marine Geology (Ghent University); ROB: Royal Observatory of Belgium; MCA: British Maritime & Coastguard Agency; UKHO: United Kingdom Hydrographic Office; SHOM: Service Hydrographique et Océanographique de la Marine; IC: Imperial College London; SC: Single-channel seismic-reflection data; MC: Multi-channel seismic-reflection data; SBES: single-beam bathymetric data; MBES: Multibeam bathymetric data.

Acquired/provided by	Year	Type of data	Processed at	Gridded at (m)	Vertical resolution (m)	Maximum penetration (m)
Lille U. – RCMG	2002	SC	ROB	–	1–3	80–100
ROB –RCMG	2010–2012	SC & MC	ROB & RCMG	–	SC: 1–3; MC: 5–10	SC: 100–150 MC: ~250
ROB –RCMG	2010–2012	MBES	RCMG	5	–	–
MCA –UKHO	2006–2007	MBES	MCA – IC	1.5	–	–
UKHO	1988–2004	SBES	MCA – IC	30	–	–
SHOM	Since 1970's	SBES	Lille U.	40 and 80	–	–
EMODnet	1946–2017	SBES & MBES	Multiple Institutions	230	–	–

Table 4
Geometry and erosional pattern of the palaeo-depressions composing the Fosse Dangeard. *Observed changes of facies that might not be linked to an erosional event, but to a change of sedimentary setting. See Table 1 for lithostratigraphic abbreviations. BES: basal erosional surface of palaeo-depressions composing the Fosse Dangeard; sub-hor.: sub-horizontally stratified.

	Fosse A	Fosse B	Fosse C	Fosse D	Fosse E	Fosse F	Fosse G
Within Lobourg Channel?	Yes	Yes	Yes	No	No	Partially	No
NE-SW cross-sectional morphology	Scoop-shape	Scoop-shape	Scoop-shape	Scoop-shape to semi-circular	Scoop-shape	Scoop-shape	Scoop-shape
Dip of NE slope of BES	22°–30°SW	10°–22°SW	7°SW	30°SW	7°SW	17°SW	7°SW
Dip of SW slope of BES	7°–23°NE	10°NE	5°NE	15°NE	3°NE	9°NE	5°NE
Maximum depth (m)	80	100	100	90	100	140	60
Length WNW–ESE axis/NE–SW axis (km)	~7/0.3–0.8	~2.6/1.5–1.8	3.5–5.9/6.3–7.3	~2.5/0.9–1.5	10–11/2–4.5	3–5.8/1.5–5.5	0.6–1.5/~5.3
Sub-cropping bedrock: composition	ZZCF, WMMF, GF, FF, SF and HF	WC	WC	HF, AF and WC	WMMF, GF, FF, SF and HF	WC	WC
Sub-cropping bedrock: structure	Major reverse fault; NE-dipping HF and sub-hor. ZZCF	No faults; sub-hor. strata	Two faults; sub-hor. strata	No faults; sub-hor. strata	One fault. NE-dipping GF, FF, SF and HF; sub-hor. WMMF	Two faults; sub-hor. strata	Possible faults; both sub-hor. and NE-dipping strata
Internal erosional surfaces	5	6	1	1*	2	1	1*
Changes of facies within same unit		1			2		

Table 5
Main characteristics of seismic facies encompassed between the various internal erosional surfaces and prominent horizons marking changes in seismic facies of the infills of individual palaeo-depressions composing the Fosse Dangeard.

Fosse A	Ea0–Ea1 Acoustically almost transparent facies; moderate-amplitude reflections in its upper part	Ea1–Ea2 Low-amplitude reflections to transparent seismic facies	Ea2–Ea3 Well-defined, subparallel high-amplitude reflections	Ea3–Ea4 Acoustically almost transparent facies; moderate-amplitude reflections in its upper part	Ea4–Ea5 Low amplitude reflections to seismically transparent seismic facies		Ea5–seafloor Diffuse reflections
Fosse B	Eb0–Eb1 Acoustically almost transparent facies	Eb1–Eb2 Moderate-amplitude reflections. Base: high-amplitude reflections.	Eb2–Eb3 Discontinuous moderate to high-amplitude reflections	Eb3–Eb4 Discontinuous moderate to high-amplitude reflections	Eb4–Eb5 Diffuse reflections with localized discontinuous reflections	Eb5–Eb6 Diffuse reflections with localized discontinuous reflections	Eb6–seafloor Diffuse reflections with localized moderate-amplitude reflections
Fosse C	Ec0–Ec1/Eb6 Acoustically almost transparent facies, presenting some low-amplitude discontinuous reflections (chaotic in its southwestern half; sub-horizontal to the northeast)						Ec1/Eb6–seafloor Discontinuous low-amplitude reflections
Fosse D	Ed0–Ed1 Subparallel low-amplitude reflections						Ed1–seafloor Diffuse reflections
Fosse E	Ee0–Fc1 High-amplitude reflections	Fc1–Fc2 Discontinuous to chaotic sub-horizontal reflections		Fc2–Ee1/seafloor Low-amplitude sub-horizontal reflections		Ee1–Ee2 Moderate-amplitude cross-bedded reflections	Ee2–seafloor Diffuse reflections
Fosse F	Ef0–Ef1 Acoustically almost transparent facies, presenting some low-amplitude discontinuous reflections						Ef1–seafloor Diffuse reflections with localized moderate-amplitude reflections
Fosse G	Eg0–Eg1 Acoustically almost transparent facies						Eg1–seafloor High-amplitude horizontal reflections

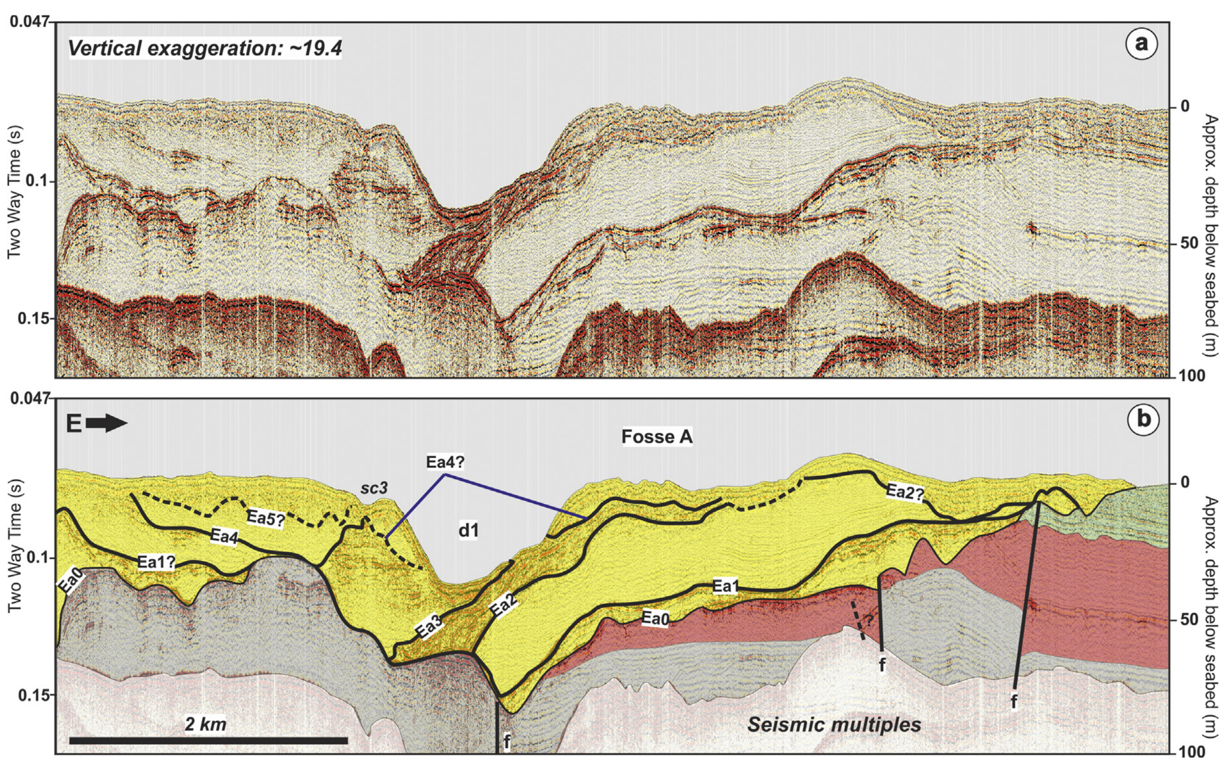

Fig. 6. High-resolution single-channel seismic-reflection profile (a) and interpretation (b) along Fosse A main axis (line 10 in Fig. 4). Yellow area: Fosse A infill; other coloured areas: bedrock geology (see legend in Fig. 5b); f: reverse fault crossed by the seismic profile at 3 different locations; black lines within Fosse A: internal erosional surfaces; sc3: longitudinal scour (see Fig. 15); d1: depression excavated into Fosse A's infill. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

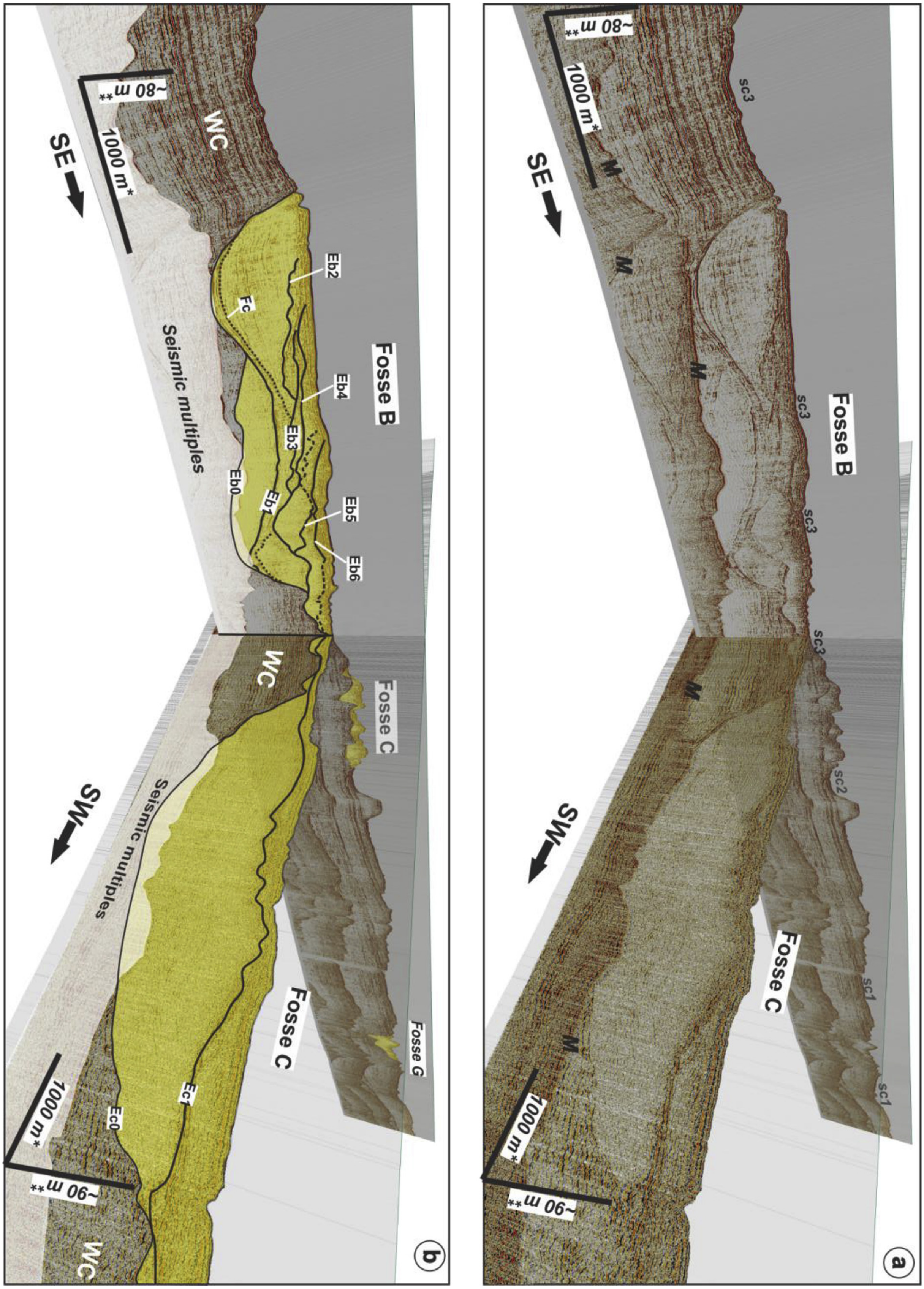

Fig. 7. 3D plot of cross-cutting single-channel seismic reflection profiles 9 and 11 (see Fig. 4 for location). Note that internal erosional surface Eb6 of Fosse B correlates with Ec1 in Fosse C. Note also the absence of other internal erosional surfaces in Fosse C. Yellow area: infills of Fosse B and C; Fc: horizon marking a change of facies within a seismic unit; WC: Weald Clay Formation; sc1, sc2 and sc3: longitudinal scours (see Fig. 11 and 15); M: seismic multiples. See Fig. 4 for other labels. *Distance measured along the different seismic profiles' orientation; **apparent differences in the sizes of depth bars are due to 3D projection.

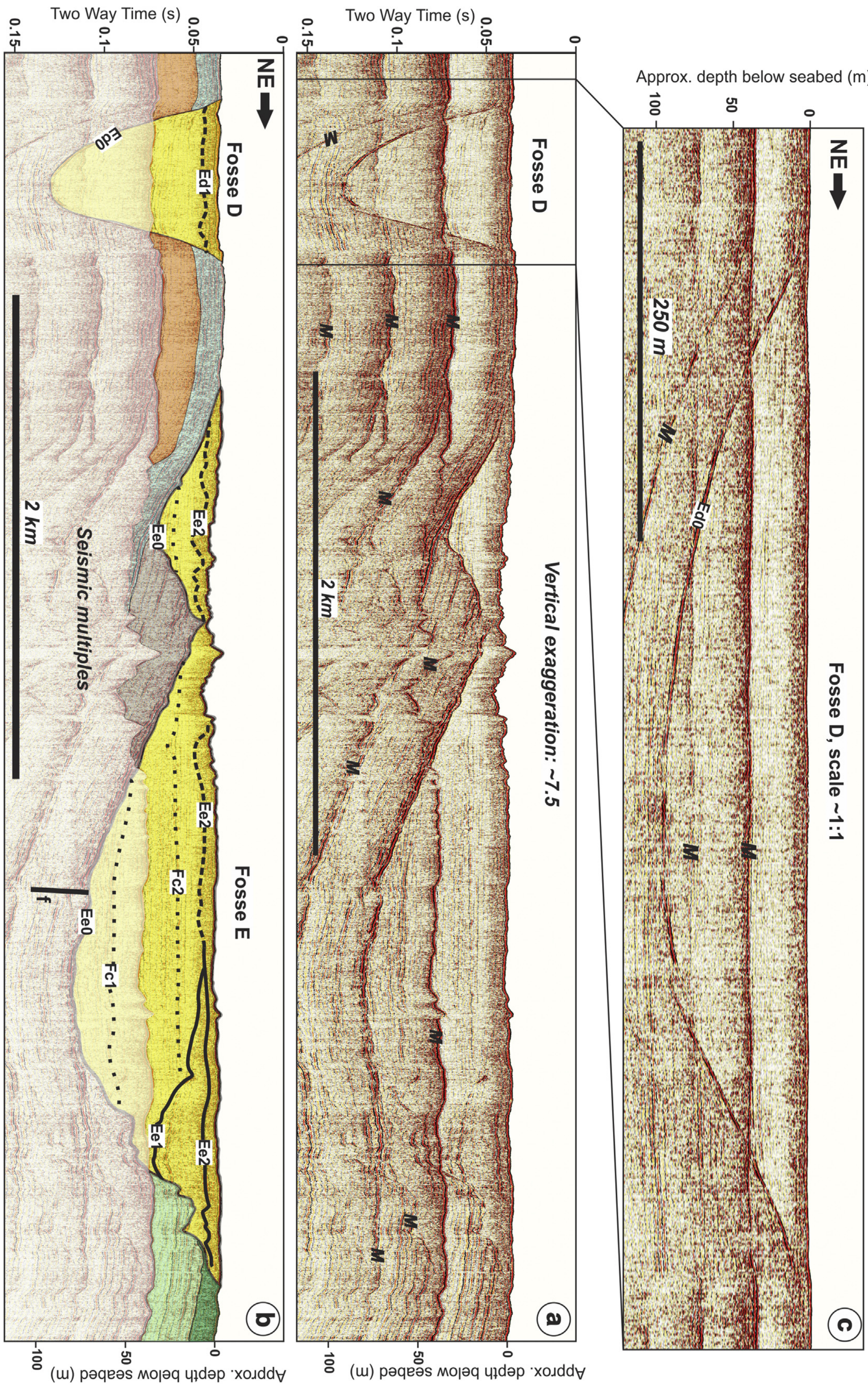

Fig. 8. High-resolution single-channel seismic-reflection profile (a) and interpretation (b) across Fosse D and E (line 2 in Fig. 4). (c) Zoom on Fosse D. Note that the bedrock units underneath Fosse D are sub-horizontally stratified. Note also the absence of bedrock faults below Fosse D. Continuous/dashed lines within yellow area: internal erosional surfaces. Dotted lines mark changes of seismic facies within seismic units. See Figs. 4 and 5 for other labels and meaning of coloured areas. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

surfaces along the seismic reflection profiles permits the interpretation of the Fosse Dangeard morphology in plan view (see Figs. 4 and 5).

The Fosse Dangeard extends subparallel to the WNW–ESE-strike of the outcropping bedrock, i.e., sub-parallel to the orientation of the axis of the Weald-Artois anticline (Fig. 2). The northwestern edges of individual palaeo-depressions are either defined by outcropping Zig Zag Chalk Formation (Fosse A and E) or by outcropping Hythe and/or Atherfield formations (Fosse B, C, D and F), whereas their southwestern terminations tend to be rounder and more irregular (Fig. 5b).

Comparison of the spatial distribution of the Fosse Dangeard with a bathymetric map of the Dover Strait shows that some of the deepest palaeo-depressions are carved either completely or partially outside the Lobourg Channel (Fig. 5). In fact, the Lobourg Channel, together with several scours and depressions associated with its incision (see next section), is incised into the uppermost infills of Fosse A, B, C and part of Fosse F, attesting to its younger age of formation (Figs. 6, 7 and 9).

Seismic data show several faults offsetting and locally deforming the bedrock sub-cropping beneath Fosse A, C, E, F and, possibly, Fosse G (see Figs. 6 and 8–10). Fosse A, for example, is carved along the plane of a major reverse fault, juxtaposing the folded Hythe Formation and the sub-horizontally stratified Zig Zag Chalk Formation. This palaeo-depression is also mainly incised into the apparently less resistant bedrock units encompassed between these two formations (Fig. 4). Hence, differential erosion due to local variations of hard/soft beds and the presence of tectonic structures (for example, fractures and faults) may have had a role in determining the morphology of Fosse A.

Fosse E is also primarily incised between the folded Hythe Formation and the sub-horizontally stratified West Melbury Marly Chalk and Zig Zag Chalk formations (Fig. 8). This palaeo-depression is shallower across the Hythe Formation and the formations composing the Grey Chalk Subgroup than across the relatively less resistant Sandgate, Folkestone and Gault formations. Note also that the dip of the southwestern slope of Fosse E appears to be controlled by the dip of the sub-cropping bedrock. Therefore, differential erosion due to lateral variations of rock type may have contributed to determining the morphology of Fosse E too. However, bedrock faulting appears to have had little, if any, control on the pattern of erosion. The only fault identified in the bedrock incised by Fosse E is indeed very localized (Fig. 8).

The morphologies of Fosse B, C, D, F and G appear, however, to be independent of rock type and tectonic structures into which they are carved. This is evidenced, firstly, by the fact that Fosse B and D are carved into sub-horizontally stratified structureless substratum (Figs. 7 and 8). Secondly, Fosse C, F and G show elongation across the strike of the faults rather than along them (Figs. 4 and 5b). Finally, the connection between Fosse C and F occur through an unfaulted area instead of along the plane of a fault (Fig. 5b).

The seismic reflection data show that the infill of the Fosse Dangeard is partitioned by several internal erosional surfaces (Fig. 4). We label these surfaces as Ea1, Ea2, Eb1, Eb2, etc., where “a” refers to internal erosional surfaces mapped in Fosse A, “b” to those mapped in Fosse B, etc., and the numbers are in order from deep to shallow. Internal Erosional surfaces are characterised by higher reflection amplitudes than overlying and underlying seismic facies and form angular unconformities with underlying reflectors (e.g. Fig. 6). These surfaces present a range of cross-sectional morphologies, separating units with different seismic facies (Fig. 4; see also Table 5). The number of internal erosional surfaces and seismic facies varies from one palaeo-depression to another. Fosse A and B contain the most, with 5 and 6 internal erosional surfaces respectably (Figs. 6 and 7). The rest of the palaeo-depressions present a maximum of 2 (see Table 4).

In general, internal erosional surfaces within Fosse A show scoop-shaped geometries along NE–SW cross sections, and more irregular geometries along WNW–ESE orientations (Figs. 4 and 6). Erosional surface Ea1 extends across the whole of Fosse A, completely truncating the sedimentary package between Ea0 and Ea1 in places (Fig. 6). The remainder of the internal erosional surfaces are, on the other hand, more localized.

The most striking internal erosional surfaces within Fosse B are Eb1 and Eb4. These surfaces truncate all underlying units and extend over the entire infill of that palaeo-depression. They show sinusoidal cross-sectional morphologies, incising more in the southwest than in the northeast (Figs. 4 and 7). Internal erosional surface Eb6 (Ec1 in Fosse C) also appears to have eroded large parts of Fosse B's older infill. This erosional surface is, however, better defined in Fosse C, where it extends over the entire palaeo-depression as a relatively planar surface (Fig. 7). Importantly, Eb6/Ec1 is the only internal erosional surface that can be confidently correlated between Fosse B and C (Fig. 7). The geometries of other internal erosional surfaces dividing Fosse B's infill are less well defined in the seismic reflection data, as significant parts of them have been eroded by one another and/or by Eb1, Eb4 and Eb6.

Internal erosional surfaces identified in the infills of Fosse C, D, E, F and G are mostly located in the uppermost part of their infills (i.e., near the seafloor). Generally, internal erosional surfaces in these palaeo-depressions show almost-sub-horizontal morphologies. Only Ee1 and part of Ef1 show a different geometry. The latter two surfaces are highly localised, extending along the northeastern edges of Fosse E and F respectively, and exhibiting channel-like cup-shape morphologies.

Seismic facies within the infill of the Fosse Dangeard do not only change across internal erosional surfaces. Some changes in facies occur across surfaces that do not form angular unconformities with underlying infill units, suggesting that these record sedimentary facies transitions. These surfaces are labelled Fc1, Fc2, etc., where the numbers are in order from deep to shallow. Clear examples of this are Fc1 and Fc2 in Fosse E, and Fc in Fosse B (see Figs. 7 and 8; Table 5).

5. The Lobourg Channel

The Lobourg Channel is the most striking geomorphological feature imprinted on the seafloor of the Dover Strait. It is a 120-km-long palaeovalley that shows up to 30 m of relief at the seabed and it is characterised by relatively straight edges, oriented NE–SW in the north and NNE–SSW in the south (Fig. 11). On its floor, this palaeovalley exhibits a series of kilometre-scale, NE–SW/NNE–SSW-elongated scours, inner channels, linear escarpments and other erosional features incised several metres to tens-of-metres into bedrock (Fig. 11b and Fig.15b).

Scours carved in the seafloor within the Lobourg Channel are labelled sc1, sc2 and sc3, where the numbers refer to the erosional phases within which they were carved; e.g., sc1 indicates that the formation of that scour is likely linked to the valley incision represented by escarpment E1. Labels d1 and d2 designate unfilled or partially infilled depressions and Ch1 and Ch2 refers to prominent channels carved within the Lobourg Channel (see Gupta et al., 2017). The Lobourg Channel also contains a number of NE–SW/NNE–SSW-elongated, kilometre-scale teardrop-shaped and ellipsoidal platforms several metres in height, which are carved into bedrock (Figures 11 and 15). In this paper, we refer to them as streamlined islands after the classification shown in Collier et al. (2015). Streamlined islands are labelled, from north to south: si0, si1, si1b, si2 and si3. The morphology across the Dover Strait is described in more detail below.

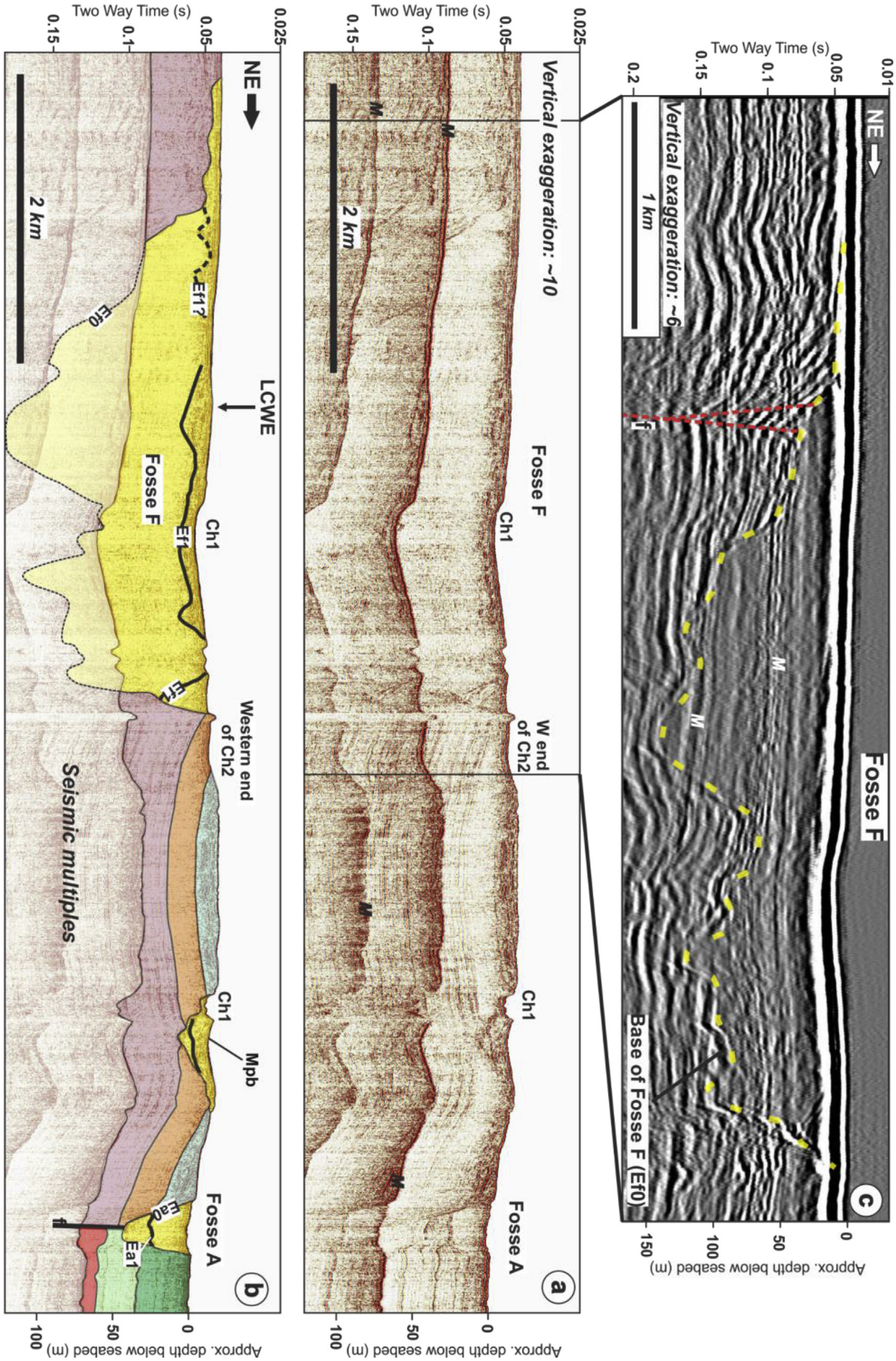

Fig. 9. High-resolution single-channel seismic-reflection profile (a) and interpretation (b) across Fosse F and A (line 7 in Fig. 4). (c) Zoom on Fosse F extracted from a multi-channel deeper-penetration seismic-reflection profile acquired parallel to (and ~1 km to the north of) seismic-reflection profile 7. Ef1 and Ea1: internal erosional surfaces; Ch1 and Ch2: amphitheatre-head channels (see Fig. 15). The western edge of the Lobourg Channel (LCWE) is indicated. See Figs. 4 and 5 for other labels and meaning of coloured areas.

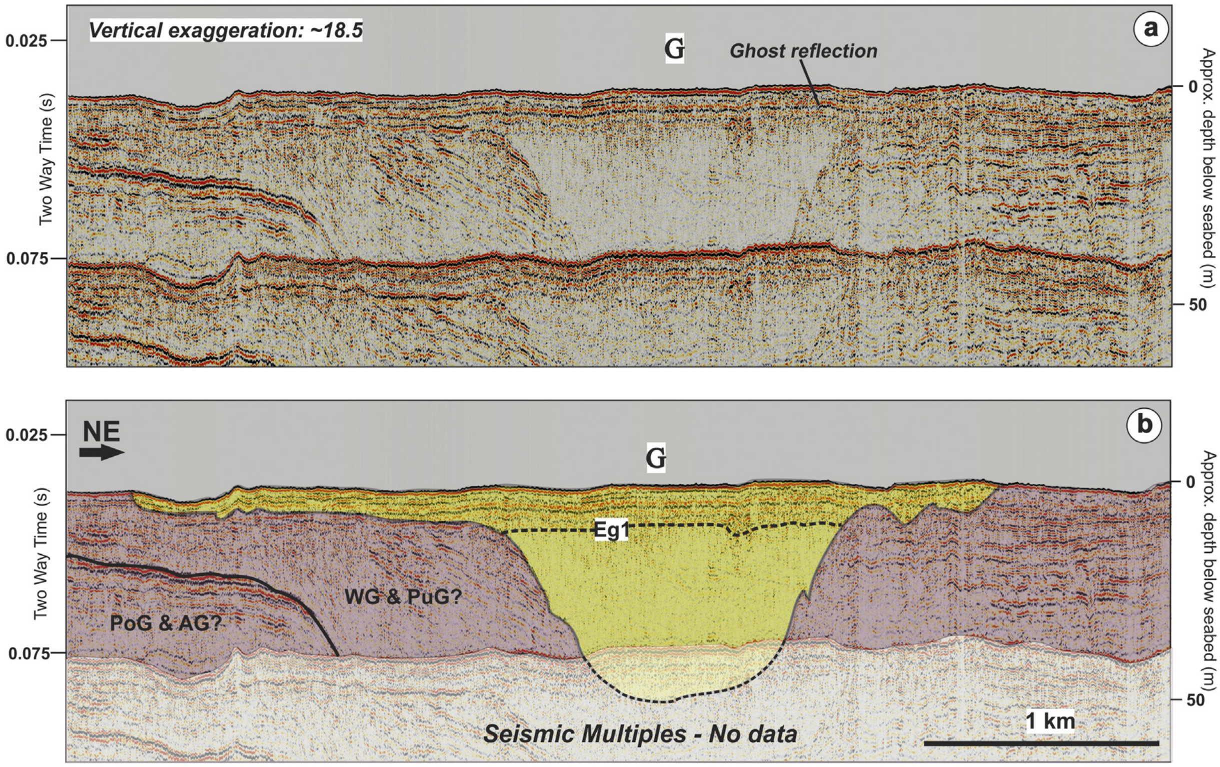

Fig. 10. High-resolution single-channel seismic-reflection profile (a) and interpretation (b) along Fosse G (line 12 in Fig. 4). Eg1: possible internal erosional surface; ghost reflection: seafloor repetition near the surface (artefact). WG & PuG: undefined units of Wealden Group and, possibly, Purbeck Group; PoG & AG: undefined units of Portland and Ancholme groups.

5.1. Northern Dover Strait

In the northern Dover Strait, the Lobourg Channel consists of a ~25 km wide, NE–SW-oriented palaeovalley, comprising three sharp, subparallel escarpments named E1, E2 and E3 (Fig. 11). Escarpments E1, which forms the eastern edge of the Lobourg Channel, and E2 are 10–15 m high and exhibit NE–SW orientations (Figs. 11 and 12). By contrast, escarpments E3 and the 'Lobourg Channel Western Edge' (LCWE) show, respectively, 5–10 m and 15–25 m of relief at the seafloor. Both E3 and LCWE trend NNE–SSW, running southward obliquely to escarpments E1 and E2. LCWE and E3 delimit a prominent ~3.5 km wide palaeovalley with a box-shaped transverse profile carved into bedrock (Figs. 11 and 12), the eastern edge of which (escarpment E3) truncates escarpments E1 and E2 in the central Dover Strait. In this study, we refer to that palaeovalley as the Lobourg inner channel (Lic).

The Lobourg inner channel defines an almost-straight channel in the northern part of the northern Dover Strait area (Fig. 11). It extends over tens of kilometres, exhibiting similar relief and shape across the various formations composing the Chalk Group (Fig. 2). The width of that palaeovalley is also rather constant in that area. However, it widens significantly at the southern end of streamlined island si0, passing from ~3.5 km to ~7.5 km wide (Fig. 11). Topographic cross-sections across the Lobourg inner channel reveal quasi-symmetrical slopes (dips: ~1.7°) and a flat bottom (Fig. 12).

The northeastern extents and morphologies of escarpments E1 and E2 (especially that of E1) are poorly constrained, since tidal sandbanks and dunes of Holocene age cover large parts of them (Fig. 11). Topographic cross-sections across the four main escarpments (LCWE, E1, E2 and E3) show morphologies similar to strath-terraced fluvial valleys (see Hancock and Anderson, 2002), with all the terraces located on the eastern side of the valley (Fig. 12). Importantly, terrace-like platforms located within the Lobourg Channel are not formed by gravels or sand aggradations; they are carved into bedrock. In this study, we have named the terrace-like platform delimited by escarpments E1 and E2 as platform LC1, while the one defined by E2 and E3 is referred to as platform LC2. Both platforms LC1 and LC2 are traversed by a minor palaeovalley system (i.e. Pvs1), which cuts through escarpments E1 and E2 perpendicularly to their main orientations (Fig. 11). This system is partially covered by sandbanks and dunes, and it appears to have been carved by a small palaeo-river system that discharged into the Lobourg inner channel.

Streamlined islands are only observed on platform LC2 and within the Lobourg inner channel (Fig. 11). These include a ~15 km long, 5 km wide, major streamlined island (si0), and some ~2.5 km long, ~1 km wide, minor ones (e.g., si1 and si1b). All streamlined islands identified in this area show similar morphologies to those identified in chalk outcrops by Collier et al. (2015) in this and other parts of the Channel palaeovalleys. They have distinct teardrop shapes, with their major axis oriented in the direction of the inferred palaeo-flow (i.e., NNE–SSW). Streamlined island si0 is located on platform LC2 and its major axis is subparallel to escarpment E2, suggesting that its formation is associated with the processes that also carved E2. Streamlined island si0 exhibits a series of ENE–WSW-oriented scours carved along its south-eastern edge and front part (Fig. 15). These scours end in the west, at their intersection with escarpment E3, in a 20-m-deep depression (d1) incised into Fosse A's infill (Figs. 6 and 15). Streamlined islands si1 and si1b are located within the Lobourg inner channel, which has similar orientation to their major axis. The formation of streamlined islands si1 and si1b thus appears to be associated with the erosional processes that also formed the Lobourg inner channel.

Other remarkable features observed in the northern Dover Strait

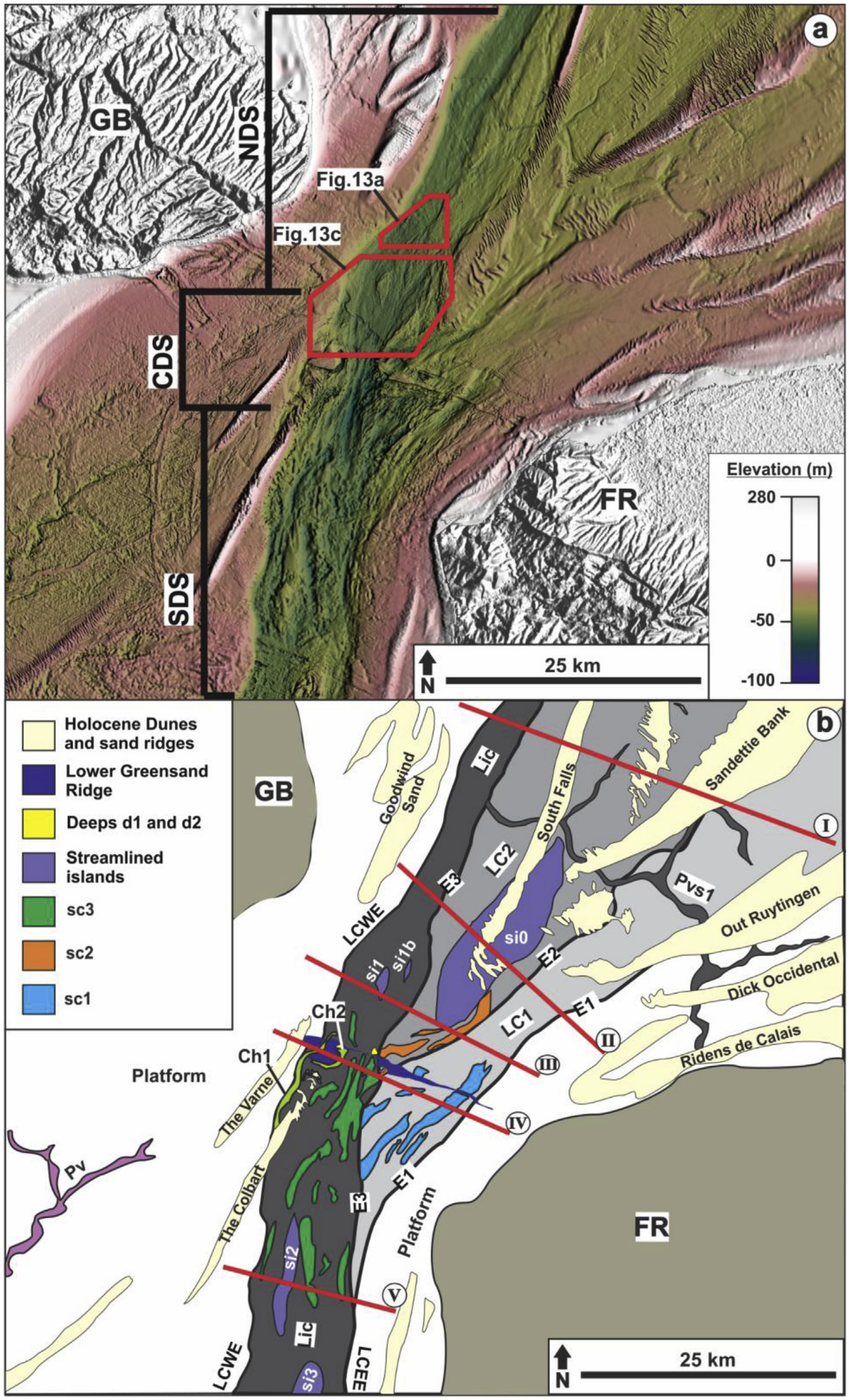

Fig. 11. Bathymetric–topographic map with a 40 m cell size (a) and geomorphological interpretation (b) of the Dover Strait region. The map in (a) is illuminated (light direction altitude: 45°, azimuth: 45°) and has a vertical exaggeration of 7.5. Red areas in (a): areas shown in Fig. 13; black lines in (a): extents of the northern Dover Strait (NDS), central Dover Strait (CDS) and southern Dover Strait (SDS) referred to in the text. Red lines in (b): topographic profiles shown in Fig. 12. Note that escarpment E3 truncates escarpments E1 and E2. Except for si0 and si1b, streamlined islands (si) are labelled after Collier et al. (2015). Major tidal sand ridges are labelled after British Admiralty Nautical Charts. Pvs1: palaeovalley system connected to Lic; Pv: valley network incised into western platform (see Gupta et al., 2017); E1, E2 and E3: escarpments defining the eastern edges of platforms LC1 and LC2, and inner channel 'Lic'; sc1, sc2 and sc3: longitudinal scours associated with valley incisions LC1, LC2 and Lic; LCWE: Lobourg Channel Western Edge; LCEE: Lobourg Channel Eastern Edge. A more detailed interpretation of the central area is given in Fig. 15.

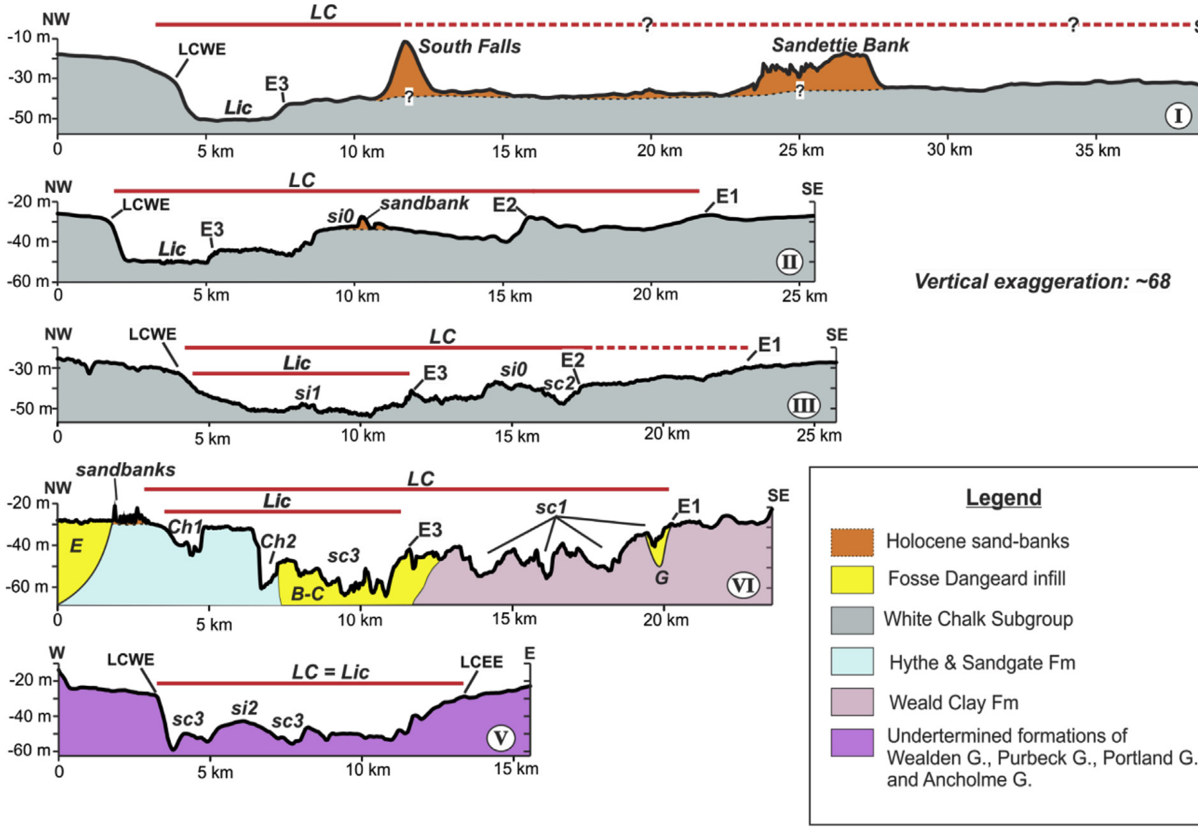

Fig. 12. Geologically interpreted topographic profiles across the Lobourg Channel (see Fig. 11b for location). Fosse B, C, E and G are shown in yellow. LC: interpreted width of the Lobourg Channel; Lic: Lobourg inner channel. See Fig. 11 for other labels. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

are several sets of parallel ridge-and-groove linear bedforms, showing 0.5–1.5 m of relief at the seafloor. In this study, they are referred to as La, Lb and Lc (Figs. 13 and 14). These features are not artefacts as they extend obliquely to ship-tracks, dunes and other Holocene sedimentary bodies. Importantly, sets Lb and Lc are located within the Lobourg inner channel. In contrast, set La is located outside the Lobourg inner Channel and appears to be truncated by the western edge of that palaeovalley (Fig. 13). The ridges and grooves appear to be carved into chalk; however, we have not been able to confirm this, as no seismic line traverses them. These features extend subparallel to the axis of the Lobourg inner channel. The lengths of individual ridges and grooves are unknown, as sand dunes and other younger sedimentary bodies cover large parts of them (Fig. 13). The three sets have slightly different azimuths, ranging between 32° and 36°, 40°–45° and 37°–41° for sets La, Lb and Lc respectively.

5.2. Central Dover Strait

The bedrock underlying the seafloor in the central Dover Strait is characterised by a range of sedimentary formations, going from the lower formations of the Grey Chalk Subgroup to the sediments infilling the Fosse Dangeard (Fig. 5). The central Dover Strait area also coincides with the northern limb of the Weald-Artois anticline and a major fault system that belongs to the regional North Artois Shear Zone (Figs. 2 and 5b).

The Lobourg Channel shows a marked change in orientation in the central Dover Strait (Fig. 11). The Lobourg inner channel and escarpment E1 bend anticlockwise at their path through the outcropping Lower Greensand Group, continuing southwards with nearly north–south orientation. The overall morphology of the Lobourg Channel also experiences a major change from the centre of the Strait southward. Notably, escarpments E2 and E1 are truncated in the central Dover Strait by the Lobourg inner channel, which defines the entire Lobourg Channel in the southern Dover Strait (Fig. 11).

The erosional features carved within the Lobourg Channel change as well from the central Dover Strait southward (see Figs. 11, 13 and 15). Bedrock units outcropping in this area form prominent scarps, ridges and grooves, attesting to differences in the resistance to erosion of the various bedrock units (Figs. 5 and 15). In addition, ridge-and-groove linear features and teardrop-shaped streamlined islands are no longer found to the south of the scarp formed in the seafloor by the base of the Holywell Nodular Chalk Formation (Fig. 15).

The most striking erosional features located in the central Dover Strait are carved within the Lobourg inner channel (Fig. 15). They include a series of NNE–SSW-oriented longitudinal scours (sc3), two deeply incised palaeo-channels (Ch1 and Ch2) and two 20–25 m deep depressions (d1 and d2).

Scours sc3 are kilometre-scale elongated incisions, exhibiting depths at the seafloor of up to 10 m. These features cut through formations of the Selborne Group, Lower Greensand Group and Wealden Group, as well as through the uppermost infills of Fosse A, B and C (Figs. 5 and 15). Ch1 and Ch2 are 0.5–1 km wide, amphitheatre-head palaeo-channels incised along the western half of the Lobourg inner channel. Ch1 is characterized by a linear to slightly sinuous morphology and relief of 10–15 m at the seafloor. This channel cuts through Fosse F's upper infill and the Cretaceous Lower Greensand Group and Wealden Group (Figs. 5 and 15). Ch1

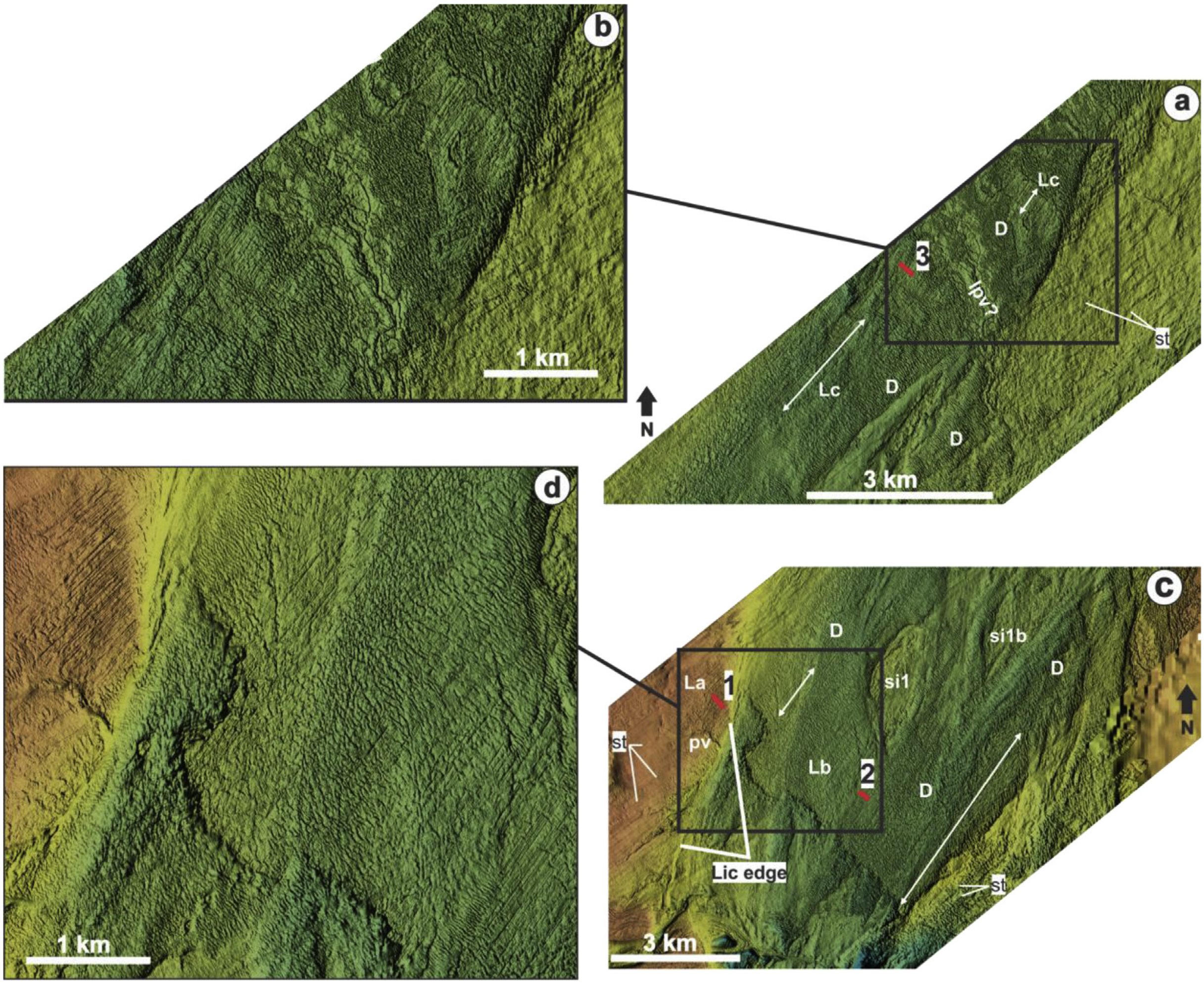

Fig. 13. Bathymetric data gridded at 1.5 m (vertical exaggeration: x15) showing linear groove-and-ridge bed-forms (i.e. La, Lb and Lc) apparently carved into chalk. Orientation of linear features is indicated by double-headed arrows. Note the apparent truncation of groove-and-ridge set "La" by the western edge of the Lic (LCWE). Note also that individual grooves and ridges are covered by, and have different orientation than, dunes (D) and minor infilled palaeovalleys (e.g. lpv). They present different orientations from linear artefacts (st: ship tracks) too. Red lines: topographic profiles plotted in Fig. 14. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

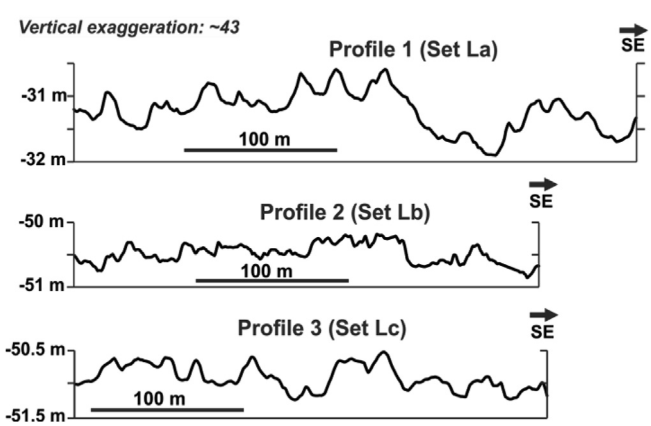

Fig. 14. Topographic profiles across the different sets of groove-and-ridge bed-forms. See Fig. 13 for location.

extends over, at least, 9 km along the western edge of the Lobourg Channel. However, its total extent is unknown, as a large tidal sand ridge cover its southern part. Ch2, on the other hand, is restricted to the eastern and southern part of the plateau formed between Ch1 and Ch2 by the outcropping Hythe Formation (Fig. 15). This channel is mostly carved into Atherfield Clay and Weald Clay formations (Fig. 5). It extends over ~4 km and shows ~20 m of relief at the seabed (Fig. 12). Palaeo-channel Ch2 branches to the northeast into two amphitheatre-shaped heads, the headwall of which reaches maximum height of ~24 m and has slopes of 10°–15° (see Gupta et al., 2017). Immediately northward of its headwall, the ~600-m-wide ovoid depression "d2" is eroded ~20 m into outcropping Hythe Formation (Fig. 15). Similarly, depression d1 is incised up to 25 m into Fosse A's infill. Depression "d1" is ellipsoidal in shape and is 560 m wide and 650 m long. Its main axis is oriented parallel to escarpment E1. In cross-section, it displays a scoop-shape along its main axis, showing a much steeper slope in the south (dip: 8°–9°) than in the north (dip: 2–3°).

Other significant erosional features carved in the central Dover Strait area, apart from the features described above, are longitudinal scours sc1. These scours are located within platform LC1 and present similar orientation to escarpment E1, suggesting that their formation is related to the erosional processes that also shaped platform LC1. Scours sc1 are mainly incised into Weald Clay Formation and some of them reach depths up to 10–20 m at the

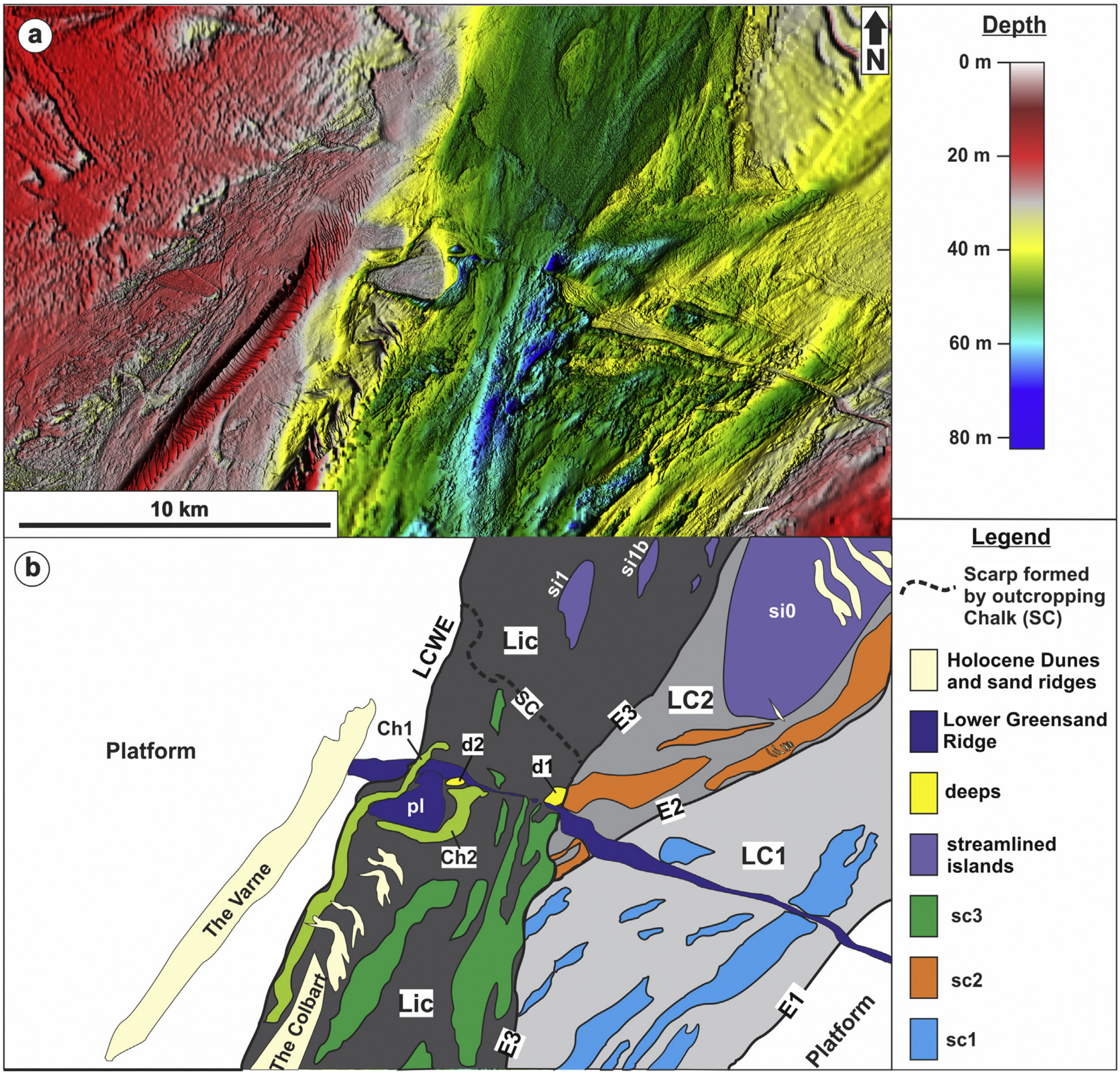

Fig. 15. Bathymetric–topographic map with a 40 m cell size (a) and geomorphological interpretation (b) of the central Dover Strait region (same area than in Fig. 5). SC: Scarp defined by base of Holywell Nodular Chalk Formation; Ch1 and Ch2: amphitheatre-head palaeo-channels (see also Gupta et al., 2017). For other labels, see legend and Fig. 11.

seafloor (Fig. 15). Importantly, scours sc1 are truncated by the Lobourg inner channel, attesting to the older age of the former.

Note that scours sc1 truncate the infill of Fosse G (Fig. 12), indicating that their formation, and so the oldest phase (LC1) of valley incision identified in the Lobourg Channel, postdates the formation of the Fosse Dangeard and the sediments infilling Fosse G. The lack of sedimentary data prevents correlations between the infill of Fosse G and those of other palaeo-depressions of the Fosse Dangeard. Hence, we have not been able to determine whether the valley incision/s that formed platforms LC1 and LC2 happened following the last infilling episode identified in Fosse A and B or, rather, they were contemporary to the incision of one or several of the internal erosional surfaces carved into the infills of these palaeo-depressions.

By contrast to escarpments E1 and E2, the Lobourg inner channel cuts through all bedrock formations outcropping in this area, including the resistant Hythe Formation (Fig. 5). The ridge formed by that stratigraphic unit across the width of the Strait is almost completely absent in the eastern half of the Lobourg inner channel (Fig. 15). Moreover, in the western half of the Lobourg inner channel, where the width of that ridge is at its maximum owing to local folding, palaeo-channel Ch1 and the ovoid depression d2 incise, respectively, ~5 m and ~20 m into that unit, attesting to the extreme erosion that carved these features. The Lobourg inner channel thus represents the most intense erosional event currently observable in the seafloor that took place following the formation of the Fosse Dangeard.

5.3. Southern Dover Strait

The Lobourg Channel in the southern Dover Strait cuts through formations of the Wealden, Purbeck, Portland and Ancholme groups, which are mainly composed of sandstones, siltstones, mudstones and shales (see Table 1). In this area, the Lobourg inner

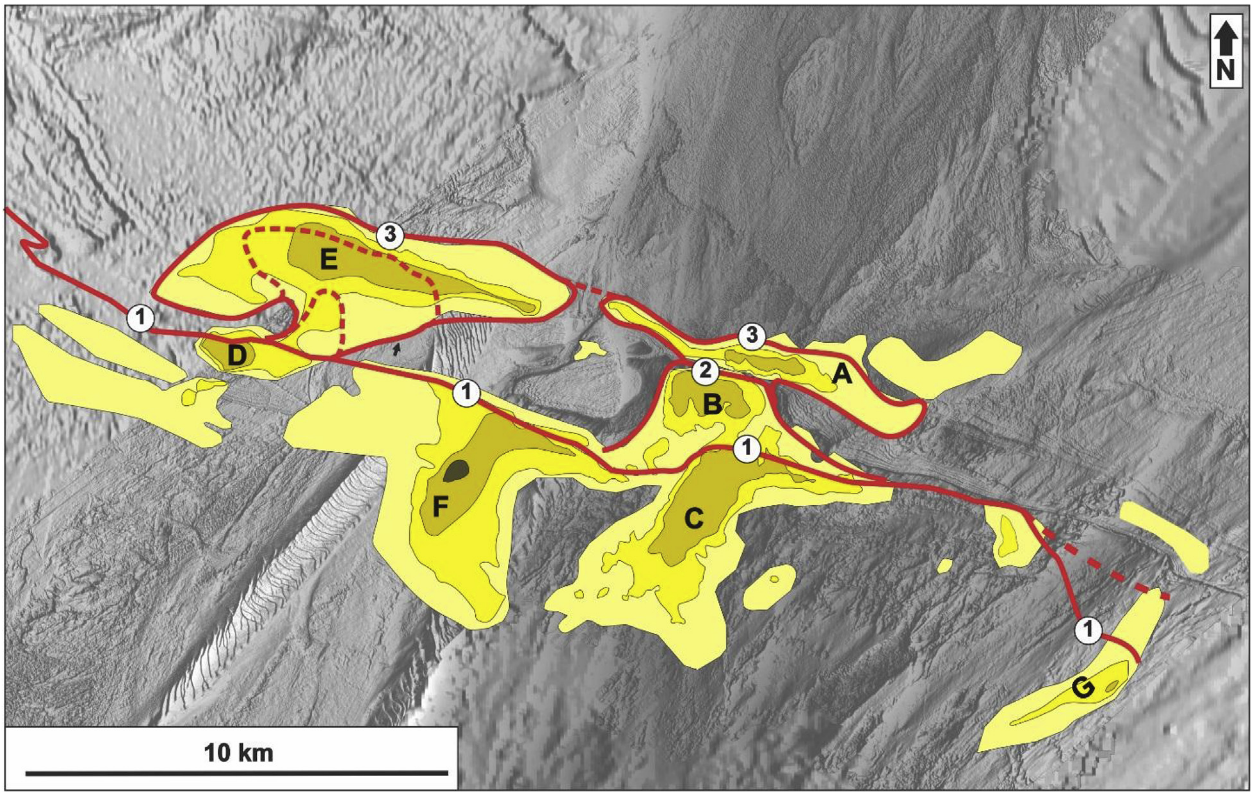

Fig. 16. Hypothetical knickpoint headward erosion during the waterfall phase that incised the Fosse Dangeard. Red line 1 indicates the position of the waterfall's wall during the incision of Fosse D and last phases of incision of Fosse F and C. Red line 2 shows the location of the waterfall during last phase of incision of Fosse B. Red line 3 indicates the last position of the waterfalls (i.e. last phase of incision of Fosse A and E) before the breach of the Weald-Artois ridge and opening of the Dover Strait. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

channel occupies the entire Lobourg Channel (Fig. 11). The Lobourg Channel thus passes from a 20–25 km wide terraced palaeovalley in the north and central Dover Strait to a 10-km-wide, channel in the south. Topographic cross sections across the southern Lobourg Channel (Fig. 12) show irregular box-shaped morphologies with steeper slopes in the west (3–4°) than in the east (1–2°). The margins of the Lobourg Channel in the southern Dover Strait are up to ~30 m tall. Erosional features within the Lobourg Channel in that area include linear and sinuous kilometre-scale scours up to 10 m deep, and kilometre-scale elongated streamlined islands up to 10–15 m high (Fig. 11).

6. Discussion

The buried and exposed landforms carved into the seafloor of the Dover Strait attest to a complex Quaternary evolution of this region, involving multiple erosional and depositional episodes. Here, we discuss the origin and inter-relationship of the various palaeo-depression and palaeovalley networks identified in this study.

6.1. Formation and infilling of the Fosse Dangeard

The remarkable localised depth of the Fosse Dangeard (up to 120–140 m) rules out marine erosional processes as a causal mechanism. We can also definitively exclude a tectonic origin, as tectonic activity appears to have been negligible in the Dover Strait since the formation of the Fosse Dangeard (García–Moreno et al., 2015). Nevertheless, faults are present beneath and cross-cut by the basal erosion surfaces of Fosse A, C, E and F (Fig. 5). The presence of these structures may have enhanced erosion along their planes by differential erosion. However, the fact that erosion of the palaeo-depressions also occurs in areas unaffected by faulting precludes structural control as the primary factor that caused the incision of the Fosse Dangeard.

The morphology of the Fosse Dangeard and their occurrence as relatively isolated sub-circular features indicates that they cannot be explained as sub-glacial tunnel valleys (Praeg, 2003; Lonergan et al., 2006; Kristensen et al., 2007). Moreover, because the Fosse Dangeard are carved into bedrock, they cannot be glacial kettle holes. Thus, a glacial erosion model as originally proposed by Destombes et al. (1975) is ruled out. The Fosse Dangeard are also tens of metres deeper than palaeo-depressions associated with fluvial erosion, such as those identified further southwest within the Channel palaeovalleys (Mellett et al., 2013). Their independence from the palaeovalleys currently exposed at the seafloor is further attested by the fact that several palaeo-depressions of the Fosse Dangeard are located outside the Lobourg Channel. The incisions of the Fosse Dangeard and Lobourg Channel were thus caused by different erosional processes that took place at different times.

The restriction of these palaeo-depressions to a belt extending along the projection of the chalk ridges forming cliffs at Calais and Dover today (Figs. 1 and 2), their morphology, and their significant depth/size support the model in which these palaeo-depressions were incised by large waterfalls. We thus interpret the Fosse Dangeard as palaeo-plunge pools initially carved by north-eastward retreating waterfalls. That is, cataracts retreating from Fosse C, D, F and G toward the northern edges of Fosse A and E, which most likely represent the last position of the waterfall's wall before the opening of the Dover Strait (Fig. 16). According to this interpretation, the waterfall phase must have ended at some point during the incision of Fosse A and E.

The interpretation of these features as the result of plunge-pool erosion along north-eastward retreating waterfalls is also supported, firstly, by the observation that all palaeo-depressions exhibit steeper slopes in the northeast than in the southwest, which indicates that the palaeo-flow that incised them came from the northeast (see Lamb et al., 2007; Pagliara et al., 2008). Secondly, the northeastern edges of Fosse C, D, F and G, similar to those of Fosse A and E, align with one another, suggesting that a linear barrier should have been located at some point along those edges. Thirdly, it is difficult to explain the location, depth and morphology of Fosse D without supposing that the cataract's wall was once located along Fosse D's northern edge. Indeed, significant erosive power would have been needed to form that palaeo-depression in such a resistant, structureless substratum (see Pagliara et al., 2008). Finally, plan-view and cross-sectional morphologies of all palaeo-depressions strongly resemble plunge pools carved at the base of natural and modelled waterfalls (e.g. Alexandrowicz, 1994; Lamb et al., 2007; Lamb, 2008; Pagliara et al., 2008; Strasser et al., 2008; Baker, 2009).

The combined interpretation of the seismic stratigraphy and seafloor morphology reveals a strong relationship between the different morphologies of individual palaeo-plunge pools and the type of rock where they were carved. For instance, the relatively high resistance to erosion of the Hythe Formation, together with the widening of its outcrop in the vicinity of Fosse F due to local folding and faulting, likely prevented the connection of Fosse F with Fosse A. By contrast, further to the east, the combination of folding and reverse faulting significantly narrowed the Lower Greensand Group immediately to the east of Fosse F. This local bedrock configuration resulted in a further northeast extent of the outcrop of the less resistant Weald Clay Formation in that area. This appears to have favoured a further northward progression of the incision, thus explaining the continuation of Fosse C into Fosse B, and the near connection of the latter with Fosse A.

The presence of tectonic structures and lithological variation within the incised bedrock may have also played an important role in defining the morphology of Fosse D. This palaeo-depression is carved along the axis of a minor anticline structure. The deformation produced by the minor anticline/syncline structure to which that fold belongs resulted in the presence of Hythe Formation to the northeast and southwest of Fosse D (Fig. 5b). That configuration would have favoured deeper incisions eastward, along the axis of the Anticline, where the Hythe Formation was most likely thinner and presented a less extensive outcrop. This explains the WNW–ESE elongation of Fosse D. Note also that Fosse D connects to Fosse E in the area where the local thickness of the outcropping Hythe Formation is at its minimum (Fig. 5b).

Having propagated past the more resistant Hythe Formation, the north-eastward retreating waterfall would have entered a faulted zone in the centre of the Strait, and encountered relatively less resistant stratigraphic units in the west, i.e., Sandgate, Folkestone and Gault formations (Fig. 5). This would have favoured deep incision again, resulting in the formation of Fosse A and E. In the case of Fosse A, the proximity of the sub-horizontally stratified, relatively resistant White Chalk subgroup, to the Hythe Formation along the fault plane would have limited the northward progression of the incision, thus constraining the width of Fosse A. Fosse E, on the other hand, extends northward across gently north-dipping Hythe, Sandgate, Folkestone and Gault formations. That is, the resistance to erosion of the bedrock across this palaeo-depression increases from south to north until it reaches the Zig Zag Chalk Formation, thus explaining the northward widening and deepening of this incision (Figs. 4 and 5).

Plunge-pool erosion is commonly attributed to vertical impact of falling water and sediments, plucking of fractured blocks and abrasion at the base of the waterfall (Whipple et al., 2000). According to analogue models, the depth and size of the incision is determined by the height of the waterfall, the volume of water falling, the sediment carried by the water and the composition of the substratum (Whipple et al., 2000; Lamb, 2008; Pagliara et al., 2008). Depths of plunge pools produced by failures of natural dams or due to steps in large rivers and/or flooded terrains usually range between a few metres to a few tens of metres (Alexandrowicz, 1994; Lamb et al., 2007; Lamb, 2008; Strasser et al., 2008; Baker, 2009). The depths of the Fosse Dangeard palaeo-plunge pools are in fact rather unusual.

The significant depths of the Fosse Dangeard may be due to the combination of several factors, i.e., the height of the waterfall, high flow discharge during lake overspill, the relatively high resistance to erosion of the waterfall crest, as well as the possible high content of sediments (e.g., flints) as tools for erosion in the water.

Even though the height of the waterfall drop is unknown, it is possible to place some bounds on its amplitude. Based on the heights of the chalk cliffs at either side of the Strait (>30 m above present sea level) and the depth of the platform into which the Lobourg Channel is carved (30–40 m below present sea level), it is likely that the waterfall was on a 60–70 m scale or higher. This is likely a minimum bound as the onshore cliff heights may not be representative of middle Pleistocene elevations of the chalk ridge across the strait.

The discharge of water cascading over the Weald–Artois ridge from a proglacial lake is impossible to constrain. However, it was most likely periodically enhanced by inputs from northwestern European palaeo-rivers and runoff from the ice complex during seasonal ice melting; specially at the end of the glacial maximum (Toucanne et al., 2009a; Gibbard, 1995; Murton and Murton, 2012; Gibbard and Cohen, 2015).

Concerning the outcropping rocks at the time of the incision, we propose that the crest of the waterfall consisted of formations of the White Chalk Subgroup. This is based on the average thickness of the Zig Zag Chalk Formation measured in our seismic dataset, which ranges between 40 and 50 m (see also Hamblin et al., 1992). Hence, an extrapolation of the base of the sub-horizontally stratified Holywell Nodular Chalk Formation to the edge of Fosse A and B would put that formation 40–50 m above the present-day seafloor at that location. The base of the waterfall, on the other hand, comprised most likely Quaternary colluvium on top of outcropping Lower Greensand Group in the west and Wealden Group in the centre and east. The relatively low erodibility of some of the formations composing the White Chalk Subgroup may have resulted in slow headward erosion of the waterfall wall, inducing longer exposure of its base to water and sediment impact. This, combined with the locally less resistant and heterogeneous bedrock outcropping at the base of the waterfall, may have favoured deep incision in this area.

Finally, the erosional power of the waterfall may have also been increased at its base by the presence of sediments forming tools for erosion. These probably came from remobilization of previously deposited colluvium, and from some gravelly and sandy subunits of the Wealden Group and Lower Greensand Group. In addition, the water may have contained numerous flints dislodged from the White Chalk Subgroup by dissolution, rock fragments resulting from seasonal frost weathering of the ridge and ice fragments removed from the glaciers.

Following the waterfall phase that formed the Fosse Dangeard, these palaeo-depressions were subjected to a number of erosional/depositional episodes that gave place to their present-day morphology and infill. This is evidenced by several internal erosional surfaces incised into their infills. We have identified the occurrence of 5–6 major erosional-and-infilling episodes in Fosse A and B, 2 in Fosse E, and a minimum of one in the remainder of the palaeo-depressions. Two of the internal erosional surfaces identified in the infill of Fosse B (i.e. Eb1 and Eb4) removed locally all previous sediments, which suggest relatively intense erosion. That seems to be corroborated by their cross-sectional morphologies, which strongly resemble scours incised by high-magnitude flood erosion and/or by high-discharge fluvial erosion (see Eilertsen and Hansen, 2008). These internal erosional surfaces thus suggest the occurrence of episodes of intense fluvial and/or flood erosion in the timespan between the formation of the Fosse Dangeard and the incision of the Lobourg inner channel. It is important to note however that these later events were not on the scale of the original erosional event that excavated the Fosse Dangeard itself.

The geometry of the other erosional surfaces identified in the various palaeo-depressions and the seismic units above them show no distinct characteristic allowing the identification of the processes that formed them. Only the uppermost (i.e., youngest) erosional surfaces (Ee2, Ed1, Ef1, Eb6 and Ea5) exhibit some similarities with erosional surfaces and seismic facies associated with marine transgressions, such as apparent widespread spatial distribution of the erosion, sub-planar geometry of the erosional surface and diffuse reflection above (see Trincardi et al., 1994). It is thus possible that the formation of those erosional surfaces is related to a marine transgression that took place before the incision of the Lobourg inner channel.

The lack of sediment cores from Fosse A, B, C, D and G precludes establishing unambiguously the palaeo-environments and chronology of the infills of the Fosse Dangeard. Currently, we can only state that the scouring-and-infilling episodes that resulted in the present-day infills of the Fosse Dangeard took place following the waterfall phase that led to the breach of the Dover Strait dam, and that they occurred before the incision of the Lobourg inner channel (see below).

6.2. Formation of the Lobourg Channel

The geomorphological analysis of the Lobourg Channel shows that it is more complex than previously thought. For instance, it comprises two terrace-like platforms (LC1 and LC2) in the north, which are truncated by a major inner channel (the Lobourg inner channel) in the centre of the Strait.

The overall morphology of the Lobourg Channel suggests that platforms LC1 and LC2 are remnants of one or several episodes of channel incision that took place before the formation of the Lobourg inner channel. The morphology of the palaeovalley(s) within which platforms LC1 and LC2 were excavated are unknown, since no trace of their western banks remains. Nonetheless, the morphology of the Lobourg Channel suggests that they may have been rather wide. Indeed, if their western banks coincided with the present-day western edge of the Lobourg inner channel, their width could have been greater than 20 km in the northern Dover Strait (Figs. 11 and 12). The fact that these platforms are parallel to each other also prevents assessing whether they were formed during the same or different erosional episodes. That is, whether they are remainders of one or several phases of incision along the Lobourg Channel. In any case, the orientation of escarpments E1 and E2 suggests that the palaeovalley(s) demarcated by platforms LC1 and LC2 narrowed south-westward as it approached the central Dover Strait (Fig. 11).

The morphology of platforms LC1 and LC2, the Lobourg inner channel and the erosional features associated with their incisions (i.e., streamlined islands, linear scours, amphitheatre-head channels) present strong similarities to features found in valleys eroded by high-discharge rivers and/or high-magnitude flood flows (Kehew and Lord, 1986; Wohl, 1993; Rains et al., 1993; Kale et al., 1996; Baynes et al., 2015). Based on these similarities, we interpret the Lobourg Channel as a palaeovalley mainly incised by several phases of intense flood and fluvial erosion (see also Collier et al., 2015; Gupta et al., 2017).

Features indicating intense fluvial/flood erosion are especially associated with the last phase of valley incision along the Lobourg Channel, i.e., the one that eroded the Lobourg inner channel. This is attested by the observation that, by contrast to platforms LC1 and LC2, scours carved within the Lobourg inner channel locally truncate the ridge formed at the seafloor by the relatively resistant Hythe Formation. In addition, this valley cross-cuts a range of bedrock sedimentary formations whilst maintaining a box-shape cross-sectional profile independently of the bedrock through which it cuts. This morphology is indeed similar to that found along high-energy fluvial systems and megaflood-eroded valleys (see Kehew and Lord, 1986; Wohl, 1993; Rains et al., 1993; Kale et al., 1996; Baynes et al., 2015). Finally, the Lobourg inner channel comprises a range of bedrock erosional features, such as isolated tens-of-metres deep depressions, amphitheatre-head channels, streamlined islands, longitudinal scours, linear ridge-and-groove bedforms, etc., which are typically found in megaflood-eroded terrains (Baker and Nummedal, 1978; Rains et al., 1993; Wohl, 1993; Lamb et al., 2007; Baker, 2009; Shaw, 2010; Baynes et al., 2015).

Taken together, the low topographic gradient across the Strait (see Lericolais et al., 2003), the morphology of the Lobourg inner channel and the erosional features carved within it suggest that at least this palaeovalley was most likely carved by one or several high discharge flood flows. Whether the formation of the previous valley incisions that produced platforms LC1 and LC2 was also caused by similar flows is unknown. Nevertheless, the presence of streamlined islands carved into chalk within platform LC2 and the deep elongated scours incised into LC1 and LC2 suggests intense fluvial and/or flood erosion also.

The occurrence of several episodes of high-magnitude flood flows in the Dover Strait might also be supported by the linear ridge-and-groove bedforms identified inside and outside the Lobourg inner channel. These features strongly resemble linear-groove bedforms excavated by megafloods in Washington State, USA, and Piccaninny Creek, Australia (see Baker and Nummedal, 1978; Wohl, 1993; Baker, 2009). Nevertheless, more data are needed from these structures to unambiguously ascertain how and when they formed.

Another important result from our geomorphological analysis is the identification of a SE–NW-oriented palaeovalley system (Pvs1) that appears to represent either a tributary to the Lobourg inner channel or a former valley truncated by it (Fig. 11). The fact that this system remains undisturbed across LC1 and LC2 implies that major rivers and/or floods flowing across the Dover Strait were channelled through the Lobourg inner channel following the incision of Pvs1. That means that the Lobourg inner channel represents the last major episode of flood and fluvial erosion that imprinted the seafloor of the Dover Strait.

6.3. Relative sequence of events in the landscape evolution of the Dover Strait

Our study provides evidence for the occurrence of several major subaerial erosional episodes in the Dover Strait, which took place prior to the Holocene Epoch. Their absolute ages are not constrained. Our investigation, however, provides sufficient information to assess the relative sequence of events. In chronological order, we identify the following episodes:

I. Excavation of the Fosse Dangeard by plunge-pool erosion at the base of waterfalls overspilling the Weald-Artois rock ridge;
II. Breaching of the Weald-Artois ridge causing initial opening of the Dover Strait;
III. Infilling of the Fosse Dangeard, which included 5 to 6 major scouring-and-infilling episodes in the centre of the Strait. At least two of these episodes are suggestive of intense fluvial scouring or flood erosion;
IV. Formation of platforms LC1 and LC2 along the Lobourg Channel possibly by highly erosional fluvial system(s) and/or high-magnitude flood flows. The relative age of these features is unclear. It is indeed possible that their formation was contemporaneous with some of the internal erosional surfaces identified in Fosse A and B;
V. Incision of the Lobourg inner channel (Lic) by high discharge flood processes.

The initial overflow of a North Sea lake and opening of the Strait is widely considered to have occurred during the Elsterian glaciation, i.e., approximately 450 ka ago (e.g. Gibbard and Cohen, 2015). This is potentially the time during which the waterfall phase and incision of the Fosse Dangeard occurred, as well as an initial breach of the Weald-Artois ridge.

Once opened, the Dover Strait became the principal outflow of the southern North Sea ice-marginal lake during the rest of the Elsterian glacial maximum. The Dover Strait area was also the main drainage route of fluvial systems traversing the southern North Sea area during the Saalian (175–150 ka BP) and Weichselian (30–19 ka BP) glacial maxima (Gibbard, 1995; Hijma et al., 2012; Toucanne et al., 2015; Sejrup et al., 2016). The palaeo-rivers traversing the Dover Strait during those time intervals were some of the largest drainage systems in Europe (e.g. Gibbard and Cohen, 2015; Patton et al., 2017). These river systems were most likely subject to recurrent increases of water volume due to seasonal ice-melting and/or lake-outburst floods produced at lakes formed within its catchment (Toucanne et al., 2009b, 2010, 2015; Meinsen et al., 2011; Patton et al., 2017; Lang et al., 2018). In addition, these river systems ran across the Dover Strait for thousands-of-years during each Middle–Late Pleistocene glaciation, possibly inducing significant gradual fluvial erosion along their valleys (Westaway and Bridgland, 2010; Mellett et al., 2013).

Therefore, the scouring-and-infilling episodes that resulted in the infilling of the Fosse Dangeard and the various flood and fluvial erosional episodes/events that carved the Lobourg Channel appears to have occurred in the time interval between the Elsterian glacial maximum (450 ka) and the Holocene marine transgression (12–8 ka). With the data at hand, it is however impossible to associate the various internal erosional surfaces identified in the infill of the Fosse Dangeard or the different erosional phases that formed the Lobourg Channel with specific glacial and interglacial stages.

7. Conclusions

Analysis of high-resolution geophysical data provides new information on the formation and evolution of the Dover Strait with particular focus on the mechanism that incised the Fosse Dangeard and Lobourg Channel. Our analysis enables us to draw the following conclusions:

(1) Detailed 3-dimensional characterization of the Fosse Dangeard provides distinct morphological evidence supporting their interpretation as palaeo-plunge pools incised at the base of waterfalls. Building on the study of Gupta et al. (2017), our analysis corroborates the existence of a rock ridge extending across the Dover Strait, and the formation of waterfalls that spilled over the top of it from a lake that formed in the southern North Sea region at least once during the Pleistocene epoch.
(2) Stratigraphic/morphologic relationships between the Fosse Dangeard and the bedrock into which these palaeo-depressions are carved indicate that the south-facing escarpment formed by the Weald–Artois ridge retreated northwards during overflow, with the northeastern edges of Fosse A and E marking its position just before the final breach phase.
(3) The various seismic units composing the infill of the Fosse Dangeard indicate that these palaeo-depressions were subject to a series of scour-and-fill episodes following their initial excavation. These episodes occurred before the last phase of valley incision along the Lobourg Channel.
(4) Mapping of the morphology of the Lobourg Channel from high-resolution bathymetric data shows evidence for multiple phases of valley incision, which are likely associated with fluvial erosion and episodes of high-magnitude flood erosion. Notably, the last phase of valley incision strongly suggests the occurrence of at least 1 episode of catastrophic flooding/megaflood.
(5) The present study demonstrates that initiation of the opening of the Dover Strait and the various palaeovalleys and palaeo-depressions carved in its seafloor are the result of fluvial erosion and episodes of high-magnitude flooding (e.g. megafloods), which most likely took place during Middle–Late Pleistocene glacial stages. By comparison with the present-day geographic/geologic setting, it is reasonable though to assume that erosion during marine highstands were also instrumental in widening the Strait by coastal erosion and by tidal flux through the seaway. However, this appears to have produced a minor overprint on the seafloor given the distinct evidence for erosion by subaerial fluvial/flood events.

Author contributions

S.G., J.S.C., M.D.B., T.C and D.G-M. conceived, designed and coordinated the study. D.G-M., J.S.C. and F.O. processed bathymetric data. D.G-M, K.VAN. and W.V. designed, coordinated and conducted RV Belgica geophysical surveys with important contributions from K.VER., H.J. and A.T. K.VAN. processed RCMG–ROB and Dangeard I seismic-reflection data with contributions from D.G-M. A.T. provided Dangeard I seismic-reflection data. D.G-M. analysed and interpreted the data with contributions from S.G. and J.S.C. All authors discussed the results. D.G-M wrote the paper with contributions from S.G. and J.S.C. All authors reviewed the manuscript.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.quascirev.2018.11.011.

References

Antoine, P., Coutard, J.-P., Gibbard, P., Hallegouet, B., Lautridou, J.-P., Ozouf, J.-C., 2003. The Pleistocene rivers of the English Channel region. J. Quat. Sci. 18, 227–243.
Alexandrowicz, Z., 1994. Geologically controlled waterfall types in the Outer Carpathians. Geomorphology 9, 155–165.
Aldiss, D.T., Farrant, A.R., Hopson, P.M., 2012. Geological mapping of the Late Cretaceous Chalk Group of southern England: a specialised application of landform interpretation. Proc. Geologists' Assoc. 123, 728–741.
Arthur, J.C.R., Phillips, G., McCormick, C.R., 1997. High definition seismic for Channel Tunnel marine route. In: McCann, D.M., Eddleston, M., Fenning, P.J., Reeves, G.M. (Eds.), Modern Geophysics in Engineering Geology. Geological Society Engineering Geology, Special Publication, 12, pp. 327–334.
Ashton, N., Lewis, S., 2002. Deserted Britain: declining populations in the british late middle Pleistocene. Antiquity 76, 388–396.
Ashton, N., Hosfield, R., 2010. Mapping the human record in the British early Palaeolithic: evidence from the Solent River system. J. Quat. Sci. 25, 737–753.
Ashton, N., Lewis, S.G., Hosfield, R., 2011. Mapping the human record: population change in Britain during the Early Palaeolithic, 39–51. In: Ashton, N., Lewis, J.E., Stringer, C. (Eds.), The Ancient Human Occupation of Britain.
Baker, V.R., Nummedal, D., 1978. The Channeled Scabland; a Guide to the Geomorphology of the Columbia Basin. NASA, Washington.
Baker, V.R., 2009. The channeled scabland: a retrospective. Annu. Rev. Earth Planet Sci. 37, 393–411.
Balson, P.S., D'Olier, B.D., 1989. Thames estuary solid geology. Sheet 51°N-00°. Br. Geol. Surv. scale 1:250 000, 1 sheet.
Baynes, E.R.C., Attal, M., Dugmore, A.J., Kirstein, L.A., Whaler, K.A., 2015. Catastrophic impact of extreme flood events on the morphology and evolution of the lower Jokulsa a Fjollum (northeast Iceland) during the Holocene. Geomorphology 250, 422–436.
Busschers, F.S., Van Balen, R.T., Cohen, K.M., Kasse, C., Weerts, H.J.T., Wallinga, Bunnik, F.P.M., 2008. Response of the Rhine-Meuse fluvial system to Saalian ice-sheet dynamics. Boreas 37, 377–398.
Carr, S.J., Holmes, R., van der Meer, J.J.M., Rose, J., 2006. The last glacial maximum in the North Sea basin; micromorphological evidence of extensive glaciation. J. Quat. Sci. 21, 131–153.
Cohen, K.M., Gibbard, P.L., Weerts, H.J.T., 2014. North Sea palaeogeographical reconstructions for the last 1 Ma. Neth. J. Geosci. 93, 1–23.
Collier, J.S., Oggioni, F., Gupta, S., García-Moreno, D., Trentesaux, A., De Batist, M., 2015. Streamlined islands and the English Channel megaflood hypothesis. Global Planet. Change 135, 190–206.
Crosby, A., Hamblin, R.J.O., Smart, J.G.O., Bouysse, P., 1988. Dungeness-Boulogne solid geology, Sheet 50°N-00°. British Geological Survey scale 1:250 000, 1 sheet.
Destombes, J.P., Shephardthorn, E.R., Redding, J.H., Morzadec-Kerfourn, M.T., 1975. Buried valley system in the strait of dover. Philos. Trans. R. Soc. London, Ser. A 279, 243–256.
Dyer, K.R., Huntley, D.A., 1999. The origin, classification and modelling of sand banks and ridges. Continent. Shelf Res. 19, 1285–1330.
Eilertsen, R.S., Hansen, L., 2008. Morphology of river bed scours on a delta plain revealed by interferometric sonar. Geomorphology 94, 58–68.
Ehlers, J., Gibbard, P.L. (Eds.), 2004. Quaternary Glaciations – Extent and Chronology, Part I: Europe. Developments in Quaternary Science, vol. 2a. Elsevier, Amsterdam.
García-Moreno, D., Verbeeck, K., Camelbeeck, T., De Batist, M., Oggioni, F., Zurita Hurtado, O., Versteeg, W., Jomard, H., Collier, J., Gupta, S., Trentesaux, A., Vanneste, K., 2015. Fault activity in the Epicentral area of the 1580 dover-Strait/ Pas- de-Calais earthquake (North-Western Europe). Geophys. J. Int. 201, 528–542.
Gibbard, P.L., 1988. The history of the great northwest European nvers during the past three million years. Phil. Trans. Roy. Soc. Lond. B 318, 559–602.
Gibbard, P.L., 1995. the formation of the strait of dover. J. Geol. Soc. (London, U.K.), Special Publications 96, 15–26.
Gibbard, P., Cohen, K.M., 2008. Global chronostratigraphical correlation table for the last 2.7 million years. Episodes 31 (2), 243.
Gibbard, P.L., Cohen, K.M., 2015. Quaternary evolution of the north sea and the English channel. Proc. of the Open University Geol. Soc. 1, 63–74.
Gibbard, P.L., Lewin, J., 2016. Filling the north Sea basin: cenozoic sediment sources and river styles. Geol. Belg. 19, 3–4.
Graham, A.G.C., Lonergan, L., Stoker, M.S., 2007. Evidence for Late Pleistocene ice stream activity in the Witch Ground basin, central North Sea, from 3D seismic reflection data. Quat. Sci. Rev. 26, 627–643.
Gupta, S., Collier, J.S., Palmer-Felgate, A., Potter, G., 2007. Catastrophic flooding origin of shelf valley systems in the English Channel. Nature 448, 342–345.
Gupta, S., Collier, J.S., García-Moreno, D., Oggioni, F., Trentesaux, A., Vanneste, K., De Batist, M., Camelbeeck, T., Potter, G., Van Vliet Lanoe, B., Arthur, J.C.R., 2017. Making Britain: two-stage catastrophic opening of the Dover Strait. Nat. Commun. 8 https://doi.org/10.1038/ncomms15101.
Hamblin, R.J.O., Crosby, A., Balson, P.S., Jones, S.M., Chadwick, R.A., Penn, I.E., Arthur, M.J., 1992. United Kingdom Offshore Regional Report: the Geology of the English Channel. HMSO for the British Geological Survey, London, p. 106.
Hancock, G.S., Anderson, R.S., 2002. Numerical modeling of fluvial strath-terrace formation in response to oscillating climate. Geol. Soc. Am. Bull. 114, 1131–1142.
Hijma, M.P., Cohen, K.M., Roebroeks, W., Westerhoff, W.E., Busschers, F.S., 2012. Pleistocene Rhine–Thames landscapes: geological background for hominin occupation of the southern North Sea region. J. Quat. Sci. 27, 17–39.
Hopson, P.M., 2005. A stratigraphical framework for the upper cretaceous chalk of england and scotland, with statements on the chalk of northern Ireland and the UK offshore sector. In: British Geological Survey Research Report RR/05/01, p. 120.
Hopson, P.M., Wilkinson, I.P., Woods, M.A., 2008. A stratigraphical framework for the lower cretaceous of england. In: British Geological Survey Research Report RR/08/03, p. 87.
James, J.W.C., Guennoc, P., Harrison, M., Le Bot, S., Philpott, S., Vinchon, C., Bee, E., Simien, F., Janjou, D., Garlan, T., Trentesaux, A., Mahieux, G., Briet, D., Augris, C., 2002. GEOSYNTH: a synthesis of the geology and sediments of the Dover Strait and its Hinterland. CR/02/078. CDROM, British Geological Survey.
Kale, V.S., Baker, V.R., Mishra, S., 1996. Multi-channel patterns of bedrock rivers: an example from the central Narmada basin, India. Catena 26, 85–98.
Kehew, A.E., Lord, M.L., 1986. Origin and large-scale erosional features of Glacial-Lake spillways in the northern Great-Plains. Geol. Soc. Am. Bull. 97, 162–177.
Kristensen, T.B., Huuse, M., Piotrowski, J.A., Clausen, O.R., 2007. A morphometric analysis of tunnel valleys in the eastern North Sea based on 3D seismic data. J. Quat. Sci. 22, 801–815.
Lamb, M.P., Howard, A.D., Dietrich, W.E., Perron, J.T., 2007. Formation of amphitheater-headed valleys by waterfall erosion after large-scale slumping on Hawaii. Geol. Soc. Am. Bull. 119, 805–822.
Lamb, M.P., 2008. Formation of Amphitheater-headed Canyons. PhD Dissertation. University of California, Berkeley, p. 311.
Lang, J., Lauer, T., Winsemann, J., 2018. New age constraints for the Saalian glaciation in northern central Europe: implications for the extent of ice sheets and related proglacial lake systems. Quat. Sci. Rev. 180, 240–259.
Lee, J.R., Busschers, F.S., Sejrup, H.P., 2012. Pre-Weichselian Quaternary glaciations of the British Isles, The Netherlands, Norway and adjacent marine areas south of 68 N: implications for long-term ice sheet development in northern Europe. Quat. Sci. Rev. 44, 213–228.
Lericolais, G., Auffret, J.-P., Bourillet, J.-F., 2003. The Quaternary Channel River: seismic stratigraphy of its palaeovalleys and deeps. J. Quat. Sci. 18, 245–260.
Lonergan, L., Maidment, S.C., Collier, J.S., 2006. Pleistocene subglacial tunnel valleys in the central North Sea basin: 3-D morphology and evolution. J. Quat. Sci. 21, 891–903.
Meinsen, J., Winsemann, J., Weitkamp, A., Landmeyer, N., Lenz, A., Dölling, M., 2011. Middle Pleistocene (Saalian) lake outburst floods in the Münsterland Embayment (NW Germany): impacts and magnitudes. Quat. Sci. Rev. 30, 2597–2625.
Meijer, T., Preece, R.C., 1995. Malacological evidence relating to the insularity of the british isles during the quaternary. Geol. Soc. Spec. Publ. 96, 89–110.
Mellett, C., Hodgson, M., Plater, A., Mauz, B., Selby, I., Lang, A., 2013. Denudation of the continental shelf between Britain and France at the glacial–interglacial timescale. Geomorphology 203, 79–96.
Mortimore, R., James, L., 2015. The search for onshore analogues for the offshore upper cretaceous chalk of the north sea. Proc. Geologists' Assoc. 126, 188–210.
Murton, D.K., Murton, J.B., 2012. Middle and late Pleistocene glacial lakes of lowland Britain and the southern North Sea basin. Quat. Int. 260, 115–142.
Pagliara, S., Hager, W.H., Unger, J., 2008. Temporal evolution of plunge pool scour. J. Hydraul. Eng. 134, 1630–1638.
Patton, H., Hubbard, A., Andreassen, K., Auriac, A., Whitehouse, P.L., Stroeven, A.P., Shackleton, C., Winsborrow, M., Heyman, J., Hall, A.M., 2017. Deglaciation of the Eurasian ice sheet complex. Quat. Sci. Rev. 169, 148–172.
Praeg, D., 2003. Seismic imaging of mid-Pleistocene tunnel-valleys in the North Sea Basin – high resolution from low frequencies. J. Appl. Geophys. 53, 273–298.
Preece, R.C., 1995. Island Britain: a quaternary perspective. J. Geol. Soc. (London, U.K.), Special Publication 96, 1–2.
Radley, J.D., Allen, P., 2012. The wealden (non-marine lower cretaceous) of the Weald sub-basin, southern england. Proc. Geologists' Assoc. 123, 245–318.
Rains, B., Shaw, J., Skoye, R., Sjogren, D., Kvill, D., 1993. Late Wisconsin subglacial megaflood paths in Alberta. Geology 21, 323–326.
Reynaud, J.-Y., Tessier, B., Auffret, J.-P., Berné, S., De Batist, M., Marsset, T., Walker, P., 2003. The offshore quaternary sediment bodies of the English channel and its western approaches. J. Quat. Sci. 18, 361–371.
Sejrup, H.P., Clark, C.D., Hjelstuen, B.O., 2016. Rapid ice sheet retreat triggered by ice stream debuttressing: evidence from the North Sea. Geology 44, 355–358.
Shaw, J., 2010. In defence of the meltwater (megaflood) hypothesis for the formation of subglacial bedform fields. J. Quat. Sci. 25, 249–260.
Smith, A.J., 1985. A catastrophic origin for the palaeovalley system of the eastern

English Channel. Mar. Geol. 64, 65–75.
Strasser, M., Schindler, C., Anselmetti, F.S., 2008. Late Pleistocene earthquake-triggered moraine dam failure and outburst of Lake Zurich, Switzerland. J. Geophys. Res.: Earth Surface 113. https://doi.org/10.1029/2007JF000802.
Stuart, A.J., 1995. Insularity and quaternary vertebrate faunas in Britain and Ireland. J. Geol. Soc. (London, U.K.), Special Publications 96, 111–125.
Sutcliffe, A.J., 1995. Insularity of the British Isles 250 000–30 000 years ago: the mammalian, including human, evidence. J. Geol. Soc. (London, U.K.), Special Publications. 96, 127–140.
Toucanne, S., Zaragosi, S., Bourillet, J.F., Gibbard, P.L., Eynaud, F., Giraudeau, J., Turon, J.L., Cremer, M., Cortijo, E., Martinez, P., Rossignol, L., 2009a. A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin. Quat. Sci. Rev. 28, 2974–2981.
Toucanne, S., Zaragosi, S., Bourillet, J.F., Cremer, M., Eynaud, F., Van Vliet-Lanoë, B., Penaud, A., Fontanier, C., Turon, J.L., Cortijo, E., Gibbard, P.L., 2009b. Timing of massive 'Fleuve Manche' discharges over the last 350 kyr: insights into the European ice-sheet oscillations and the European drainage network from MIS 10 to 2. Quat. Sci. Rev. 28, 1238–1256.
Toucanne, S., Zaragosi, S., Bourillet, J.F., Marieu, V., Cremer, M., Kageyama, M., Van Vliet-Lanoë, B., Eynaud, F., Turon, J.-L., Gibbard, P.L., 2010. The first estimation of Fleuve Manche palaeoriver discharge during the last deglaciation: evidence for Fennoscandian ice sheet meltwater flow in the English Channel ca 20–18ka ago. Earth Planet Sci. Lett. 290, 459–473.
Toucanne, S., Soulet, G., Freslon, N., Jacinto, R.S., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J.-F., Bayon, G., 2015. Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate. Quat. Sci. Rev. 123, 113–133.
Trincardi, F., Correggiari, A., Roveri, M., 1994. Late Quaternary transgressive erosion and deposition in a modern epicontinental shelf: the Adriatic semi-enclosed basin. Geo Mar. Lett. 14, 41–51.
Turner, C., 2000. The Eemian interglacial in the North European plain and adjacent areas. Neth. J. Geosci. 79, 217–231.
Westaway, R., Bridgland, D.R., 2010. Causes, consequences and chronology of large-magnitude palaeoflows in Middle and Late Pleistocene river systems of northwest Europe. Earth Surf. Process. Landforms 35, 1071–1094.
Whipple, K.X., Hancock, G.S., Anderson, R.S., 2000. River incision into bedrock: mechanics and relative efficacy of plucking, abrasion, and cavitation. Geol. Soc. Am. Bull. 112, 490–503.
Wohl, E.E., 1993. Bedrock channel incision along Piccaninny Creek, Australia. J. Geol. 101, 749–761.